Subsurface and outcrop characteristics of fluvial-dominated deep lacustrine clinoforms

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ABSTRACT

Deep lacustrine deposits provide records of palaeoclimatic and tectonics, and often host major hydrocarbon reservoirs, but their facies description and long-term stratigraphic architecture are not sufficiently reported. A combined outcrop and three-dimensional seismic dataset in the western Dacian Basin of Romania is used to decipher depositional systems, basin fill architecture and clastic sediment dispersal across a narrow shelf. The lake clinoforms are about 400 m high in seismic images and all of their aggradational bottomsets total up to 350 m in thickness. Depositional elements and morphology of fluvial channels, delta lobe complexes, sublacustrine channel forms, sublacustrine canyons and deep-lacustrine lobes are interpreted on the seismic attribute maps. Outcrops show that sharp-based deltaic units contain thin delta front deposits. The slope succession is dominated by channel-levée thin-bedded turbidites with terrestrial debris. Thicker and coarser turbidites are founded...
in sublacustrine channels. The channelized sandstones on the slope are 10 to 25 m thick and often overlie tens of metre thick mass-transport deposits. Tabular turbidite beds, sandy-conglomeratic debrites with shallow-water fossils, mud-rich mass-transport deposits and hybrid event beds within fan lobes are found on the basin floor. The integrated seismic-outcrop analysis suggests that low accommodation on the narrow (10 to 30 km) morphological shelf and high sediment supply resulted in the prograding lacustrine shelf-margin clinoforms with fluvial-dominated topsets and significant sediment bypass to the deep-lacustrine. The late Miocene–Pliocene Dacian Basin provides a typical example of a supply (river) dominated basin margin and possible recognition criteria of deep-lacustrine clinothems including: fluvial dominated topset deposits with abrupt vertical facies changes, bottomset-dominated sediment partitioning, and frequent sediment gravity flow activities denoted by closely-spaced and aggradational channel-levée systems, thick bottomsets and rare indication of sediment starvation in the deep-lacustrine deposits.

INTRODUCTION

Studies of linked shallow to deep depositional systems of deep lacustrine/lake basins are relatively few when compared to those of their marine counterparts. Concepts of deep marine depositional systems and sequence stratigraphy have been frequently extrapolated to deep lake analyses (e.g. Scholz & Rosendahl, 1990; Changsong et al., 2001) even though lake systems differ significantly from marine systems due to their isolation of open-sea influences and smaller water mass (Caroll & Bohacs, 1999). The amplitude and frequency of lake-level fluctuation are sensitive to climate change induced by Milankovitch orbital cycles and the magnitude of change can be larger and/or more frequent than marine systems (e.g. Olsen, 1990; Scholz et al., 1990; Baltzer, 1991; Sáez & Cabrera, 2002; Lyons et al., 2011; Feng et al., 2013). In modern sublacustrine systems such as Lake Baikal (Kuzmin et al., 2000), Lake Mead (Twichell et al., 2005) and Lake Texoma (Olariu et al., 2012) hyperpycnal flows are as significant as failure-generated turbidity currents and other mass-wasting processes. As a result, thick and coarse-grained (sand-gravel) river-derived turbidites are common features on lake floors for examples, in Lake Brienz (Sturm & Matter, 1978), Lake Malawi (Soreghan et al., 1999) and the Dongying depression, East China (Feng et al., 2013).
Most of the existing deep-lake examples occurred in extensional settings such as the Pleistocene–Holocene Lakes Tanganyika and Malawi (Scholz et al., 1990; Soreghan et al., 1999; Lyons et al., 2011) and the over 1500 m deep Lake Baikal (Nelson et al., 1999). The Hungarian late-Miocene Lake Pannon, of back-arc origin, is an example of a well-studied deep lacustrine basin that was filled by multiple sets of 500 to 700 m high clinoforms (e.g. Juhász et al., 2007; Uhrin & Sztanó, 2012; Magyar et al., 2013; Sztanó et al., 2013). Hyperpycnal flows from three large rivers deposited significant volumes of turbidites in the deep-lacustrine Lake Pannon (Juhacz et al., 2007; Sztanó et al., 2013; Csato et al., 2015). While subsurface imaging, core and wireline log are powerful tools for investigating deep lake systems (e.g. Lyons et al., 2011) outcrop data are still needed for understanding sub-seismic sedimentary architectures of such systems.

This study focuses on the western Dacian Basin of Romania (Fig. 1) in which a lake system with 400 to 500 m palaeowater depth and sufficient sediment supply (Fongngern et al., 2016) allowed shelf-margin scale clinoforms (Steel & Olsen, 2002) to form. The case study provides an example of a deep lake in a foreland tectonic setting (Jipa, 1997). Continuous outcrops of the deep-lake infill of the western Dacian Basin, including fluvio-deltaic, slope and basin-floor environments are described. Here, it is emphasized that the western Dacian Basin was hundreds of metres deep at least during Meotian–Pontian times and the conventional stratigraphy of sand-rich regressive and mud-rich transgressive units actually represents motifs of sand-rich basin-floor deposits, mainly mud-rich slope deposits and sand-rich shelf deposits (Figs 2 and 3).

The subsurface-outcrop investigation of the deep-lacustrine clinoforms focuses on: (i) documenting the characteristics of the shelf to basin-floor deposits in low-accommodation and narrow shelf setting, and reduced salinity conditions; (ii) demonstrating the direct linkage between fluvio-deltaic and sublacustrine sediment gravity flow systems; (iii) investigating controlling parameters of the depositional geometry; and (iv) discussing how the deep-lacustrine depositional system might be distinguished from marine systems.
GEOLOGICAL BACKGROUND

The Dacian Basin is a foreland basin of the South Carpathian Mountains of Romania (Fig. 1A), formed due to a continental collision between the Inner Carpathian unit and East European–Scythian–Moesian platforms during late-Middle Miocene (Sarmatian) (Rabâgia & Maţenco, 1999; Schmid et al., 2008; Jipa & Olariu, 2009). Eastward motion of the Carpathian unit and its collision with the European platforms changed the stress regime along the east–west oriented plate margins from dextral transtension that formed the Getic depression to dextral transpression (north–south compression) (Rabâgia & Maţenco, 1999; Krézsek et al., 2013). This tectonic event resulted in the thrusting of a Cretaceous–Palaeogene sedimentary prism, known as the External Carpathians, onto the Moesian platform and created high flexural subsidence at an average rate of 1400 m/Ma during the late-Middle Miocene (Sarmatian) times (Maţenco et al., 2003). This subsided area of the Moesian platform became the narrow (ca 100 km wide) but deep (up to 1000 m) Dacian Basin (Maţenco et al., 1997; Rabâgia & Maţenco, 1999; Maţenco et al., 2016) (Fig. 1B).

Due to the Carpathian uplift, the Dacian Basin became isolated from the adjacent, the Pannonian Basin of the Central Paratethyan lake, at around 11 Ma (early-Late Miocene) (Olteanu & Jipa, 2006). Later, in late Miocene times, the connection between the Dacian and Black Sea/Euxinian basins was more limited due to an uplift in the Dobrogea area (Fig. 1B). The Dacian–Black Sea intermittent connections occurred through a narrow seaway called the ‘Galaţi passage’ (Popov et al., 2006; Jipa & Olariu, 2009). The separation of the basins caused a decrease in relative base-level in the Dacian Basin during late Miocene times or Upper Sarmatian to Lowermost Meotian (Olteanu & Jipa, 2006). A decrease in the salinity of the Dacian Basin and a change into a brackish or fresh water lacustrine environment from late Miocene (Meotian) times onward were attributed to the basin isolation events (Olteanu & Jipa, 2006). However, Marinescu et al. (1981) and Rögl (1999) suggest that the brackish water conditions, 14 to 18 ‰ salinity, (data from basin margin by Marinescu et al., 1981) of the Dacian Basin is reported to have begun after the basin was formed in the late-Middle Miocene (Sarmatian), before the isolation. The late Miocene–Pliocene (Meotian–Dacian) times were tectonically quiescent until later in the Pliocene (during latest Dacian to Romanian) when the tectonic

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‘Wallachian Phase’ caused an uplift of the basin fill and formed the outcrops along the South Carpathian foothills (Rabăgia & Mațenco, 1999). The late Miocene–Quaternary basin fill attained a thickness up to 3 km (Jipa & Olariu, 2009) and has been reported to consist of near-shore, deltaic and fluvial deposits that were controlled by transgressive–regressive base-level cycles (Jipa & Olariu, 2009), (Fig. 2) with a fauna indicative of overall low palaeowater depth (e.g. Marinescu et al., 1981; Olteanu & Jipa, 2006; Jipa & Olariu, 2009). The basin infill has been classified into lithostratigraphic units that imply horizontal timelines (Fig. 3A). However, publication of shelf-margin scale clinoforms in the western Dacian Basin (e.g. Leever, 2007; Munteanu et al., 2012; Fongngern et al., 2016) suggest that, for the large part of the Upper Miocene stratigraphy, the timelines within the basin fill are not horizontal but inclined and each of the depositional segments (i.e. shelf, slope and basin floor) can be separated as diachronous lithostratigraphic units (Fig. 3B). Fongngern et al. (2016) mapped the Meotian shelf-margin clinoforms that prograded from the Southern Carpathians towards the south and south-west (their Stage 2 and 3 clinoform succession respectively, see also fig. 6b) and interpreted the western Dacian Basin as a deep basin with deepwater depositional environments.

