1. Introduction

Restricted sedimentary basins constitute complex depositional environments. They form semi-isolated basins when they display limited water connections to the marine realm, or isolated basins when all connections are interrupted. The sedimentary infill of (semi-)isolated basins is mostly controlled by the relative importance of accommodation space, water supply and sediment supply (Bohacs et al., 2003).

Restricted sedimentary basins generally record limited accommodation space compared to the open ocean (Carroll and Bohacs, 1999). An important control on the limited accommodation space is the position and relative height of the spill point of the basins (Leever et al., 2010; Yanina, 2014; Pongnern et al., 2016), which may at the same time affect their connectivity to the marine realm (Leever et al., 2011; Ter Borgh et al., 2014). Restricted connectivity with the open ocean results in the prevalence of brackish to freshwater depositional environments and is usually accompanied by the development of faunas endemic to the basin (e.g., Jones and Simmons, 1996; Rögl, 1998; Wesselingh et al., 2006). Limited water interchanges between (semi-)isolated basins and the open ocean lead moreover to low energy depositional environments. Therefore, (semi-)isolated basins generally lack tidal influence and often show reduced wave interference (Medvedev et al., 2016). The isolated nature of restricted basins also enhances their...
sensitivity to external forcing factors, such as changes in precipitation, runoff and (cyclic) astronomical controls (e.g., Müller et al., 2001; Abels et al., 2009; Litt and Anselmetti, 2014; Wagner et al., 2014; Constantinescu et al., 2015). If properly understood, (semi-)isolated sedimentary basins may provide valuable environmental, climatic and biological archives. Understanding depositional processes operating in these basins may provide spatial and temporal insights into the distribution of sedimentary facies and fauna, and could elucidate potential climate forcing.

Although the understanding of (semi-)isolated basins has significantly increased in the past decades through various deep-drilling projects (e.g., Ross, 1978; Francke et al., 2016), depositional processes occurring in restricted basins and their impacts on sedimentary facies are not still well known. The sedimentary infill of restricted basins is likely controlled by forcing factors that are different to the open ocean. When (semi-)isolated basins are infilled by prograding deltas, the resulting deltaic architectures are likely influenced by both external and internal forcing factors specific to restricted depositional settings. The impact of these forcing factors may however be difficult to identify in deltaic sedimentary records. In particular, the relative importance of alluvial forcing factors driving accommodation space, water supply and sediment supply acting of the deltaic progradation compared to autogenic deltaic avulsion processes remain poorly studied in restricted basins.

As a result, deltaic facies models for (semi-)isolated basins are less well developed than those for open marine environments (Andrews et al., 2016; Nutz et al., 2017). This is remarkable as some of the present-day (semi-)isolated basins are infilled by major deltas. The restricted Black Sea and isolated Caspian Sea accommodate for instance the Danube and Volga deltas, the two longest rivers in Europe (Overeem et al., 2003; Giosan et al., 2005). Unfortunately, continuous exposures recording long periods of deltaic deposition in (semi-)isolated basins are relatively rare.

In this study, we investigate the mid-Pliocene sedimentary architecture of a river-dominated delta entering the semi-isolated Dacian Basin in Romania. The Dacian Basin formed at that time a brackish embayment of the ancient Black Sea. Deltaic and alluvial sediments prograded on the northern margin of this restricted basin. We investigated an 835 m thick, continuous sedimentary section of fossil-rich sediments, cropping out along the Slănicul de Buzău River, in Romania (Andreeescu et al., 2011; Van Baak et al., 2015). In this article we combine detailed analyses of sedimentary facies, with a description of the accompanying biofacies. The quality and continuity of the exposure allows for establishing a detailed sedimentological and sequence-stratigraphic framework, which permits investigating the drivers of the internal deltaic architecture. Moreover, because of available magnetostratigraphic time constraints (Van Baak et al., 2015), the impact of autogenic versus alluvial forcing factors on the deltaic sedimentary architecture can be discussed. Facies models developed in this paper may form analogues for more poorly exposed or subsurface deltaic successions in (semi-)isolated basin elsewhere.

2. Geological background

The Paratethys Sea formed one of the largest intercontinental seas that ever existed (Rögl, 1998; Popov et al., 2006). During the Oligocene, convergence between Africa and Eurasia generated a topographic barrier, which isolated the Paratethys Sea from the Tethys Ocean (Allen and Armstrong, 2008; Schmid et al., 2008). Oligocene and Miocene tectonic activity produced numerous mountain belts (Vincent et al., 2007, 2016; Schmid et al., 2008), which further fragmented the Paratethys Sea into several semi-isolated basins (Popov et al., 2006). From west to east, the four major sub-basins were the Pannonian, Dacian, Euxinian (Black Sea) and Caspian basins (Fig. 1a).

This paper focuses on the semi-isolated Dacian Basin, a former embayment of the Black Sea. This basin represents the late Miocene to present-day foreland basin in the eastern and southern parts of the Carpathians (Matenco and Bertotti, 2000; Cloetingh et al., 2004; Panaiotu et al., 2007; Jipa, 2015). Following mountain belt uplift, a deep depocentre was formed in front of the Southeast Carpathians (Bertotti et al., 2003; Târâpoacă et al., 2003). The depression was progressively filled with the erosional products of the uplifted mountains (Jipa, 1997; Sanders et al., 1999; Târâpoacă et al., 2003; Panaiotu et al., 2007) (Fig. 1b). Open water deposits with brackish water faunas, which accumulated in the basin during the late Miocene-early Pliocene (Pontian regional stage - Stoica et al., 2013), were gradually replaced by alluvial deposits with freshwater faunas towards the late Pliocene (Romanian regional stage - Van Baak et al., 2015). This transition of depositional environments occurred during the intermediate Dacian regional stage, which lasted from 4.8 to 4.2 Ma (Vasileiv et al., 2005; Van Baak et al., 2015) (Fig. 2a). At that time, a major delta prograded east of the Carpathians towards the northeastern margin of the Dacian Basin (Jipa, 1997; Jipa and Olariu, 2009; Fonngern et al., 2016; Matoshko et al., 2016), whereas sediments shed from the Southern Carpathians mainly accumulated in the western Dacian Basin (Jipa and Olariu, 2009; Jipa et al., 2011; Ter Borgh et al., 2014; Fonngern et al., 2017). Palaeogeographic and provenance data indicate that the basin was eventually entirely filled during the late Pliocene to early Pleistocene, when sediments started to overspill into the Black Sea (De Leeuw et al., 2018; Olariu et al., 2018).

Post-collisional shortening affected the Carpathian Foredeep during the Quaternary (Necea et al., 2005; Leever et al., 2006; Maynard et al., 2012). Faulting and large-scale folding of the foreland infill occurred in the Buzău area of Romania. As a result of this Quaternary foreland inversion, long and continuous exposures of the late Miocene to early Pleistocene foreland infill can be found at the surface.