STUDY AREA AND DATASET

The study area is located in the western Dacian Basin between the Olt and Gilort rivers of Romania (Fig. 1C). The datasets consist of: (i) a Mio-Pliocene aged outcrop belt that exposes hundreds of metres of continuous stratigraphy along the northern basin margin; and (ii) three-dimensional seismic volumes located around 30 km south of the outcrops (Fig. 1C). The strata (structural attitude) are orientated almost east–west (90 to 110 azimuth) with a 10° to 15° dip to the south. The focus of this study is the Meotian–Dacian aged basin fill (Fig. 2) characterized by the seismically well-imaged shelf-margin clinoforms in the subsurface data. The entire basin fill succession can be traced out in the north-south aligned river valleys from east to west: the Otăsău, Luncavăt, Cernisoara, Cerna and Tariia river valleys (Fig. 1). The 3D seismic dataset is made up of three overlapping seismic volumes that cover ca 1700 km². With a frequency of 20 to 30 Hz and 2200 m/s average seismic velocity, the highest vertical resolution the seismic data provide is around 20 to 25 m.
METHODOLOGY

The seismic data study of Fongngern et al. (2016) provides the information of clinoform geometry (Fig. 4) and dimensions of large-scale depositional features such as sublacustrine channels and canyons that are difficult to measure from the outcrops. Variance and sweetness attributes (Chopra & Marfurt, 2007) are extracted on the mapped clinoform surfaces and on time slices in the flattened seismic volumes (Fig. 5). Time slices that showed the clearest depositional features were selected from the interval within 10 ms above and below the clinoform surfaces used to flatten the seismic volume. Interference from the overlying and underlying reflectors might have occurred but was kept minimal. The variance attribute shows variability of the acoustic impedance from one seismic trace to another and is useful for detecting features that cause lithological discontinuity such as faults, channels and slump scars. Sweetness is a ratio of reflection strength (or amplitude) and the square root of an instantaneous frequency. In this study, Resistivity (R) and Spontaneous Potential (SP) well logs were used for calibration with the seismic attribute and high sweetness amplitudes represent relatively coarser-grained deposits than dim amplitudes. A combination of variance and sweetness attributes was used to locate channels, slump scars, faults and the distribution of associated sedimentary deposits.

In five river valleys that cut the basin fill (Fig. 1C), sedimentological data including grain size, bed orientation, sedimentary structures, natures of bed contact, trace fossil occurrence and fossils were collected at bed scale (centimetre to decimetre) from the outcrops. The measured sections are 500 to 1000 m thick. In each measured section, the deposits were first separated into facies and then grouped into facies associations. Based on the facies associations and stratigraphic relationships, depositional elements and systems were interpreted. The age constraints on the stratigraphic sections were mainly derived from the 1 : 200,000 scaled geological map of Romania (1967): Tîrgu-Jiu and Pitești maps, and the age dating of detrital volcanic samples. For the subsurface dataset, the ages of seismic reflections rely on a combination of well data biostratigraphy that is propriety of Petrom OMV and magnetostratigraphy work of Vasiliev et al. (2005). The Sarmatian–Meotian age boundary (Fig. 2) was selected as a regional datum for stratigraphic correlation across the river valleys because

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it records a transition from the late Sarmatian basin-wide lowstand (e.g. Olteanu & Jipa, 2006) to a deep basin with seismically imaged clinoforms downlapping onto this chronological surface (e.g. Leever, 2007; Fongngern et al., 2016). It is important to note that the deposits above seismic surface I in Fig. 4, i.e. the Meotian–Pontian age boundary in the subsurface study (Fongngern et al., 2016) also marks a boundary between clinoforms and sub-horizontal to horizontal seismic reflections. Thus, the Pontian–Dacian deposits on the measured sections are most likely not genetically related to the examined deep-lacustrine deposits but probably belong to the topset or shallow-water deposits of younger clinoforms.

Two samples of detrital volcanic sediment were collected at around 50 m above the section bases for age dating (Fig. 6). Uranium–lead (U-Pb) age dating of detrital zircon grains in the samples was carried out using laser ablation–inductively coupled plasma–mass spectrometry (LA-ICP-MS) technique at the (U-Th)/He and U-Pb Geo-Thermochronometry Laboratory, Jackson School of Geosciences, University of Texas at Austin. Any analyses with <95% concordance were discarded and the refined results were plotted as age histograms with probability density distributions (see Appendix). The ages of the youngest zircon grains are interpreted here as the maximum depositional age of the detrital volcanic sediment (Fedo et al., 2003).

RESULTS

BASIN MARGIN CLINOFROMS, DEPOSITIONAL ELEMENTS AND IMAGED DEPOSITIONAL SYSTEMS IN THE SUBSURFACE

Basin margin clinoforms and depositional elements in the subsurface

The seismic data in the western Dacian Basin image basin-margin scale clinoforms (Leever, 2007; Leever et al., 2010) that prograded south-westward from the Southern Carpathians (Figs 1 and 4). The clinoform geometry, dimensions of the depositional elements (summarized in Table 1), cross-sections and time slices of the basin margin have been integrated with the outcrop-lithological data. The lacustrine clinoforms in the western Dacian Basin are different from most marine shelf-margin clinoforms because of their consistently thin topsets (Fig. 4) due to a sill-limiting shelf.
accommodation (Fongngern et al., 2016). With limited shelf accommodation, the oblique clinoforms were strongly progradational with marked aggradational bottomsets and sediment deposition rate in a range of 1620 to 4860 km$^3$/Ma (Jipa & Olariu, 2013). Clinoforms C, D and E (Fig. 4) are selected for examples of the lake clinoform depositional architectures. The interpretation of the shelf-edge and base of slope locations was made based on the clinoform geometry (Figs 4 and 5, see also Fongngern et al., 2016).

**Depositional systems imaged by variance and sweetness seismic attributes**

**Observations**

On Clinoform C variance surface (Fig. 5A), the shelf edge is indented at multiple locations and numerous, 1 to 2 km wide, canyons are present on the slope with a gradient of around 3° (Fongngern et al., 2016). The incisions are interpreted as ‘canyons’ because they are wide and deep (relative to the margin thickness), cut down multiple sequences and contain smaller sediment conduits i.e. channels (see also Fongngern et al., 2016). Towards the south, a channel is discernible right below the incised (ca 50 m deep) shelf edge. On the basin floor, sublacustrine channels can be mapped for over 3 km from the base of slope (Fig. 5A to C). A basin-floor scour is imaged to extend basinward for >10 km from the base of slope. The scour is characterized by 0.4 to 1.5 km wide area with low variance flanked by curved edges with higher variance values (Fig. 5A) and in cross-section, is overlain by chaotic and dim seismic reflections. Sweetness amplitude (Fig. 5B) shows high value on the shelf and near the shelf edge. Channels 3, 11 and 12 are highlighted with linear amplitude patterns on the slope whereas the rest of the channels have dim amplitude (Fig. 5B and C). Most channels show connection with deposits, 2 to 8 km wide and 5 to 10 km long, on the basin floor, even though they have dim amplitudes on the slope (Fig. 5B and C). Fluvial channels and an incised valley are observed on the clinoform D topset and the shelf edge is more deeply incised than on clinoform C (Fig. 5D). The fluvial channels are connected to the 1.0 to 2.5 km wide sublacustrine canyons, with and without channels observed on the slope (Fig. 5D) (Fongngern et al., 2016). In the north, the slope channels imaged by the variance attribute are not brightened up by sweetness attributes but do show connection to deposits on the basin floor. Instead, fairly straight (amplitude patterns) channels are

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observed, especially near the faults (Fig. 5E and F), and possible sandy deposits are observed basinward of these feature (ca 10 km from the base of slope, see Fig. 5F).

The degree of shelf incision and shelf-edge erosion decreases on clinoform E relative to the older clinoforms (Fig. 5G). A fluvial channel can be observed near the shelf edge towards the south (Fig. 5G). The clinoform E foreset (slope) is less rugose than the older clinoforms and fewer slope channels can be observed (Fig. 5G and H). Among them, only channel 2, situated in front of the imaged fluvial channel, and channel 3 show bright amplitudes (Fig. 5F). Basin floor deposits are mainly imaged in the south (Fig. 5H and I).

**Interpretation and discussion**

The seismic data show that the western Dacian Basin fill is characterized by hundreds of metres thick clinothems with coarse-grained deposits present in both topset and basin-floor segments, as suggested by higher sweetness amplitudes. The degree of slope channelization and mass wasting decreased with decreasing gradient as evidenced by progressively lower variance values on the younger clinoform surfaces (Fig. 5). It is possible that more fluvial channels are present along the shelf edge but they are too shallow to be extracted by variance attribute. The high amplitude lobate features on the topset are interpreted to be delta lobe complexes. (Fig. 5C, F and I). The topset and basin floor environments are linked by the channels on the overall fine-grained slope inferred by dimmer amplitudes. The slope channels with high and low sweetness amplitudes are interpreted to be filled with sand-rich and mud-rich sediments and are represented by black and grey line drawings respectively (Fig. 5C, F and I). Both sand-rich and mud-rich channel fills link to the high amplitude lobe shapes that are approximately 2 to 8 km wide and 5 to 10 km long on the basin floor. When comparing their plan-view size to other studies, these deposits are of turbidite lobes hierarchical classification (3.5 to 13 km wide and 5 to 27 km long by Prélat et al. (2009) and 6 to 15 km wide and 9 to 28 km long by Koo et al. (2016). Channels with sand-rich infills, however, show clearer connection to the basin-floor lobes as exemplified by channels 1 to 3 on clinoform E (Fig. 5H and I). This implies that the slope channels normally served as pathways for shelf-derived coarse-grained
sediment to bypass to the basin floor. In addition, channel 2 on clinoform E exhibits a direct linkage between the fluvial and basin-floor environments (Fig. 5G, H and I) and suggests that hyperpycnal flows could be an importation process.

The observed high-amplitude linear features on the basin floor (Fig. 5B, E and H) are interpreted to be channel-levée systems. Since these basin-floor channel-levée systems are flanked by the faults, it is possible the structure controlled the path of the channels by creating topography on the basin floor. The lack of sand-rich basin-floor lobes, for example, on the north-west corner of Fig. 5E maybe because: (i) coarse-sediment was stored on the shelf and/or in the sublacustrine channels; and (ii) the channel-levée systems bypassed sediments onto the floor out of the mapped area. The scours observed on the basin floor (Fig. 5A) are interpreted to have been potentially formed by debris flows as suggested by the overlying chaotic and dim seismic facies. On a low gradient sea-floor, debris flows are known to be able to continue to flow, plough and deposit sediment further away from the base of slope (e.g. Mohrig et al., 1998; Moscardelli et al., 2006). However, limited downdip coverage of the data precludes unequivocal interpretation of the process.

DEPOSITIONAL SYSTEMS IN THE OUTCROPS OF THE MIO-PLIOCENE DACIAN BASIN FILL

The western Dacian Basin fill deposits described in the outcrops are classified into sixteen lithofacies (F) that compose six facies associations (FA). The facies description, photographic examples, type logs, and interpreted depositional processes and environments of five continuous sections are summarized in Table 2. The facies associations are composed of groups of facies that have distinct architectures and imply different depositional environments. The clinoform progradation is towards the south and south-west (Stages 1 and 2 in fig. 4 of Fongngern et al., 2016, respectively); thus the fairly closely spaced (few kilometres) measured sections represent a slightly oblique to depositional-strike section (Fig. 6). The palaeocurrent directions have a distribution from south-east to north-west but are mainly towards the south-east and south-west (Fig. 6). Overall, the facies in the up
to 1000 m thick stratigraphic sections comprise 28% sandstones or conglomeratic sandstones, 14% siltstones and 58% mudstones.

**Facies Association 1: Cross-bedded sandstone and silty mudstone**

Facies Association 1 (FA 1) is only observed towards the 850 to 1000 m interval of the Cernisoara measured section (Figs 1 and 6) and is mainly composed of cross-bedded, coarse-grained sandstone with occasional conglomeratic lenses (F1, Table 2). The cross-bedding foresets are often normally graded (Fig. 7B). The sandstone set thickness and grain size normally decrease upward to ripple cross-laminated or flat-laminated fine-grained sandstone. Amalgamated sandstones with thickness up to 4 m with erosional basal and sharp top contacts (Fig. 7A) are encased in ripple cross-laminated or structureless silty mudstone (F2, Table 2). Disturbance by plant roots on original sedimentary structures of the mudstones were observed. Mudstone beds of 1 to 5 cm thickness can stack to form 1 to 5 m mudstone intervals (Fig. 7A).

**Interpretation and discussion**

The cross-bedded sandstones constitute mainly unidirectional flow bed load of a fluvial-channel fill (Table 2). The overall upward-fining pattern of the deposit suggests point bars (Miall, 1985) interpreted to be within the distributary channels of the lake delta feeders. The mudstones with centimetre-thick siltstone interbeds and plant roots represent floodplain or interdistributary deposits that settled out from suspended sediment, or weak outside channel currents when the river was high above its natural levees (McKee et al., 1967).

**Facies Association 2: Cross-bedded sandstone and laminated mudstone**

Facies Association 2 (FA 2) is abundant in the top 200 m interval of the Luncavăt and Otăsău sections and in the top 100 m interval of the Cerna and Tarăia sections (Figs 1 and 6): FA 2 comprises sets of cross-bedded sandstone (F3), cross-bedded sandstone with mudclasts (F4), sandstone with swaley cross-stratification (SCS) (F6), and laminated mudstones (F5) that are commonly of bluish colour and bioturbated (Table 2). Abundant shells and bioturbation can be found in both sandstone and mudstone facies (Figs 8 and 9). Parallel laminated or structureless mudstones and ripple cross-laminated or

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laminated siltstone (F5) are encountered in the lower part of FA 2 and form up to 10 m thick intervals (Figs 8A and 9). The bioturbated mudstone and occasional 10 to 15 cm thick pebbly sandstone lenses (of F4) with fossil *Viviparus sp.* and bivalve or brachiopod shell fragments become notable towards the top of the mudstone succession. A transition from mudstone to sandstone intervals is typified by low-angle cross-laminated, poorly-sorted pebbly sandstone with shells and mudclasts that gradually grade to parallel or low-angle laminated fine sandstones (F3) (Fig. 8C). Cross-bedded and fossil-rich sandstones (F3), stacked up to 2 to 4 m of total thickness, overlie the poorly sorted sandstones with a sharp contact. The shell fragments are mostly concentrated on the cross-stratal bottomsets where unidirectional ripples are also present. The large cross-bed sets are truncated and overlain by a 20 to 30 cm thick, shell-rich medium-coarse sandstone bed (coquina) (Fig. 8D) which is often overlain by stacked sandstone sets (0.5 to 2.0 m) with trough-cross strata, common shell debris and rare mudclasts and structureless sandstone beds with small mudclasts throughout but rare to absent shell debris (F4): F4 sandstones are commonly interstratified with 30 to 50 cm thick lens-shaped conglomerates with fossil *Viviparus sp.*, interbedded with 1 to 2 cm thick mudstones (Fig. 8E) or with 1 to 4 m thick shell-rich silty mudstone (F5, Fig. 8F).

Trace fossils *Teichichnus* and *Planolites* are commonly found in burrowed F3 deposits (Fig. 9A). *Skolithos* Ichnofacies with *Thalassinoideas, Skolithos, Arenicolites* and *Rosselia* are found in sandstone beds of F3 (Fig. 9B and C). Swaley cross-stratified sandstones (F6) are often found overlain by at least 5 to 7 m thick F5 mudstone intervals (Fig. 9D) of which fossil rich beds are encountered towards the top (Fig. 9E).

*Interpretation and discussion*

Facies Association 2 is interpreted to be lake-shoreline or delta-front deposits. The high content of shells, bioturbation, cross-strata and overall coarser grain size distinguish FA 2 from the overlying and underlying mudstones (Fig. 6) which represent shallow-water and sublacustrine environments, respectively. The tabular sandstones (F3) are interpreted to be formed by jet flows that exited the river mouth and deposited sediment as river mouth bars (Wright, 1977; Wellner et al., 2005; Olariu &
Bhattacharya, 2006). This interpretation is supported by the presence of unidirectional ripples on the inclined master surface (underlying surface) that indicate forward accretion. The sorting of the sandstones and lack of mud also suggests high energy current prior to deposition.

The structureless or laminated mudstones (F5) were deposited from suspended sediment fall-out from the river plumes during or after floods, basinward of the mouth-bar deposits and they represent prodelta sediments (Bhattacharya & Walker, 1992; Orton & Reading, 1993). The poorly sorted coarse sandstones that are interstratified with the prodelta mudstones are interpreted to have formed by resedimentation of a collapsed distributary-channel mouth bar (F3) (Nemec et al., 1988). The mouth-bar is likely to have been truncated and continuously reworked by waves as suggested by erosional base and concentration of the shells in the overlying coarse sandstone (Fig. 8D). The cross-bedded sandstones with clasts (F4) that have sharp to erosional bases are interpreted to be distributary channel fill based on the smaller content of shell debris, the overall upward fining trend (Fig. 8B), the presence of mudstone layers, mud rip-up clasts, trough-cross strata and unidirectional ripples. The parallel laminated sandstones and conglomeratic beds with freshwater fossil Viviparus sp. in the cross-bedded sandstone imply highest energy in FA 2 and a terminal distributary channel environment during river flood (Olariu & Bhattacharya, 2006). River floods could have triggered hyperpycnal flows in front of the deltas (Pattison, 2005; Olariu et al., 2010) and these flows potentially reached the deeper water (Plink-Björklund & Steel, 2004; Petter & Steel, 2006).