3. The Slănicul de Buzău section

The investigated section crops out along the Slănicul de Buzău River, between the villages Cernăuți and Minzălești (Fig. 2b). The river cuts through a 6.4 km thick stratigraphic succession, recording folded foreland late Miocene to Pleistocene deposits (Snel et al., 2006; Andreeescu et al., 2011; Van Baak et al., 2015).

Our study focuses on the part of the section corresponding to the mid-Pliocene Dacian regional stage. The existing age model for this part of the valley (Van Baak et al., 2015) shows that this section contains two normal magnetozones, interpreted as paleomagnetic chron C3n.2n (Nunivak), aged 4.631–4.493 Ma and C3n.1n (Cochiti), aged 4.300–4.187 Ma (absolute ages from Gradstein et al., 2012). Including the under- and overlying sediments, the entire studied section was therefore deposited between 4.8 Ma and 4.11 Ma. These age constraints are in line with studies of the Dacian regional stage at other locations in the Dacian Basin (e.g., Vasileiv et al., 2004).

4. Methods

4.1. Sedimentological data collecting

The Dacian segment of the Slănicul de Buzău section starts under the bridge North of the village Niculești (N45°26′26.87″, E26°44′37.97″) and continues for 2 km northwards in the river bed (N45°27′14.81″, E26°44′41.29″) (Fig. 2c). The section exposes an 835 m thick stratigraphic succession. Thicknesses were measured in the field and later checked by GPS measurements.

This part of the section had previously been logged at a m-scale (Van Baak et al., 2015). This more generalized log unfortunately missed a 40 m interval in the middle part of the section, which is now included in our work (Fig. 3a). To support our sedimentological analyses, the section was analysed in greater detail and several key intervals were described at a cm-scale. Variations in lithology, grain size and sedimentary structures were recorded in the field. Particular attention was
paid to sedimentary structures, such as graded bedding, laminations, cross-stratification and ichnofossils. Samples were collected for sedimentological and petrographic optical microscopic descriptions. Thin-sections with a thickness of 30 μm were made perpendicular to the sedimentary structures for petrographic descriptions.

Detailed sedimentological observations allowed for several typical lithofacies to be established, based on sediment grain size, sedimentary structures, ichnofossils and faunal composition. Lithofacies repeatedly occurring together along the section were then grouped into facies associations, each of which related to a distinct depositional environment and an estimated water depth. A depth ranking scale was subsequently constructed by attributing a number from 0 to 9 to each facies association, with 0 being the deepest and 9 the shallowest depositional environment (Table 1). This ranking scale permitted the reconstruction of a relative water-level curve and the identification of parasequences, including superimposed lower- and higher-order sequences (Fig. 3b).

4.2. Palaeocurrent determination

Along the sedimentary succession, 41 palaeocurrent directions were measured on 3D cross-beds. As this section was affected by post-depositional folding, palaeocurrent directions needed correction to remove any tectonically-induced rotations (Supplementary material 1). Corrections were realized with the help of the available palaeomagnetic dataset. We proceeded with a first step of deplunging the fold axis, followed by a second step of correcting for the true vertical axis rotation.
To correct for the plunging fold axis, all obtained palaeomagnetic directions, bedding planes and their poles were plotted in a stereographic projection using Stereonet 9 software (Allmendinger et al., 2011; Cardozo and Allmendinger, 2013). Bedding planes and palaeomagnetic directions were subsequently rotated 19° around a rotation axis with a 121° azimuth and 0° plunge to restore the fold axis to horizontal. The corrected palaeomagnetic directions were then entered as pre-tilt directions in the statistics portal of paleomagnetism.org (Koymans et al., 2016), together with their associated plunge-corrected bedding planes.

For the second step of unfolding, regular tilt-correction on the basis of these bedding planes was applied to place palaeomagnetic directions in their correct tectonic reference frame. Subsequent regular statistical analysis revealed a plunge-corrected anticline with a mean direction of 171°. This implies a 9° counter clockwise horizontal plane rotation of 14°. These results are in line with the previously determined 9° rotation (Slănic site of Dupont-Nivet et al., 2005; Vasiliev et al., 2009; Van Baak et al., 2015). They illustrate tectonic rotation in the Buzău area, due to the deformation of the Carpathian Bend zone.

Once the palaeocurrent directions were corrected, their statistical distribution was calculated for 12 sectors of 30° and plotted on four rose diagrams with a maximum representability of 50%. One of the rose diagrams represents the overall flow direction along the section. Palaeocurrent directions were additionally plotted on three rose diagrams according to their respective depositional environments.

4.3. Cyclostratigraphical analysis

An age model was constructed using the existing magnetostratigraphic timescale of the studied section. For each magnetozone, the sedimentation rate was calculated in order to evaluate variations of sediment input into the basin through time.

A cyclostratigraphical analysis was performed on the frequencies of the parasequences, low- and high-order sequences, to evaluate potential astronomical forcing on sedimentation. Blackman-Tuckey spectral analyses were realized using standard settings on an equally-spaced data series in the Analyseries 2.0.4b program. Analyses were performed at 90% confidence levels. Bandpass filters were generated with a bandpass-width defined arbitrary between 50 and 138 m, 64 and 111 m, 25 and 37 m and 13.6 and 25 m. Filters were then plotted against the facies rank data and the astronomical target curves.

4.4. Faunal analyses

Working in Paratethyan basins may introduce a certain ambiguity between palaeontological and sedimentological terminology (Matoshko et al., 2016). These basins registered episodic periods of connectivity and disconnectivity with the open ocean and therefore display lowered salinity environments, where endemic faunas developed though time (Marinescu et al., 1978; Stoica et al., 2013). Here, the terms “brackish water” and “freshwater” are used to specify water salinity on the basis of palaeontological indicators. The term “open water” is used in a sedimentological context in order to describe offshore to shoreface depositional environments.

The Slănicul de Buzău section display very rich assemblages of endemic mollusc and ostracod faunas. For this study, both groups were analysed in order to corroborate environmental reconstructions based on sedimentology. Mollusc assemblages were studied from 58 sediment samples. Samples of 1000–2000 g were taken throughout the section (Fig. 3a, SBD16-nf). Sample preparation for molluscs was performed at the Natural History Museum in Vienna. Samples were cleaned using pneumatic micro-chisels. They were then washed through sieves of 1 mm. The general preservation of the shells was moderate to poor. Shells were finely cracked due to secondary gypsum mineralisation and carbonate crystal growth. The taxonomic identifications follow Wenz (1942) and Marinescu and Papaianopol (1995). Taxonomic revision incorporates results by Neveskaya et al. (1997, 2001, 2013) and Neubauer et al. (2014) (Supplementary material 2).