In addition to the river dominated characteristic of the deltaic deposits, wave reworking appears to have been significant, particularly in the distal and lateral reaches of the river mouth. This is suggested by the presence of clean sandstones with Skolithos Ichnofacies (Fig. 9B and C) that are commonly found below the mouth bar deposits and/or interstratified with prodeltaic mudstones. *Skolithos* trace fossils are common in sandy shoreface environment (Pemberton et al., 2012). High wave energy (storm) is also indicated by the presence of shoreface sandstones with SCS (Fig. 9D, Greenwood & Sherman, 1986). The mudstone intervals that overly the sandstones with SCS and the intercalated fossil-rich beds (Fig. 9E) imply transgression and deepening of the shoreline that allowed waves to rework and concentrate large number of fossils in one bed.

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Facies Association 3: Thick sandstone with erosional base

Beds and bed sets of structureless sandstone (F7), graded sandstone (F8) and thin-bedded sandstone (F9) (Table 2; Fig. 10) constitute FA 3 that commonly reach thickness of 10 to 20 m (Fig. 6). Facies Association 3 is encountered between 300 to 800 m, 100 to 800 m and 300 to 700 m on the Tariia, Cerna and Cernisoara stratigraphic sections, respectively (Fig. 6). The sandstone bed sets have lens-shaped geometry when followed laterally in the outcrop for over 50 m and the deposits onlap or downlap onto erosional bases. Graded sandstone beds (F8) are normally graded from medium-grained to very fine-grained sandstone and observed sedimentary structures from base to top include structureless, parallel laminations, low-angle laminations unidirectional ripple cross-laminations and climbing ripple cross-lamination. Internal weakly erosional but sharp contacts, are sometimes observed within the sandstone bodies (for example, Fig. 10B).

The deposits of FA 3 are commonly underlain by internally deformed mud-rich deposits with rotated mudstone and sandstone blocks (F15, Fig. 10A and C), thin-bedded and 10 to 30 cm thick structureless sandstone beds (F9) (Fig. 10D) and/or thin-bedded, 1 to 5 cm, siltstone (F10). The F7 structureless beds (up to 80 cm thick), contain large, 5 to 20 cm, round-shaped or slightly angular-shaped mudclasts, and sand quartz granules near the erosional base (Fig. 10D). The amalgamated structureless beds can form 2 to 4 m thick successions. Occasionally, the laminations of F8 sandstones are less well-defined, thickening to 2 to 3 cm and showing normal grading (Table 2; Fig. 10B). The laminated intervals gradually change to an interval of 1 to 3 cm thick, fine-grained to very-fine grained sandstone with ripple cross-laminations (RCL) or climbing-ripple cross-laminations (CRCL) (Fig. 10B). If not truncated and/or overlain by the same facies, F8 often vertically grades to laminated siltstone–mudstone layers (F9).

Interpretation and discussion

Facies Association 3 deposits are interpreted to have been deposited by turbidity currents that at a specific location, waned and lost their capacity to carry sediment and thus the sediment settles out of the flow and deposits (e.g. Hiscott, 1994). The depositional process is indicated by the nature of the
deposits of being non-graded but clean and graded with distinctive sedimentary structures. The presence of the sandstones and silty mudstones within and onlapping the incisional surfaces suggests deposition in confined turbidite channels.

Following Bouma divisions (Bouma, 1962), the structureless sandstone of F7 and F8 (Table 2) are Ta turbidites. Within planar laminated sandstone beds (Tb), there are thicker (1 to 5 cm) and less well-developed laminated intervals that have intermediate characteristic between Ta and Tb divisions and might indicate higher bed aggradation rates than the normal well-defined laminations (i.e. spaced laminations of Sumner et al., 2008). Repetition of Ta, Tb and Tc beds and internal weakly erosional contacts within graded sandstone (for example, in Tc interval in Fig. 10B) suggest that the turbidites were deposited from a quasi-steady sustained turbidity current (Kneller & Branney, 1995; Mulder & Alexander, 2001). The deformed mud-rich deposits (F15) are interpreted to be mass-transport deposits. The mass-transport deposits with highly deformed sandstone and mudstone blocks and gravels (for example, Fig. 10A and C) could have been originated by a collapse of deposits from the upslope locations with some distance away such as the shelf-edge or upper slope (Mulder & Cochonat, 1996), whereas the deposits with bended or folded blocks of facies that is lateral to or not far in the vertical succession with undeformed turbidites (such as F10) might suggest an in situ origin such as from collapse of channel margin or levées (Deptuck et al., 2003).

**Facies Association 4: Thin-bedded sandstone and mudstone**

Facies Association 4 (FA 4) deposits are typically stacked couplets of sandstone and mudstone of F9 and F10 (Table 2) that form up to 100 m thick successions (Figs 6 and 11): FA 4 is found alternating with FA 3 and, by thickness, is the most dominant in the stratigraphy between ca 500 to 900 m, 500 to 700 m and 400 to 800 m on Tarfia, Cerna and Cernisoara measured respectively (Fig. 6). Thin-bedded sandstones of FA 4 (F9; for example Fig. 11A and B) are found adjacent to FA 3 and, even though lateral facies transition is often subtle, they can be differentiated by their thinner but more constant bed thickness characteristics.
Thin-bedded sandstone (F9; Table 2) can form a package of up to 20 to 30 m thick and vertically grade to the more mud-rich F10 that reaches 30 to 70 m thickness. Facies F10 is dominated by thin-bedded (2 to 3 cm) ripple cross-laminated sandstone and mudstone beds (Fig. 11B) which can be traced horizontally for a significant distance (>200 m). The turbidite beds are often truncated and overlain by successive discordant beds (Fig. 11B). The ripples in the F10 very-fine sandstones or siltstones are often overturned or sand-starved (Fig. 11C). Laminated siltstone or silty mudstone (Td) with common flame structures are occasionally found above Tc or directly on top of Tb (Fig. 11D). The presence of flames (Fig. 11D), wood detritus (Fig. 11E), folded beds or laminae and water escape structures are common. At some locations, especially lateral to FA3, sandstone with mudclasts towards bed tops and carbonaceous debris-rich sandstone (F11) and argillaceous sandstones (F14) are found interstratified with F9 and F10 (Figs 11D and F).

Interpretation and discussion

The thin-bedded sandstones and siltstones (F9 and F10) are interpreted to be a product of dilute turbidity currents that built up sublacustrine channel-levées through flow stripping (Piper & Normark, 1983) and overspill (Hiscott et al., 1997) processes. Both processes are due to the inertial force of the flow that tends to preserve its initial path, but flow stripping occurs continuously along the channel length and is dominantly controlled by pre-existing levée height. Conversely, overspill occurs preferentially at a sharp outer bend (e.g. Piper & Normark, 1983; Peakall et al., 2000; Di Celma et al., 2013; Morris et al., 2014). Characteristics of the channel-levée deposits in the Dacian Basin is their proximity to the deposits interpreted as turbidite channels, their thin bed thickness, the presence of small scours, and wedging bed-set geometry typified by discordant bed contacts often observed within 20 to 30 m lateral from the channel margin (for example, Fig. 11B). The channel-levée deposits in the Dacian Basin make up around 40 to 60% of the entire deep-lacustrine deposits. The F9 and F10 beds represent deposits that were formed relatively close and farther away to the channel axis, so called channel-proximal levée and channel-distal levée, respectively (Kane et al., 2007). The climbing-ripple cross-laminations and flame structures in F9 beds indicate high depositional rates. Besides abundant tractional beds, the proximal levée deposits sometimes contain
structureless sandstones (Ta) which differentiates them from the distal levée deposits. The latter is characterized by faint laminations, starved ripples and thinner but more constant bed thickness (Table 2, see also Kane et al., 2007).

**Facies Association 5: Tabular sandstone**

Facies Association 5 (FA 5) is encountered from the bottom of the Tariia, Cerna and Cernisoara measured sections to 300 to 500 m level (Fig. 6): FA 5 includes tabular sandstones (F13), argillaceous or banded sandstone (F14), tabular argillaceous sandstone with mudclasts (F11), structureless sandstone with clasts (F7), thin-bedded siltstone (F10) and laminated mudstone (F12) (see Table 2; Fig. 12). Flame structures and dewatering pipes are ubiquitous (for example, Fig. 12A and B). Shell fragments are abundant in some sandstone beds (for example, Fig. 12C). Mudstone clasts are often found near the sandstone bases, particularly in those very-coarse grained to pebbly sandstones (F7) overlying finer grained deposits or deformed deposits (Fig. 12D and E). The F11 sandstone beds are normally capped by a thin, 0.5 to 1.0 cm, mudstone and carbonaceous or mudclast breccia. The breccias have similar characteristics to the top of F11 beds but have greater concentration of mudclasts and carbonaceous debris, and thus named here F11a. The argillaceous and banded sandstones (F14) contain clay, mudstone clasts, and carbonaceous fragments that form dark layers alternating with lighter and cleaner very fine to fine-grained sandstones towards the top of the beds (Fig. 12F). The sandstone has sharp to slightly erosional contact with an underlying mudstone bed, on which flame structures are common, and is capped by thin-bedded mudstone. A vertical stack of the individual F11, F13 and F14 sandstone beds, occasionally interbedded with thin (centimetre) F10 siltstone beds, can attain 2 to 8 m thickness and all of these sandstone facies can stack up to an average thickness of 15 to 30 m and are capped by tens of centimetres thick thin-bedded siltstones (F10) or mudstones (F12).
Interpretation and discussion

The thick normally graded sandstones of F13 are turbidites deposited where turbidity currents decelerated or expanded and deposited sediment as they experience slope gradient decrease or loss of confinement (cf. Piper & Normark, 1983). Flames and dewatering structures are common in high concentration turbidites and suggest rapid sedimentation rate (Lowe, 1975). The finer grained sediments (F10 and F12) were deposited from the more dilute tails of waning turbidity currents. The sandstones with mudclast dominated tops (F11), mudclast breccia (F11a) and argillaceous banded sandstones (F14) are hybrid event beds and classified as H1b, H3 and H2, respectively (Haughton et al., 2009; Fonnesu et al., 2015). The absence of a clear bed boundary with underlying turbidite sandstone indicates that the mudclast-rich argillaceous sandstone or mudclasts breccia, so called linked debrites, were emplaced almost instantaneously after the turbidite (Haughton et al., 2003; Talling et al., 2004; Amy & Talling, 2006; Haughton et al., 2009). This suggests that the deposit could have derived as co-genetic sediment gravity flows or a flow that went through transformation (Baas & Best, 2002; Mohrig & Marr, 2003). The banded and argillaceous sandstones (F14) are interpreted to have been formed by transitional flows in which the flow turbulence was periodically dampened by high sediment concentration or by increasing clay particles from disintegrated mudclasts (e.g. Lowe & Guy, 2000; Haughton et al., 2003; Davis et al., 2009; Haughton et al., 2009; Kane & Ponten, 2012; Sumner et al., 2012). A periodic turbulence suppression by the clay particles within the flow produces dark and light bands observed in the transitional flow deposit (F14 or H2 of Haughton et al., 2009).

At an outcrop location where up to 10 m of stratigraphy can be observed, turbidite sandstones are sheet-like, interbedded with thin (0.5 to 2.0 cm thick), locally erosional, siltstones or silty mudstones and are interpreted to compose a basin-floor ‘lobe’. This is supported by their intercalation or juxtaposition to the transitional flow deposits, facies reported as commonly located on distal and lateral fan fringe settings (e.g. Haughton et al., 2003; Haughton et al., 2009) and channel to lobe transition zone (CLTZ) (e.g. Jackson et al., 2009; Terlaky & Arnott, 2014; Fonnesu et al., 2015). Over 10 m thickness, sandstones become thinner while mudstone beds thicken to 5 to 15 cm (see type lithological logs of F11 and F13 in Table 2). This could represent an interlobe site (cf. Prélat et al.,

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2009) and indicate decreasing sandy sediment supply to the locality. For the entire recorded stratigraphy (Fig. 6), the basin-floor deposits form 30 to 50 m thick lobe complex(es) encased in thick mudstone accumulation (cf. Prélat et al., 2009).

Carbonaceous material in the transitional flow beds indicates that the sediment was derived from a continental or shallow water area. Hence, the sediment gravity flows in the Dacian Basin could have been triggered by failure of the shelf edge and upper slope or the sediment was sourced directly from rivers.

**Facies Association 6: Chaotic and deformed deposit**

In addition to FA 5, the chaotic and deformed facies of FA 6 (F15 and F16; Table 2) make up significant portions of the lowermost 300 to 350 m of the stratigraphy but are also occasionally found throughout the 0 to 700 m interval of the Tariia, Cerna and Cernisoara sections. The deposits are mainly mud-rich (F15), but a sand-rich (F16) example is shown in Fig. 13A. Coarser chaotic deposits (F16) have varying characteristics including matrix-supported, mud-clast dominated conglomerates and structureless sandstones (Fig. 13A), and are around 5 to 10 m thick on average. Mudclasts, 5 to 25 cm long, are the most abundant (Fig. 13A) but clasts of pebble-sized quartz, granitic and metamorphic rock are occasionally observed. Patches of the mudclasts and small pebbles are typically observed in the structureless sandstones. Intact shells of bivalves and gastropods and wood detritus are also present. The amalgamated, up to 30 m thick, conglomerate or sandstone beds are normally emplaced above laminated mudstone or thin-bedded siltstone (Fig. 13A). The chaotic muddy deposits contain deformed blocks of mostly thin-bedded fine-grained facies such as F10 or F12; however, blocks of sandstone are occasionally observed (Fig. 13B to E). These deformed muddy deposits are normally overlain by thin-bedded siltstone (Fig. 13B), structureless sandstone (F7) (Fig. 13E) or graded sandstone (F8) (Fig. 13C and D) and tabular sandstones (F13).
Interpretation and discussion

The chaotic muddy (F15) and sandy (F16) facies were deposited from cohesive and non-cohesive or sandy debris flows, respectively (Mulder & Alexander, 2001; Talling et al., 2012). The sand-rich deposits that contain shells or well-rounded pebbles (for example, Fig. 13A) suggest that the debris flows could have been generated from mass-wasting at the shelf-edge and speculatively during floods when deposition rate was high. Deformed mud-rich deposits that contain contorted blocks of thin-bedded siltstone and mudstone such as F10 (Table 3.2) are interpreted to have been formed by collapse of canyon walls or channel-levées of which F10 consists. The mass-transport deposits (MTDs) are occasionally found truncated and overlain by structureless, coarse-grained, sandstones with mudclasts interpreted to be channelized turbidites (Figs 8A, 8C, 14E and 15). This suggests the possibility that either: (i) the MTDs were emplaced in proximity to the base of slope where the turbidity currents remained confined or semi-confined; or (ii) the MTDs created at least a minor topographic relief on the lacustrine floor.

Shelf to basin-floor architectures and depositional elements

The combination of the 3D seismic and outcrop facies observations allows detailed reconstruction of the depositional architecture of this fluvial-dominated deep lacustrine fill. Here, the focus is on the slope and basin floor deposits because they represent the highest thickness within the stratigraphy and they have not previously been documented in the Dacian Basin. The seismic data provide the width, length and depth of the depositional elements but not the lithology or thickness of sub-seismic features. The latter were obtained from the outcrops. A summary of the depositional elements on the lake clinoforms is shown in Fig. 14. Amplitude and variance surfaces of clinoform E (Fig. 14A and B) display coarse-grained sediment accumulation and erosional features. Seismic cross-sections of key depositional elements, their sedimentary logs from the outcrop and dimensions are shown in Fig. 14C to F.
The Dacian lacustrine clinoform evolution, under conditions of narrow morphological shelf and its limited accommodation, is characterized by persistent coarse-grained sediment transport from the shelf to the deep lacustrine environments and relatively fast shelf-edge migration of 7 to 8 km/Myr (Fongngern et al., 2016). The efficiency of sediment bypass across the topset of the 400 m high clinoforms, without significant aggradation or erosion, is consistent with the Dacian flat, non-aggradational shelf-edge trajectory (see also Henriksen et al., 2009). The data and interpretation of the depositional systems based on the stratigraphic relationship of the depositional elements were integrated with the north-west/south-east depositional-strike seismic cross-section (fig. 6B of Fongngern et al., 2016) and the depositional dip section (Fig. 4) both located ca 40 km south of the field area (see the location in Fig. 1). The stratigraphic logs (Fig. 6) were placed on the strike cross-section with their actual space distances. Note that the Otăsău section is too far east to be included on the cross-section. The strike and dip views of the lake fill, constructed based on the sedimentological data, the well log patterns and the seismic reflection characteristics are shown in Fig. 15A and B, respectively. On the measured sections in Fig. 15A, the uppermost 200 m of the stratigraphy is dominated by fluvio-deltaic and shoreline deposits of which only the lower 100 m (below surface I) are genetically linked to the deep-lacustrine deposits on the clinoforms (Fig. 15). The fluvial deposits above this surface are suggested to be part of the subsequent basin fill when the Dacian Basin was essentially infilled and then too shallow for shelf-margin clinoforms to develop. The sandy-pebbly fluvio-deltaic deposits correspond to the sub-horizontal high amplitude reflections on the topsets (Figs 6 and 14). A large incision on the deltaic sequences close to the centre of Fig. 15A is interpreted to be an incised valley that was initially filled with fluvial facies.