Ostracod assemblages were studied from 35 sediment samples of 500–1000 g taken throughout the section (Fig. 3a, SBD16-nf). Sample preparation for ostracods was carried out at the Department of Palaeontology at the University of Bucharest. Samples were dried to remove interstitial water from the sediments. Dry samples were subsequently boiled for 30–60 min with a sodium carbonate solution for better disaggregation. Samples were washed through several sieves of 63 to 500 μm. The residues were studied under a ZEISS-Stemi SV11
microscope. Pictures of microfaunas were taken with a NIKON digital camera. The general preservation of the ostracods was moderate to poor. Ostracods were often fragmented due to strong diagenetic processes. The taxonomic identifications follow Hanganu (1976, 1985), Hanganu and Papaianopol (1977), Stancheva (1990) and Olteanu (1995) (Supplementary material 2).

5. Results

5.1. Sedimentary facies associations

The 835 m thick succession displays a generally regressive trend, superposed by a rhythmic alternation between more distal clays and more proximal sands. Our field observations and subsequent microscope descriptions were compared to well documented sedimentological classifications (Postma, 1990; Miall, 2006). We distinguished thirteen lithofacies, formed by distinct sedimentary processes (Table 2, Figs. 4–7). Lithofacies were grouped into eight depositional facies, representing five main facies associations, each of which being related to a distinctive depositional environment.

5.1.1. Prodelta facies association

5.1.1.1. Description. The first facies association is generally 1 to 5 m thick. It consists of three types of dark-bluish-gray (GLEY2-4/5B) to bluish-gray (GLEY2-5/5B) mudstone that occur successively. There are massive (Fm), laminated (Fl) and lenticular (Fs) mudstones. Massive mudstones (Fm) occur at the base of the prodelta facies association strata. They are generally 0.5 to 1 m thick, but sometimes are absent from this facies association. They display a dark-bluish-gray color (Fig. 4b). They may contain well-preserved in situ brackish water molluscs, such as Euxinicardium olivetum, Pontalmyra tohanensis or Chartoconcha rumana (Supplementary material 3, Fig. 8). This facies is progressively replaced by 1 to 3 m thick, laminated gray mudstones (Fl). These mudstones have mm-scale horizontal laminations of silt (Fig. 4c). They may also contain cm-scale horizontal laminations of silt with mm-scale fragments of terrestrial organic material. Upwards, the muddy succession may contain 0.5 to 1 m thick, gray mudstones with lenticular bedding (Fs). Lenticular bedding consists of 1 to 5 cm thick isolated lenses made of silt to very fine sand showing trough cross-stratification (Fig. 4d). The sandy layers are occasionally affected by cm-scale convolute bedding. Throughout this facies association, laminations and lenses become thicker, more frequent, and composed of coarser sediments towards the top. Palaeocurrent directions were measured on 3D trough-cross stratifications present in lenticular-bedding. Measurements in these deposits demonstrate a mean direction of 225° (n = 17; Fig. 7a). They display a wide range of current directions from 180° to 270°. In addition, deposits occasionally show intercalations of cm-thick beds of gray (GLEY1-6/N) fine to medium grained sandstones (Sfr). Their bases form a wavy, sharp surface, highly perturbed by vertical burrows 3 to 5 cm wide and 5 to 15 cm deep (Fig. 5e). These sandstones are moderately-sorted and are composed of subangular quartz grains with high sphericity. They contain many abraded or broken, reworked brackish and freshwater cardiids, unionids, dreissenids or viviparids.

5.1.1.2. Interpretation. This facies association is interpreted as a prodelta environment because of the distal depositional setting and the evidence...
of distal fluvial input. The massive mudstones (Fm) indicate deposition from suspension in open water. The progressive transition to mm-scale silty laminations (Fl) is related to large volumes of fine-grained sediments being transported into the basin over long distances during intermittent sandstone deposition. As in previous studies (Starek et al., 2010; Hampson et al., 2011), these sandstones are interpreted as storm deposits. The coarse-grained, unidirectional current ripples, and possible erosive bases highlight the development of several delta-lobes, feeding a wide prodelta region.

### Table 1
Facies depth ranks used to create the relative water-level curve of the section.

<table>
<thead>
<tr>
<th>Index</th>
<th>Depositional environment</th>
<th>Sedimentary structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>Open water</td>
<td>Mudstones, structureless, in situ brackish water faunas</td>
</tr>
<tr>
<td>1</td>
<td>Distal prodelta</td>
<td>Mudstones, mm-scale planar laminations of silts</td>
</tr>
<tr>
<td>2</td>
<td>Proximal prodelta</td>
<td>Mudstones, cm-scale lenticular bedding of silts and very fine sandstones</td>
</tr>
<tr>
<td>3</td>
<td>Distal delta-front</td>
<td>Mudstones, cm- to dm-scale cross-beding of very fine sandstones</td>
</tr>
<tr>
<td>4</td>
<td>Proximal delta-front</td>
<td>Fine grained sandstones, dm-scale cross-beding of very fine sandstones</td>
</tr>
<tr>
<td>5</td>
<td>Distributary mouth bar/Interdistributary bay</td>
<td>Medium grained sandstones, dm to m-scale cross-beding, erosive base, reworked freshwater faunas</td>
</tr>
<tr>
<td>6</td>
<td>Hardground</td>
<td>Medium grained sandstones, erosive base, glauconite, reworked brackish and freshwater faunas</td>
</tr>
<tr>
<td>7</td>
<td>Channel fill</td>
<td>Fine grained sandstones, cm-scale cross-beding, organic material fragments, ichnofossils</td>
</tr>
<tr>
<td>8</td>
<td>Coastal plain</td>
<td>Clays, structureless, organic material fragments, ichnofossils, in situ freshwater faunas</td>
</tr>
<tr>
<td>9</td>
<td>Peat</td>
<td>Clays, structureless, cm-scale coal layers, roots</td>
</tr>
</tbody>
</table>

### Table 2
Description of the main lithofacies characteristics and associated sedimentary processes identified along the section.