Fluvio-deltaic, shoreface, and offshore deposits constitute the clinoform topsets. Deltaic deposits in the Dacian Basin tend to have a sharp base (rather than a gradual upward coarsening trend) and thin delta front (Fig. 8B). Mudstones 1 to 7 m thick interstratified with deltaic or shoreface sandstones probably indicate deepening or transgression of the shoreline, particularly when fossil-rich beds are present (for example, Figs 8D and 9F).
Sediment gravity flow channels are ubiquitous on the Dacian basin deep-lacustrine slope (Fig. 15). The medium-grained to very coarse-grained sublacustrine sandstones documented on the measured sections can be part of the canyon fills or channel-levée complexes seen in the seismic data, though the lensoid isolation of these sandstone bodies prevents accurate correlation between the stratigraphic sections, and the seismic images suggest that they are most likely to be parts of different systems (Fig. 15A). In this dataset, it is common that shelf-indenting sublacustrine channels (some are within canyons) are present along the lake-margin shelf edge. Fluvial channels are sited along this zone connecting to the slope channels and thus they serve as a link between the fluvio-deltaic and the basin floor environments (see example of a connection between fluvial and slope channels in Fig. 15B). This shelf-slope configuration also suggests that besides relatively thin deltaic and shoreline deposits, sediment consistently bypassed the shelf to the deep-lacustrine environment and, significantly, through direct connection between fluvial and sublacustrine conduits. In other words, hyperpycnal flow (e.g. Mulder & Syvitski, 1995; Plink-Björklund & Steel, 2004) is likely to have been an important process for sediment transport to the basin floor in addition to surge-type turbidity currents and debris flow generated from collapses of the shelf edge and upper slope. Abundant terrigenous materials found in the deep lacustrine deposits (for example, Figs 11E and F), the presence of well-rounded pebbles in turbidite beds, and the reduced salinity condition of the lake support the idea of hyperpycnal-flow transported sediment. Moreover, because the Dacian Basin’s narrow shelf, shoreline was located only a few kilometres away or at the shelf edge which are the optimal position to promote coarse-grained sediment transport (Sweet & Blum, 2016) by hyperpycnal plumes through the slope conduits to the basin floor.

Thin-bedded very-fine sand and silty turbidites (F9 and F10) are the most abundant facies on the Dacian deep-lacustrine slope. Domination of the fine-grained sediment on the slope suggests that coarse-grained sediment transport over the shelf edge bypassed to the basin floor. Moreover, because of the presence of closely spaced slope channels, parts of the levée deposits resulted from the interference of adjacent sediment gravity flow systems (Fig. 15B). As a result, the channel complexes appear to be aggradational and the high levées could have increased bypass efficiency of the systems.

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(see also Hodgson et al., 2016). A transition of the channel-levée facies to turbidite lobes and lobe complexes occurs in a lower slope-proximal basin floor position (Fig. 15B). The turbidite lobes, but not in all clinothems, are made of sandstones with grain size range from very-fine sand to granule and they correspond to high seismic amplitudes lobe imaged on the basin floor (for example, Fig. 5).

Collapses of the shelf edge or upper slope (see the high variance amplitude near the clinoform D rollover in Fig. 5D) and collapses of canyon walls or channel levées are invoked for generation of mass-transport deposits found throughout the deep-lacustrine succession. Sand-rich (coarse-sand to small granule) debrites are also observed on the basin floor (Figs 6 and 13A). The common motif of turbidite beds above mass-transport deposit observed in the outcrops (for example, Figs 12E, 13C and 13D) probably suggests possible co-genetic relationship of the turbidity current and debris flows (Talling et al., 2013).

Due to the multiple sediment entry points along the shelf edge (for example, 200 m wide fluvial channels and localized shelf-edge collapse in Fig. 5), contemporaneous but varying sediment gravity flow systems formed turbidite lobes, sandy and muddy debrites and thin-bedded mudstone on the basin floor. Consequently, lateral facies changes can occur abruptly or within a few kilometres distance (Fig. 15B). An analogous case of various turbidite systems building out onto one basin floor has also been reported from late Quaternary Lake Baikal where axial and transverse sedimentary systems simultaneously delivered sediment to the lake floor (Nelson et al., 1999).

**DISCUSSION AND CONCLUSIONS**

**Depositional geometry of the basin fill and controlling factors**

*Accommodation versus sediment supply*

The late Miocene-Pliocene lake Dacian Basin fill is characterized by the series of 400 m high oblique clinoforms with prominent bottomset aggradation or the sedimentary wedge consisting of shelfal to deep-lacustrine deposits in about a 1:5 thickness ratio (Figs 6 and 15). The basin-fill depositional geometry and facies suggest an interplay between the controlling factors that are accommodation...
being outpaced by sedimentation supply. As a result, a significant amount of sediment bypassed to the deep-lacustrine environments. Tectonic activities have been reported minimal and the subsidence was down to 150 m/Myr during the late Miocene times after the uplift ceased around 12 Ma (Mațenco et al., 2003). Almost the entirety of sediments eroded from the Southern Carpathians was fed to the Dacian Basin (Jipa & Olariu, 2013). The other parameter that controls base-level in lakes is water supply into the basin (which is derived from direct river flow or through a connection with other lake basins, e.g. Carroll and Bohac, 1999; Bohac et al., 2000). Periodic connection with the neighbouring Paratethyan lakes, the Pannonian Basin (Müller et al., 1999) and the Black Sea (Saulea et al., 1969; Vasiliev et al., 2010), have been speculated. However, this study is in agreement with Jipa & Olariu (2013) that an influence of the water supply through the connection on the base level of the Dacian Basin was minimal because: (i) the study of late Miocene to early Pliocene clinoforms in the Pannonian Basin (Sztanó et al., 2013) shows that the clinoforms have prominently aggradational topsets (>400 m thick) which reflects ample accommodation from long-term rising base-level; and (ii) the Dacian clinoforms did not record the same major base-level drop (>1000 m) in the Black Sea during the Messinian Salinity crisis (Bartol et al., 2012). This implies that despite high water supply into the basin through rivers or inter-basin passages, the western Dacian Basin had certain capacity to store water as the basin water level was increasing with the rate a little higher than sediment accumulation rate. Thus, the excess water drained off to the adjacent Black Sea, i.e. overfilled basin type of Bohacs et al. (2000). Due to the presence of the topographic barrier or sill between these lake basins (see also Fongngern et al., 2016), when the base level in the other basins dropped below the sill level, the Dacian Basin became isolated and its base level became a function of basin catchment precipitation and river inputs.

Climate’s effects on lake sedimentation

Because the supplies of sediment and water to a lake depend on river input into the basin, sediment supply increases when water-level rises (Scholz et al., 1990; Bohacs et al., 2000). Thus, it is commonly known that higher amount of sediments is likely to be delivered to the lake floor during periods of wetter climate. In the adjacent Pannonian basin (see Fig. 1B) the clinothems of similar age...
have been reported to comprise aggradational topsets and long bottomsets with slope-detached basinal lobes interpreted to be formed by efficient turbidity currents during base-level rise (Sztanó et al., 2013). The same scenario seems to hold for this Dacian Basin study. Clinothems 2 and 4 or the deposits above clinoforms C and E (Fig. 6B and H) that exhibit relatively high amount of topset aggradation ca 60 m and 30 m, respectively (table 1 of Fongngern et al., 2016) have fluvio-deltaic system or the shoreline situated at the shelf-edge. This configuration means that sediment flux to the shoreline could easily be delivered over the shelf edge. Increased frequency of flood events during wet climate could have enhanced generation of hyperpycnal-flow derived turbidity currents (Mulder & Syvitski, 1995) and consequently more coarse-grained deposits are observed on the basin floor (Fig. 6B and H). In contrary, the shelf-edge delta and basin-floor turbidite lobes are less prominent within Clinothem 3 or the deposits above Clinoform D (Fig. 6E) that has attained topset aggradation <10 m, implying lower base-level rise and thus less sediment input.

*Impacts of high-frequency climatic cycles on lake base-level changes*

Climate driven high-frequency base-level changes have been commonly reported to influence depositional systems in lakes (e.g. Sáez & Cabrera, 2002). However, in this study the shelf-edge trajectory analysis (Fig. 4) shows that the clinoforms were formed under normal regression condition, except for a forced regression during a deposition of Clinothem 5 (see also Fongngern et al., 2016). This suggests that during the formation of this sedimentary wedge (ca 3 Myr duration), a base-level fall below the shelf-edge, observable under the seismic resolution, was rare despite the thin topset geometry that reflects a shallow-water shelf. It also means that high-frequency climatic cycles did not influence significant base-level change during the time span of the individual clinothems of around >200 kyr (Fongngern et al., 2016). Although, lake-level fluctuations could have occurred in high frequency (in >200 kyr), the magnitudes might have been too low to affect the overall geometry of the lake clinoforms (see also ter Borgh et al., 2014).
Comparison of the Dacian Basin clinothems to marine shelf-margin examples

One of basic criteria to identify lake-basin fills is a lack of tidally influenced deposits, unless the basin has been periodically connected to an open-marine basin (Talbot & Allen, 1996). When comparing the Dacian Basin clinothems with their marine counterparts (approximately the same relief and in a tectonic foreland setting) the following characteristics may help indicating clinothems in the deep lake.

(i) *Dominant process and vertical facies transition*

The presence of fluvial dominated topset deposits (river and delta) and subordinate wave facies might be typical of lakes. Fluvial channels on the upper slope suggest a prolonged siting of the fluvial feeder system at the shelf edge and its influence on building out the clinothems. As a consequence, vertical facies shift from upper-slope mudstones or sandy-turbidite channel fills to middle-inner shelf delta front or fluvial-channel fills is abrupt (Figs 6, 8 and 15). Even though occurrence of the fluvial feeders above upper slope deposit is exhibited in marine clinothems as well, it is often the case that other process such as wave dominates the previous or succeeding clinothems (see example in the Karoo Basin by Jones et al., 2013).