<table>
<thead>
<tr>
<th>Code</th>
<th>Grain size</th>
<th>Sedimentary structures</th>
<th>Inclusions</th>
<th>Processes</th>
<th>Fig.</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sfr</td>
<td>Fine to medium grained sandstone</td>
<td>Hints of low-angle cross-stratification, erosive base, induration, reddish oxidation</td>
<td>Reworked brackish and freshwater shells, glauconite</td>
<td>Very high energy flow, sediment reworking, post-diagenetic induration and oxidation</td>
<td>5f</td>
</tr>
<tr>
<td>Sfg</td>
<td>Fine to medium grained sandstone</td>
<td>Trough cross-stratification, preserved cross-set thickness of 10–50 cm, possible erosive base, unidirectional</td>
<td>Reworked brackish and freshwater shells, ichnofossils</td>
<td>Very high energy flow, sediment reworking</td>
<td>5e</td>
</tr>
<tr>
<td>Sr</td>
<td>Very fine to medium grained sandstone</td>
<td>Orginal material fragments along foresets</td>
<td>Organic material fragments along foresets</td>
<td>Moderate energy flow, migration of sinuous crested dunes</td>
<td>5d</td>
</tr>
<tr>
<td>Ss</td>
<td>Fine to medium grained sandstone</td>
<td>Sigmoidal cross-stratification, preserved cross-set thickness of 20–100 cm, possible erosive base, possible inverse grading, unidirectional</td>
<td>Organic material fragments along foresets</td>
<td>Moderate to high energy flow, migration of straight crested dunes</td>
<td>5c</td>
</tr>
<tr>
<td>Sr</td>
<td>Very fine to medium grained sandstone</td>
<td>Climbing ripple, preserved cross-set thickness of 10–50 cm, unidirectional</td>
<td>Organic material fragments along foresets</td>
<td>Moderate energy flow, migration of curved crested ripples, abundant suspended material</td>
<td>5b</td>
</tr>
<tr>
<td>Sc</td>
<td>Fine to medium grained sandstone</td>
<td>Asymmetrical current ripple, amplitude of 3–5 cm and wavelength of 7–10 cm, unidirectional</td>
<td>Organic material fragments along foresets</td>
<td>Moderate energy flow, migration of ripples, abundant suspended material</td>
<td>5a</td>
</tr>
<tr>
<td>Sh</td>
<td>Fine to medium grained sandstone</td>
<td>Horizontal lamination, possible normal and inverse grading, possible erosive base</td>
<td>Possible organic material fragments along lamina tions</td>
<td>Low to high energy flow, plane-bed flow</td>
<td>5g</td>
</tr>
<tr>
<td>Sl</td>
<td>Fine to medium grained sandstone</td>
<td>Low-angle cross-stratification, preserved cross-set thickness of 30–200 cm, possible inverse grading, possible erosive base, unidirectional</td>
<td>Organic material fragments along foresets</td>
<td>High energy flow, migration of low relief dunes</td>
<td>4f</td>
</tr>
<tr>
<td>Sm</td>
<td>Fine to medium grained sandstone</td>
<td>Structureless, inverse grading, dm-scale troughs infilled with organic material fragments</td>
<td>Reworked and in situ freshwater shells, ichnofossils</td>
<td>Low energy flow</td>
<td>4e</td>
</tr>
<tr>
<td>Fs</td>
<td>Mudstone</td>
<td>Lenticular bedding made of trough cross-stratified silt and very-fine sand, preserved cross-set thickness of 1–5 cm</td>
<td>Organic material fragments along foresets</td>
<td>Fluctuation between very low and moderate energy flow, deposition from suspension and distal outflows</td>
<td>4d</td>
</tr>
<tr>
<td>Fl</td>
<td>Mudstone</td>
<td>Horizontal lamination</td>
<td>Silts and organic material fragments along lamina tions</td>
<td>Fluctuation between very low and moderate energy flow, deposition from suspension and distal outflows</td>
<td>4c</td>
</tr>
<tr>
<td>Fm</td>
<td>Mudstone</td>
<td>Structureless, in situ brackish water shells</td>
<td>Organic material fragments along foresets</td>
<td>Very low energy, deposition from suspension and distal outflows</td>
<td>4b</td>
</tr>
<tr>
<td>C</td>
<td>Clay</td>
<td>Structureless, possible coal layers, possible induration</td>
<td>In situ freshwater shells, ichnofossils</td>
<td>Very low energy, deposition from suspension, possible post-diagenetic induration</td>
<td>4a</td>
</tr>
</tbody>
</table>
composed of coarser sediments towards the top of the facies association. As in the previous prodelta facies association, deposits are occasionally interrupted by the same cm-scale beds of gray (GLEY1-6/N) fine to medium grained sandstones (Sfg). They are structureless and composed of well-sorted sediments with highly-spherical and subangular quartz grains. Their bases display the same wavy, sharp, highly bioturbated surface (Fig. 5e). They also contain many reworked, abraded and broken brackish and freshwater molluscs.

5.1.2.2. Interpretation. This facies association is interpreted as representing a distal delta-front environment more frequently influenced by fluvial input. Mudstones were deposited from suspension in open waters. Episodic intercalations of sandstones are related to increases of sand input coming from the distal margin of distributary channels. Once these sands reach the distal delta lobe, they record a relative deceleration and form cm- to dm-scale, migrating submarine dunes comprising small-scale cross-bedding. The lenticular bedding (Fs) is related to wave action and/or winnowing (De Raaf et al., 1977). The climbing ripples may be related to rapid sedimentation rates and non-uniform flows, due to a loss of flow confinement or a decrease in slope gradient (Jobe et al., 2012). The cm-scale, organic-rich, silty laminations could be related to hyperpycnal flows, occurring during episodic larger river discharge events (Mulder et al., 2003; Bhattacharya and MacEachern, 2009; Lamb and Mohrig, 2009). The convolute bedding could have been created when the sandstones were rapidly deposited on the underlying water-saturated mudstones, causing an expulsion of the fluids contained in the mud (Oliveira et al., 2009). The shell-rich sandy beds (Sfg) sporadically intercalated in this facies association suggest storm deposits (Starek et al., 2010; Hampson et al., 2011). This facies association, showing coarsening up, displays an increase in the energy of the depositional process and illustrates

Fig. 4. Photos of the different lithofacies identified along the section, facies codes in brackets (Part I). (a) Organic-rich clays C. (b) Massive mudstone Fm. (c) Laminated mudstone Fl. (d) Lenticular mudstone Fs. (e) Massive sandstone Sm. (f) Low-angle cross-stratified sandstone Sl. (g) Horizontally laminated sandstone Sh.
a depositional environment closer to the distributary system. In line with previous studies on similar facies associations (Fielding, 2010; Hampson et al., 2011), the depositional setting was interpreted as a distal delta-front environment.

5.1.3. Proximal delta-front association

5.1.3.1. Description. The third facies association is marked by a lack of mudstones. It is composed of 0.5 to 2 m thick, grayish-brown (2.5Y-5/2), fine grained sandstones. The sands are moderately sorted, subangular and have low sphericity. They form dm-scale layers and comprise the same three types of cross-beddings, but at a dm-scale (Sr, Ss, St). Sandstones contain 10 to 50 cm thick sets of climbing ripples (Fig. 5b), 20 to 100 cm thick sigmoidal cross-stratification (Fig. 5c) and 10 to 50 cm thick trough cross-stratification (Fig. 5d). Sedimentary structures are draped by mm-scale laminae of fragments of terrestrial organic material. Palaeocurrent directions in this facies association were measured on trough cross-stratification, sigmoidal cross-stratification and climbing ripples. They have a mean direction of 180° (n = 19; Fig. 7b), with a range from 90° to 330°.