(ii) *Sediment partitioning*

The thickness of deposits in the shelf, slope and basin floor of the Dacian clinothems is around 1:3:3 whereas the ratio 1:1:1 has been reported from a study of Lewis-Fox Hills clinothems (Carvajal & Steel, 2012). Besides low accommodation on the shelf, another explanation for high amount of sediment accumulation in the deep-lacustrine compartments could be that the brackish to fresh water of the lake basin allowed hyperpycnal flows to have a long run-out length without being stratified and lifted off by the interstitial fluid (freshwater) buoyancy (e.g. Spark et al., 1993; McLeod et al., 1999; Steel et al., 2017). A reverse buoyancy (freshwater) interstitial fluid is speculated to cause hyperpycnal flows to
cease and deposit sediment on the slope in the Central Eocene of Spitzbergen (Plink-Blörklund et al., 2001)

(iii) Sediment gravity flow activities

Due to the persisting low accommodation and narrow width of the shelf, a significant portion of the basin’s sediment supply was transported to the deep-lacustrine environments. Deep-lacustrine deposits derived from interference between by multiple sediment-gravity flow systems on the slope and basin floor (Fig. 15A) (e.g. the Lake Baikal study of Nelson et al., 1990). Although not emphasized in this study, a lake-basin infilling can be a process of multiple clinoform sets prograding from around the basin margin as reported in the Pannonian Basin by (Magyar et al., 2013; ter Borgh et al., 2014). In addition, homogeneous hemipelagic mudstones were rarely observed in the field and is speculated to be due to frequent sediment-gravity flow activities, however, more detailed studies are needed to confirm this.

Furthermore, another characteristic that is believed to typify lake clinoforms is an alternation between progradational and aggradational clinothems, reflecting drier and wetter climatic cycles, and it is very well-illustrated in the late Miocene Pannonian Basin clinoforms (Sztano et al., 2013; ter Bourgh et al., 2014). This trait is not obvious in the Dacian clinothems, perhaps because the water level changes by similar climate cycle frequencies were too small in amplitude to allow such geometry to form or even if they were, cannot be detected under the seismic resolution.

In summary, this study is the first to document and interpret deep-lacustrine deposits that make up 70 to 80% of the stratigraphy in the lacustrine western Dacian Basin. Description of linked lacustrine shelf-slope–basin floor depositional systems from integrated outcrop and 3D seismic data is provided. The key points from this investigation include:

1. Limited accommodation on the narrow lake-clinoform topsets along with high sediment supply resulted in sediment bypassing the shelf and fast prograding shelf-margin clinoforms. Ubiquitous channels and canyons on the slope promoted sediment partitioning to the notably thick clinoform bottomsets.

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2. The intercalated deltaic and shoreface sandstones with transgressive mudstone intervals indicate deepening and shallowing up base-level cycles (5 to 10 m amplitude) which cannot be observed in the seismic data. Channel-levée deposits dominate the deep-lacustrine slope and abruptly or gradually change into turbidite and sandy debrite lobes on the lacustrine floor. The presence of mass transport deposits throughout the deep-lacustrine stratigraphy suggests common mass removal from the shelf edge-upper slope areas and collapse of channel levée deposits.

3. The common direct connection of fluvial and turbidite channels at the shelf edge observed in the seismic data and, probably but not always, terrestrial material in the sediment gravity-flow deposits suggest that hyperpycnal flows were frequently generated in the lake basin. The reduced water salinity in the basin played an important role in sediment transport to the highly aggrading deep-lacustrine floor.

4. Sediment supply was the main driver in out-building the shelf-margin clinothems in this low-accommodation setting. The influence of base-level change due to the possible connection with the adjacent Paratethyan basins and the climatic cycles including the Messinian Salinity Crisis is thought to be minimal or not observed on the clinoforms.

5. From this study, characteristics that may be indications of deep-lacustrine deposits include: fluvial-dominated clinoform topsets with abrupt vertical facies shift, bottomset-dominated sediment partitioning on the clinoforms, and frequent sediment gravity flow activities denoted by closely-spaced and aggradational channel-levée systems, thick bottomset deposits and rare indication of sediment starvation in the deep-lacustrine environment.

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FIGURE CAPTIONS

Fig. 1: Location of the study area. (A) Location of the western Dacian Basin in Eastern Europe in a rectangle. (B) Palaeo-geographic setting of the Dacian Basin, the neighbouring Paratethyan lakes and possible communicating passages during late-Miocene to early Pliocene times (modified from Jipa & Olariu, 2013). (C) Geological map of the western Dacian Basin (modified from geological map of Romania, 1967). The southward dipping Meotian–Dacian outcrop belt provides lithological data of the shelf-margin clinoforms imaged in the 3D seismic dataset situated 30 km in the south. The faults are from fig. 3 of Krézsek et al. (2013). Note the location of the data present in Figs 2, 4, 5, 14 and 15. PCL = Peri-Carpathian Line or the frontline of the basal decollement thrust of the Getic Depression (Krézsek et al., 2013).

Fig. 2: Late Miocene–Pliocene (Meotian–Dacian) study interval previously interpreted to be formed by transgressive–regressive shoreline cycles. However, a kilometre-thick stratigraphic log from this study shows a continuum of shelf to basin-floor deposits and the boundary between the shallow and deepwater environments nearly corresponds to the Pontian–Dacian boundary. An absolute age from zircon grains in the detrital volcanic sediment samples is shown. See the location of the stratigraphic log in Fig. 1.

Fig. 3: Schematic models of: (A) layer-cake transgressive–regressive sedimentary units; (B) prograding clinoforms. The significant differences between the two models are: (i) the time lines show contrasting growth style for the basin fill; and (ii) the linkage between depositional environments that defines basin depth and sediment supply; (C) basin-floor to shelf facies that can be walked out through the south-dipping strata in the river valleys.
Fig. 4: Depositional dip cross-section of the Dacian Basin clinoforms (from Fig. 5B of Fongngern et al., 2016). The lacustrine clinoforms are characterized by an oblique geometry with consistently thin topsets and thick bottomsets. The clinoform foresets are erosional due to sediment-gravity flow activities (see Fongngern et al., 2016). Numbers above each clinothem (C) represent amount of topset aggradation in meters from Fongngern et al. (2016).

Fig. 5: Clinoform surfaces with variance attribute (left), with variance and sweetness attributes (centre) and an interpretation of depositional elements (right). The black arrows point towards the north. High degree variance means more discontinuity of the seismic traces or unsmooth surface and high sweetness amplitudes imply relatively coarser grained deposits. (A) to (C) Clinoform C; (D) to (F) Clinoform D; (G) to (I) Clinoform E (see Figs 1 and 4 for the map location and a cross-sectional view of the clinoforms).

Fig. 6: Simplified measured sections with paleocurrent observations. The measured sections make up a cross-sectional view that is slightly oblique to the depositional strike orientation. The age boundaries marked by the arrows are from the geological map of Romania (1967). The Sarmatian–Meotian boundary is selected as a datum. See Table 2 for legend and text for detail. Note that the age boundaries are mark with the same (arrow) colours as those in Fig. 3.

Fig. 7: Examples of the Facies Association 1 (FA 1) fluvial deposits. (A) Fluvial channel fills (2 to 3 m thick) encased in flood-plain mudstones. (B) and (C) Trough cross-bedding and conglomeratic lenses found in the fluvial channels, the palaeoflow is out of the outcrop towards the right (south-east). The scales in (B) and (C) are 15 cm and 160 cm long, respectively.

Fig. 8: Outcrop of a shelf delta. (A) Deltaic sequence, ca 15 m thick in this view, sharply overlie 10 to 20 m thick offshore lacustrine mudstone. (B) Stratigraphic log of the outcrop shown in (A). The three digit numbers by the log are palaeocurrent directions. (C) Poorly-sorted pebbly sandstone beds with fossil Viviparus sp., other shell fragments and mudclasts occur at the base of sand-rich deltaic deposit. (D) Very-coarse to coarse-grained sandstone with large cross-beds and shell fragments are truncated and overlain by fossiliferous sandstone. (E) Lens-shaped conglomerate with fossil gastropods, 1 to 2 cm thick mudstone intercalation and erosional base above mudstone. (F) Shell-rich silty mudstone.

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Fig. 9: Shoreline deposits and trace fossils: (A) *Teichichnus* (T) and *Planolites* (P) in structureless silty mudstone containing a sand-filled groove of unknown origin. The pencil is 15 cm long. (B) *Thalassinoides* (Th), *Skolithos* (S) and *Arenicolites* (Ar) in medium-grained sandstone bed. (C) *Skolithos* (S) and *Rosselia* (R) in fine-grained sandstone bed. (D) Fine-grained sandstone with swaley cross-stratifications and an overlying bioturbated sandstone, all overlain by 5 to 7 m thick mudstone. (E) Shell-rich bed found towards the top of the mudstone succession above (D). A division of the measuring stick in (D) and (E) is 10 cm.