5.1.3.2. Interpretation. This facies association is interpreted as deposited in a proximal delta-front environment, in agreement with comparable studies of similar facies associations (Fielding, 2010; Hampson et al., 2011; Forzoni et al., 2015). Sediments were deposited under higher energy conditions compared to the distal delta-front deposits and are therefore mostly sand-dominated. The thicker sandy beds contain larger-scale sedimentary structures, formed by migration of larger-scale dunes. Previous authors deduced that similar sediments were transported towards the basin by subaqueous terminal

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**Fig. 5.** Photos of the different lithofacies identified along the section, facies codes in brackets (Part II). (a) Current rippled sandstone Sc. (b) Climbing rippled sandstone Sr. (c) Sigmoidal cross-stratified sandstone Ss. (d) Trough cross-stratified sandstone St. (e) Grayish shelly sandstone Sfg. (f) Reddish shell-rich sandstone Sfr.
5.1.4. Delta-top facies association

The fourth facies association groups four different facies, deposited under specific sedimentary processes, in the same depositional environment.

5.1.4.1. Interdistributary bay facies

5.1.4.1.1. Description. Sediments deposited in the first facies consist of 1 to 5 m thick sandstones, directly overlying prodelta facies. The greenish-gray (GLEY1-5/5GY), moderately sorted, sandstones (Sm) have low sphericity and are subangular. Sandstones form m-thick continuous layers with a diffuse base (Fig. 4e). The layers coarsen upwards from very fine to medium grain-sizes towards the top of the sandy beds. Sandstones are massive and structureless. They contain a few, dm-scale troughs, which are infilled with mm-scale fragments of terrestrial organic material. The sandstones include some well-preserved in situ freshwater mollusks, such as unionids or viviparids (Supplementary material 3, Fig. 8). They also contain many vertical and horizontal burrows 0.5 to 1 cm wide and 5 to 10 cm deep, made by Crunizana ichnofossils, such as Cylindrichnus.

5.1.4.1.2. Interpretation. This facies was deposited from suspension under low-energy and low salinity conditions. Environmental conditions are corroborated by the presence of freshwater mollusks, burrows, and terrestrial organic material. In line with previous studies (Elliott, 1974; Overeem et al., 2003), this facies is interpreted to be deposited in an interdistributary bay environment, between distributary channels. Sand-laden currents entered and progressive encroached the interdistributary bay, producing a coarsening upwards succession. Finer-scale sedimentary structures and thin intervening bay mudstones and sandstones were probably erased by intensive bioturbation.

5.1.4.2. Distributary mouth bar facies

5.1.4.2.1. Description. The second facies forms 2 to 5 m thick sandstone beds. The grayish-brown (2.5Y-5/2), fine to medium grained sandstones are moderately-sorted and comprise low-sphericity and subangular quartz grains. They form m-scale beds with weak inverse grading from the base to the center of the beds. Sand beds may also occasionally record weak normal grading from the middle to the top of the beds. Sandstones contain four types of dm- to m-scale cross-bedding (Sr, Ss, Sl, Sh). They display 10 to 50 cm thick climbing ripples at the bases and/or at the tops of the sandy beds (Fig. 5b), 50 to 200 cm thick sigmoidal cross-stratification (Fig. 5c), 200 to 20 cm thick low-angle cross-stratification (Fig. 4f) and mm-scale horizontal laminations (Fig. 4g). Cross-bedding foresets are draped by mm-scale laminae of fragments of terrestrial organic material (Fig. 6a, b). Some cm-scale clay pebbles are sometimes found at the base of this facies.

5.1.4.2.2. Interpretation. This facies is interpreted to be formed in distributary mouth bars, under high-energy depositional processes. Climbing ripples at the bases and tops of these beds were formed by migration of small-scale current ripples, whereas larger-scale cross-bedding in the middle parts were formed by migration of large-scale dunes. Clay pebbles at the base of the sand beds were likely formed by erosion of the underlying muddy substratum. Scouring occurred due to relatively high-energy and concentrated density flows (Mulder and Alexander, 2001). In comparison to studies based on similar facies...
(Allen, 1983; Olariu and Bhattacharya, 2006; Forzoni et al., 2015), we interpreted these deposits as being formed in shallow channelized channels, by lateral and longitudinal accretion of distributary mouth bars.

5.1.4.3. Channel fill facies

5.1.4.3.1. Description. The third facies displays 2 to 3 m thick sandstones forming several dm-thick layers. The grayish-brown (2.5Y-5/2), fine grained sandstones are moderately sorted, have low-sphericity and are subangular. Sandstones show a haphazard succession of various cm-scale cross-bedding types (Sc, Sh, Sr, Sl). They contain asymmetrical current ripples with an amplitude of 3 to 5 cm and a wavelength of 7 to 10 cm (Fig. 5a), mm-scale horizontal laminations (Fig. 4g), 10 to 50 cm thick climbing ripples (Fig. 5b) and occasionally 30 to 50 cm thick low-angle cross-stratification (Fig. 4f). Cross-bedding foresets and horizontal laminations are draped by mm-scale laminae of fragments of terrestrial organic material. Palaeocurrent directions were measured on trough cross-stratification, sigmoidal cross-stratification, climbing ripples, low-angle cross-stratification and asymmetrical current ripples, recorded within the interdistributary bay, distributary mouth bar and channel fill deposits. The measurements display a mean direction of 165° (n = 5; Fig. 7c) and a range from 0° to 210°. They highlight a very wide range of flow directions. Unfortunately, the available amount of data is insufficient to extract any other useful information.

5.1.4.3.2. Interpretation. These sandstones were gradually deposited on top of the distributary mouth-bar deposits. They display various sedimentary structures that are formed by migration of small-scale current ripples or by deposition of fine sediment from suspension. The enrichment in terrestrial organic material draping the sedimentary structures is indicative of waxing and waning of fluvial flow. Sediments were deposited under fluctuating fluvial current velocities. According to work on similar deposits (Elliott, 1974; Fielding, 1986; Bhattacharya, 2006), this facies was interpreted as the infill of a channel, progressively affected by avulsion.