Fig. 10: Channelized turbidite deposits (FA 3). (A) Graded turbidite sandstone (F8) that shows a mud-rich zone above the contact with the mass-transport deposit. The pencil is 15 cm long. (B) Close-up view of the turbidite in Fig. 8A with a sketch of sedimentary structures and turbidite divisions. The white arrows point to minor erosional surfaces. (C) Example of a common occurrence of channelized turbidite above a mass transport deposit in the Dacian Basin. (D) Coarse to very coarse-grained sandy turbidite channel deposit (cSS-vcSS) with 15 to 20 cm sized mud clasts above the erosional base.

Fig. 11: Examples of levée deposits (FA 4). (A) Thin-bedded sandstone facies (F9). (B) Commonly observed discordant bed contact (dashed line). (C) Sunken ripples and climbing-ripple cross-laminations (CRCLs) implying flow direction to the left or south-west. (D) Structureless fine-grained sandstone bed with abundant carbonaceous material content (darker colour) overlain by cleaner, very-fine grained sandstone with ripple cross-laminations (RCLs). Flame structures on mudstone beds. (E) Abundant wood detritus in thin-bedded siltstone (F10) and thin-bedded sandstone (F9) facies. (F) Structureless sandstone that is internally graded to mudclast-rich towards the bed top are found interstratified with the fine-grained turbidites.

Fig. 12: Examples of facies comprising the basin-floor deposits (FA 5). (A) Tabular turbidite beds (F13) with dewatering pipes. The scale is 15 cm long. (B) Very-coarse grained sandstone with flames capped by thin carbonaceous breccia beds, interpreted to be ‘linked’ debrite. (C) Structureless sandstone with abundant whitish shell fragments. (D) Conglomeratic sandstones with erosional bases (F7) found above finer grained deposit. (E) Graded very-coarse grained sandstone with quartz granules above mud-rich mass-transport deposits. (F) Banded sandstone with carbonaceous fragments and mudclasts.

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Fig. 13: Examples of mass-transport deposits found in the proximity of the base of slope and on the basin floor (FA 6). (A) Sand-rich chaotic deposit with patches of coarse-grain segregation and large mudclasts. The scale is 15 cm long. (B) Mud-rich slump deposit containing deformed block of thin-bedded turbidites, overlain by similar but undeformed facies. (C) and (D) Graded turbidite beds contain rip-up mudclasts from the underlying mass transport deposit. The pencil is 15 cm long. (E) Structureless sandstone often found above mass transport deposits with erosional contact.

Fig. 14: Summary of depositional elements. (A) Clinoform E amplitude map. (B) Clinoform E variance map with light source directed from the north-west to highlight the clinoform morphology. (C) to (F) Seismic cross-sections of the depositional elements and their representative sedimentary logs.

Fig. 15: A very tentative seismic interpretation of the lacustrine depositional architecture based on careful and high-resolution data from outcrop measured sections (Fig. 6) and well logs, interpolated into the seismic lines. The datum is Sarmatian–Meotian age boundary marked by the orange arrows near the bottom of the measured sections: (A) depositional-strike view; (B) depositional-dip view. Each environment is composed of the colour coded facies from Table 2. See the locations of the cross-sections and wells in Fig. 1.
REFERENCES


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Talling, P. J., Paull, C. K., and Piper, D. J. (2013) How are subaqueous sediment density flows triggered, what is their internal structure and how does it evolve? Direct observations from monitoring of active flows. Earth Sci. Rev., 125, 244-287.


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Table 1: Summary of dinoflag geometry and dimensions of associated depositional elements from seismic data (Fugger et al., 2016)

<table>
<thead>
<tr>
<th>Features</th>
<th>Dimension</th>
</tr>
</thead>
<tbody>
<tr>
<td>Topset (shell) width</td>
<td>2-50 km</td>
</tr>
<tr>
<td>Clinoform height</td>
<td>280-470 m</td>
</tr>
<tr>
<td>Fore reef length</td>
<td>5-13 km</td>
</tr>
<tr>
<td>Slope gradient</td>
<td>1.5°-5.5°, average 2°-4°</td>
</tr>
<tr>
<td>Sublacustrine canyons</td>
<td>0.8-4 km wide, 50-100 m deep</td>
</tr>
<tr>
<td>Sublacustrine channels</td>
<td>100-200 m wide, 25-50 m deep</td>
</tr>
<tr>
<td>Sublacustrine lobes</td>
<td>2-8 km wide, 5-10 km long</td>
</tr>
</tbody>
</table>

Comment (BFL): Sublacustrinelobe dimension was added at Dr. Stern's suggestion.
### Table 2: Summary of Lithofacies

<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Type log and grain size</th>
<th>Sedimentary Structures</th>
<th>Contact</th>
<th>Bed thickness</th>
<th>Secondary features</th>
<th>Environments and processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>F1: Cross-stratified sandstone</td>
<td>Gravelly cross-stratification (SCS)</td>
<td>Sheet lamination</td>
<td>Gradational Contact</td>
<td>18-20 cm</td>
<td>Fine shell fragments</td>
<td>Storm wave reworking on high shoreline sandstones</td>
</tr>
<tr>
<td>F2: Fluvial channel sandstone</td>
<td>Gravelly cross-stratification (SCS)</td>
<td>Sheet lamination</td>
<td>Gradational Contact</td>
<td>18-20 cm</td>
<td>Fine shell fragments</td>
<td>Storm wave reworking on high shoreline sandstones</td>
</tr>
<tr>
<td>F3: Thick cross-beded sandstone</td>
<td>No grading or upward fining of coarse sandstone, planar lamination or discontinuous shell fragments</td>
<td>Sheet lamination</td>
<td>Gradational Contact</td>
<td>18-20 cm</td>
<td>Fine shell fragments</td>
<td>Storm wave reworking on high shoreline sandstones</td>
</tr>
<tr>
<td>F4: Cross-beded sandstone with silt</td>
<td>Grade lamination</td>
<td>Sheet lamination</td>
<td>Gradational Contact</td>
<td>18-20 cm</td>
<td>Fine shell fragments</td>
<td>Storm wave reworking on high shoreline sandstones</td>
</tr>
</tbody>
</table>

Comment (R2F): Typo in the table were corrected.
<table>
<thead>
<tr>
<th>Lithofacies</th>
<th>Type top and grain size</th>
<th>Sedimentary structures</th>
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<th>Secondary features</th>
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<th>Environments and processes</th>
</tr>
</thead>
<tbody>
<tr>
<td>PS: Graded sandstone</td>
<td>Normal grading, wave to normal grading; no structures, often amalgamated</td>
<td>Normal grading, parallel or step plane lamination, ripple and bedding, ripple cross-lamination and RCLs</td>
<td>Erosional bases, sharp or transgressive tops</td>
<td>20-100 cm, thickening up to 10-20 m</td>
<td>Opaque or foliated, parallel and massive, drop-up clasts above the base, grading to parallel laminations and RCLs</td>
<td>Channel-fill deposits, thickening</td>
<td>Channel-fill deposits (T), channel bank deposits (B) and braided distributary deposits (R) and channel bank deposits (B) are deposited from high density turbidity currents (S)</td>
</tr>
<tr>
<td>BS: Graded sandstone</td>
<td>Normal grading, parallel or step plane lamination, ripple and bedding, ripple cross-lamination and RCLs</td>
<td>Erosional bases, sharp or transgressive tops</td>
<td>20-100 cm, erosional surface bound beds, 5-11 cm to 20 cm, drop-up clasts above the base, parallel laminations and RCLs</td>
<td>Mud clasts of silt and mud, organic material fragments and lenses, shell fragments and bioclasts</td>
<td>Channel-fill deposits, thickening</td>
<td>Clayey channel and sheet or deltaic deposits (B) and braided distributary deposits (R) and channel bank deposits (B) are deposited from high density turbidity currents (S)</td>
<td></td>
</tr>
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<td>PS: Graded sandstone</td>
<td>Normal grading, parallel or step plane lamination, rippled and bedding</td>
<td>Erosional bases, sharp or transgressive tops</td>
<td>20-100 cm, erosional surface bound beds, 5-11 cm to 20 cm, drop-up clasts above the base, parallel laminations and RCLs</td>
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Delta lobes are about 10-30 m thick, 2-3 km wide and 3-4 km long. The fluvial-deltaic system constitutes the clinoform basal deposits. Pond-rich mudstones are interstratified with the shoreline deposits suggesting transgression of unknown magnitude.

Fluvial channels:
Fluvial channels are 4-25 m deep and 100-200 m wide in the seismic images. Most fluvial channel might be under seismic resolution and/or small-filled. Multiple channels observed along the incised shelf edge have connection with the slope channels.

Slope channels:
Slope channels are 25-50 m deep, 100-200 m wide, found within 0.8-4 km wide, 50-100 m deep canyons. Levede channel are 50-80 m deep and 0.6-1 km wide. The deposits are 15-20 m thick and reflect oversteepening deposits.

Basin-floor lobes:
Turbidite lobes are up to 10 m thick together they form ca. 30 m thick lobe complex(es) that are 2-8 km wide and 5-10 km long in the seismic data. Turbidite lobes commonly over 15-30 m thick mass transport deposits.