5.1.4.4. Coastal plain facies

5.1.4.4.1. Description. The last facies includes 0.2 to 0.5 m thick structureless clays (C). Clays are very-dark-gray (GLEY1-3/N) and are rich in dispersed mm- to dm-scale fragments of terrestrial organic material. Palaeocurrent directions were measured on trough cross-stratification, sigmoidal cross-stratification, climbing ripples, low-angle cross-stratification and asymmetrical current ripples, recorded within the interdistributary bay, distributary mouth bar and channel fill deposits. The measurements display a mean direction of 165° (n = 5; Fig. 7c) and a range from 0° to 210°. They highlight a very wide range of flow directions. Unfortunately, the available amount of data is insufficient to extract any other useful information.

5.1.4.3.2. Interpretation. These sandstones were gradually deposited on top of the distributary mouth-bar deposits. They display various sedimentary structures that are formed by migration of small-scale current ripples or by deposition of fine sediment from suspension. The enrichment in terrestrial organic material draping the sedimentary structures is indicative of waxing and waning of fluvial flow. Sediments were deposited under fluctuating fluvial current velocities. According to work on similar deposits (Elliott, 1974; Fielding, 1986; Bhattacharya, 2006), this facies was interpreted as the infill of a channel, progressively affected by avulsion.
such as unionids (Rumanunio rumanus) or pachychilids (Tinnyea abchasica) (Supplementary material 3, Fig. 8). The top of the clay beds occasionally shows cm-scale ichnofossils, such as Planolites, forming horizontal burrows 0.5 to 1 cm wide and 1 to 3 cm deep. More rarely, the top of these clay beds displays vertical roots 0.5 to 1 cm wide and 5 to 10 cm deep. The upper 5 to 10 cm of this facies is occasionally indurated.

5.1.4.4.2. Interpretation. The organic-rich mudstones are interpreted as having been deposited from suspension, in low-energy coastal plain mires. The upper parts of this facies, affected by burrows and roots, point to sporadic subaerial exposure of the environments. More prolonged subaerial exposure may have indurated the upper parts of these deposits. In line with previous studies of similar facies (Fielding, 2010; Hampson, 2010; Forzoni et al., 2015), we interpreted the organic-rich layers to be deposited on coastal plains during fluvial flooding.

5.1.5. Hardground

5.1.5.1. Description. The last facies association consists of 0.2 to 0.4 m thick sandstones. Sandstones (Sfr) form dm-thick layers with wavy, sharp, erosive bases (Fig. 5f). Sediments are quartz-rich with highly spherical and subangular grains. They are fine to medium grained and well-sorted sandstones. The weathered surfaces of these sandstones have a noticeable reddish-brown color (2.5YR-4/4), whereas the fresh surface is more grayish (GLEY1-4/N). Sandstones are mostly structureless, but occasionally show cm- to dm-scale low-angle cross-stratification. Microscopic observations realized on thin-sections show
enrichment in subangular glauconite grains (Fig. 6c, d). The sandstones also contain high concentrations of shells, which are often abraded or broken, and composed of a mix between brackish and freshwater molluscs (Supplementary material 3). Sandstones display iron cement, that is post-diagenetically oxidized, distributed throughout the entire sand bed, leading to the formation of indurated layers.

5.1.5.2. Interpretation. These sandstones form hardgrounds, created under high-energy depositional processes. The formation of erosive sand beds, comprising mature sands and many abraded and reworked shells, requires erosion and sediment reworking along the shoreface (Weimer, 1988; Scarponi et al., 2013). The formation of glauconite necessitates slow sedimentation rates down to slight erosion (Cloud, 1955). Subsequent winnowing processes may have diminished the sedimentation rate and caused episodic sediment starvation (Kidwell and Aigner, 1985; Brett, 1995), leading to the formation of condensed layers (Kidwell, 1989; Abbott and Carter, 1994; Brett, 1995; Scarponi et al., 2013). Similar to previous studies (Nummedal and Swift, 1987; Weimer, 1988; Murakoshi and Masuda, 1992; Cattaneo and Steel, 2003; Hurd et al., 2014), we interpret the formation of such oxidized shell-rich hardgrounds to occur during relative water-level rises and represent therefore flooding surfaces. Along our section, these deposits display a red weathering color and are cemented, which is probably the result of post-diagenetic oxidation during subaerial exposure.

5.2. Fauna assemblages

The Slănicul de Buzău section contains very rich mollusc and ostracod assemblages. We identified 25 ostracod species (Figs. 9–12) and

47 mollusc species (Fig. 8). Molluscs comprise about 70% bivalve and about 30% gastropod species.

5.2.1. Biofacies

Based on macrofaunal observations, three major biofacies were identified within the studied deltaic sedimentary succession.

The first biofacies comprised an autochthonous assemblage of several cardiid species, such as *Euxinicardium olivetum*, *Pontalmyra tohanensis* and *Chartoconcha rumana* (Fig. 8). They were preferentially found in clayish prodelta deposits and occasionally in clayish distal delta-front sediments (Supplementary material 3). These species demonstrate brackish water environments (Nevesskaya et al., 2001).

The second biofacies included an autochthonous assemblage of unionid, pachychilid and viviparid species, such as *Rumanunio rumanus*, *Tinnyea abchasica* and *Viviparus rumanus* (Fig. 8). They were mostly recorded in clayish delta-top environments, such as coastal plains, and at times in proximal delta-front sediments (Supplementary material 3). These species indicate fresher water conditions (Mandic et al., 2015; Rundić et al., 2016).

The third biofacies consisted of a mixture of broken and abraded shell and shell fragments, recorded in erosive sandstone beds. Molluscs were transported post mortem from proximal to more distal deltaic environments. They were commonly deposited within storm events or flooding surfaces (Supplementary material 3).

5.2.2. Biostratigraphy

The evolution of mollusc and ostracod assemblages was analysed throughout the studied section, in order to identify the stratigraphic position of the boundary between the Lower and Upper Dacian regional substages, as defined by Marinescu and Papaianopol (1995).

In the lower part of the investigated section, we found several index mollusc species of the Lower Dacian, such as *Stylodacna heberti*, *Pachydacna* (*Parapachydacna*) *serena*, *Psilodon munieri*, *Zamphiridacna orientalis* and *Viviparus argesiensis* (Fig. 8). They all display their latest occurrences around 445–503 m, except for *Stylodacna heberti* which extends to 621 m (Supplementary material 4). The Lower Dacian is similarly marked by several characteristic ostracod species. The most common is *Cyprideis* ex gr. *torosa*, associated with *Candona neglecta*, *Caspicypris alba*, *Camptocypris balcanica*, *Pontoniella ex gr. quadrata*, *Scottia dacica*, *Amloocypris* sp. and *Cytherissa boghatschovi* (Figs. 9, 10). Beside these species, we also noted in this interval the presence of *Amnicythere multituberculata*, *A. andrusovi*, *A. ex gr. cymbula*, *Loxoconcha schweyeri* and *L. babazananica*.

The two major index mollusc species for the Upper Dacian encountered in the investigated section are *Psilodon haueri* and *Zamphiridacna zamphiri* (Fig. 8). Their first occurrence is around 445–561 m (Supplementary material 4). We also found several characteristic ostracod species for the Upper Dacian. In this part, *Cyprideis* ex gr. *torosa* becomes more abundant. It is associated with several other species, such as *Cytherissa bogathschovi*, *C. lacustris*, *Caspicypris ornatus*, *Cyprinotus* sp., *Amloocypris* sp., *Scottia kempfi* and *S. bonnei* (Figs. 11, 12). We also

Fig. 10. Most common ostracods present in the Lower Dacian along the Slănicul de Buzău section (Part II, 1–20). 1–8. *Cytherissa bogatschovi*; 9–20. *Cyprideis* ex gr. *torosa*.
noticed low abundance of several additional species like *Pontoniella ex gr. quadrata*, *Ilyocypris bradyi*, *I. gibba*, *Darwinula stevensoni* and *Cyclocypris laevis*.

On the basis of these observations, the stratigraphic position of the boundary between the Lower and Upper Dacian regional substages was identified around 445–503 m in the Slănicul de Buzău section.

5.3. Deltaic stratigraphy

5.3.1. Regressive parasequences

The litho- and biofacies form facies associations which tend to appear in the same stratigraphic order throughout the entire section. They generally form sedimentary successions of about 15 m thick (Fig. 13), which may occasionally extend to a maximum thickness of about 40 m. These sedimentary successions begin with 1 to 13 m thick massive or laminated prodelta mudstones with autochthonous cardiid species from biofacies 1 (Fig. 13, logs A–F). Prodelta deposits are overlain by 1 to 5 m thick distal delta-front mudstones with thin sandy intercalations (Fig. 13, logs A, C–F). The transition from distal delta-front to proximal delta-front is marked by a progressive coarsening-up and the deposition of 0.5 to 3 m thick small-scale cross-bedded sandstones with thin muddy intercalations (Fig. 13, logs C–F). Successions continue upwards with several delta-top deposits, which occasionally correspond to 1 to 5 m thick massive sandstones deposited in interdistributary bays (Fig. 13, log B). These sandstones are deposited directly on top of the prodelta and delta-front sediments. They mark the transition from distal to shallower and more restricted depositional environments, without recording any deltaic sandy input. On other occasions, the deltaic input is recorded and prodelta and delta-front sediments are overlain by 2 to 5 m thick distributary mouth bars, forming...
large-scale cross-bedded sandstones (Fig. 13, logs C, E–F). Distributary mouth bars sometimes erode the underlying proximal delta-front and are directly deposited on top of distal delta-front deposits (Fig. 13, log A). Distributary mouth bars are infrequently overlain by 1 to 3 m thick channel fill deposits, creating small-scale cross-bedded organic-rich sandstones (Fig. 13, log C) or 0.2 to 0.5 m thick coastal plain deposits, with bioturbated organic-rich clays and an autochthonous faunal assemblage of unionid, pachychilid and viviparid species from biofacies 2 (Fig. 13, log E).

Each sedimentary succession displays a shallowing-upward trend, regressing from deeper open water towards shallower fluvial environments. Regressive successions are bounded by oxidized, shell- and glauconite-rich flooding surfaces with broken and abraded shell assemblage from biofacies 3 (Fig. 13, logs A–C, E–F). Flooding events produced basal erosional unconformities. These surfaces were formed during relative water-level transgressions, corresponding to delta-lobe switching, which formed in total 64 shallowing-upwards successions, defined in the literature as parasequences (Catuneanu et al., 2011). Parasequences are illustrated in a relative water-level curve, based on attributing a depth rank to the facies associations (Table 1), which highlights numerous relative water-level variations of low magnitude (Fig. 3b).

5.3.2. Regressive sequences

In addition to water-level variations of low magnitude, the relative water-level curve displays variations of higher magnitude. The 64 parasequences can be stacked in larger-scale regressive events. In other well documented cases, larger-scale events are bounded by major unconformities, usually correlated throughout the entire basin, and are defined in the literature as sequences (Catuneanu et al., 2011).
In our section, the parasequences can be grouped into nine low-order regressive sequences (Fig. 3b), each of them enclosing between 17 and 29 regressive parasequences. The low-order regressive sequences are delimited by well-developed delta-top facies, such as m-thick distributary mouth bars or channel fill deposits. The low-order sequences can themselves be stacked into three high-order regressive sequences (Fig. 3b). Each high-order regressive sequence encloses three regressive low-order sequences. The high-order regressive sequences are bounded by even shallower delta-top facies, such as m-thick channel fills or dm- to m-thick coastal plain deposits, marked by enrichment in terrestrial organic material, ichnofossils and freshwater faunas. Low- and high-order sequences highlight larger scale regressive events, related to larger relative water-level variations in the basin.

5.3.3. General regressive trend

The section records numerous water-level variations of various amplitudes, which are superposed onto a general regressive trend seen on the scale of the entire sedimentary succession (Fig. 3b). The base of the section is mostly mud-dominated and comprises 15 m thick parasequences, composed on average of 77% mudstones (Fig. 14). The muddy regressive parasequences start with m-thick prodelta deposits, showing numerous density-driven and hyperpycnal flows (Fig. 14a). Prodelta deposits are overlain by m-thick distal delta-front deposits, often disturbed by convolute bedding (Fig. 14b). The succession regresses up to dm-thick proximal delta-front deposits, forming thin sandy beds with small-scale sedimentary structures (Fig. 14c). Delta-top deposits hardly occur in the basal part of the section. Furthermore, at the base, parasequences are grouped in 94 to 137 m thick low-order sequences and in 350 m thick high-order sequences (Fig. 3b).

Towards the top of the section, the sedimentary succession becomes sand-dominated. Parasequences are on average composed of 56% mudstones, whereas the amount of sand has doubled compared to the base of the section (Fig. 15). In the sandy regressive parasequences, prodelta deposits are only dm- to m-thick, or are absent. They are overlain by dm- to m-thick distal delta-front deposits, showing frequent density-driven flows (Fig. 15a). On top of the distal delta-front deposits, m-thick proximal delta-front deposits are deposited and formed by m-thick sandstone layers. These sand beds contain various large-scale cross-beds. Sandy parasequences commonly regress up to m-thick erosional distributary mouth bar deposits (Fig. 15b), or more rarely into m-thick channel fill deposits. They are occasionally capped by cm- to dm-thick coastal plain deposits, showing organic-rich sediments with roots and burrows (Fig. 15c). The parasequences thickness decreases to 9 m towards the top of the section. These thinner parasequences can be grouped into thinner low- and high-order sequences. The low-order sequence thicknesses decrease to 65 to 121 m and the high-order sequence thickness decrease to 208 m (Fig. 3b).

5.4. Autogenic delta-lobe switching

Deltaic progradation formed numerous regressive parasequences and sequences, generated by frequent delta-lobe switching. Thanks to the robust time frame available for this section, it is possible to test if the sediment rhythmicity of this delta is autogenic or allogenic (Fig. 17). On a small-scale, we observed 64 parasequences, when there are only 27 precession cycles in the corresponding time-interval. Parasequences have a frequency of 12 ± 9 kyr. Filters with 13.6–25 m and 25–37 m bandpass-width are not in tune with the 23 kyr precession cycle. Parasequences repeat too frequently to be coeval with any astronomical cycles. On a larger scale, the nine low-order sequences display a frequency of 81 ± 44 kyr. The 64–111 m and 50–138 m filters do not correlate with the 40 kyr obliquity cycle or with the 100 kyr eccentricity.

![Fig. 13. Sedimentological logs of some of the most representative regressive parasequences observed along the section.](image)
cycle. The low-order sequences can likewise not be reliably linked to astronomical cycles. On an even larger scale, the three high-order sequences have a frequency of 243 ± 61 kyr. Similarly, they occur too often compared to the 400 kyr eccentricity cycle. In summary, it appears that neither the parasequences, nor the sequences reflect astronomical climatic forcing.

The absence of correlation between the sequences/parasequences and the astronomical cycles may be due to condensed intervals and minor hiatuses recorded in the sedimentary succession. These events might have affected the time frame of deltaic progradation. Each parasequences is topped by shell-rich oxidized layers, marked by a basal erosional unconformity, formed during flooding events. Each of these events generated a minor hiatus, due to sediment starvation and winnowing occurring during relative water-level rises. Furthermore, at about 250 m in the section, the boundary between two low-order sequences is marked by a series of four erosional shell-rich oxidized layers, stacked together in a 5 m thick interval. Each of these beds represents a full parasequence of about 1 m thick. As parasequences are on average 15 m thick elsewhere in the section, we may estimate that about 55 m sediments have been eroded or were not deposited. This resulted in a dramatic decrease in sedimentation rate from 144 cm/kyr to 65 cm/kyr (Fig. 16). The sedimentary succession seems therefore to have recorded a major hiatus in this part of the section, which may have impacted its time frame and possible correlation with astronomical cycles. However, even if the section displays some hiatuses, the parasequences and sequences are most likely the result of autogenic relative water-level variations.

6. Discussion

6.1 Palaeoenvironmental evolution of the mid-Pliocene eastern Dacian Basin

During the mid-Pliocene Dacian stage, the Dacian Basin received erosion products of the uplifting Carpathians (Fig. 1a, b). The eastern part of the Carpathians was drained by a river running parallel to the mountain belt (Jipa, 1997; Popov et al., 2006; Jipa and Olariu, 2009; Leever et al., 2010; Stoica et al., 2013; Fongngern et al., 2016; Matoshko et al., 2016). The Dacian alluvial and deltaic river system prograded southwards, with a mean palaeocurrent direction of 195° (Fig. 7d). The system prograded on the northern margin of the Dacian Basin and progressively infilled the deep southern foreland depression (Jipa, Sanders et al., 1999; Tărăpoancă et al., 2003; Panaiotu et al., 2007). The basin became overfilled during the Romanian regional stage (Jipa and Olariu, 2009). At that time, the Dacian deltaic system merged with the Danube system and sediments started to overspill into the Black Sea at around 4 Ma (De Leeuw et al., 2018; Olariu et al., 2018).

The Dacian deltaic system remained relatively stable during the entire Dacian regional stage. The long-term stability of the sediment system was ensured by an equilibrium between subsidence and sedimentation rates (Bertotti et al., 2003). The regional subsidence rate of 90 cm/kyr (Tărăpoancă et al., 2003) was balanced in the eastern Dacian Basin by average sedimentation rates of 90 cm/kyr along the more northern Râmnicu Sărat section (Fig. 2b) (Vasiliev et al., 2004) and
139 cm/kyr along the more southern Slănicul de Buzău section (Fig. 17). This balance permitted a major storage of sediments within only 0.6 Ma. About 1300 m of deltaic sediments were recorded along the Râmnicu Sărat section (Vasiliev et al., 2004) and about 835 m along the Slănicul de Buzău section. These two sections recorded the southern progradation of the Dacian deltaic system through the eastern Dacian Basin. Deltaic progradation generated a north to south decrease of sediment grain size and a thinning of the deltaic and alluvial sand bodies. In the northern Râmnicu Sărat section, sand bodies are about 2 to 3 m thick (Jipa and Olariu, 2009) and composed of medium grained sediments (Vasiliev et al., 2004). In the southern Slănicul de Buzău section, sand bodies are only 1 to 2 m thick and are composed of fine grained sediments. The depositional environment thus became progressively more distal towards the northern margin of the Dacian Basin.

The Slănicul de Buzău section records progradation of the entire Dacian deltaic system through time. The section registers a gradual coarsening-upward trend, coeval with progressive thinning of the regressive parasequences and sequences (Fig. 3b). These trends demonstrate an increase in energy of depositional processes and a decrease in accommodation space. The sedimentary succession records in parallel two major changes in faunal assemblages. The first change marks the boundary between the Lower and Upper Dacian regional substages at about 445–503 m (Supplementary material 4). This change occurred gradually, as the transitional interval extends to 621 m for some of the species. Importantly, the transition occurred independently from changes in the depositional environment. This suggests that the boundary between the Lower and Upper Dacian regional substages might be synchronous throughout the entire Dacian Basin. On the basis of the present age model, we estimate this boundary to be within chron C3n.1r at an age around 4.42 Ma (Fig. 17). The second major change in faunal assemblage corresponds to the boundary between the Upper Dacian and the Romanian regional stages. The boundary, marked by a relatively abrupt transition from brackish water to freshwater faunas, is located at about 835 m in the sedimentary succession and was dated at around 4.2 Ma (Van Baak et al., 2015). The transition in faunal assemblages is coeval to the first coal layers deposited in delta-top environments, which are observed in the upper-most part of the section (Fig. 3a). As the boundary between the Upper Dacian and the Romanian stages is linked to the depositional setting, it might be diachronous throughout the Dacian Basin.

6.2. Distinctive deltaic features in semi-isolated basins

The Slănicul de Buzău section documents the mid-Pliocene infill of the semi-isolated Dacian Basin by a substantial prograding deltaic system. This enclosed depositional environment seems to have influenced the sedimentary facies and internal architecture of this delta. The restricted basin formed a protected depositional environment, with limited wave and tide activity. Deposits are only occasionally disturbed by minor wave and storm action, creating small-scale lenticular bedding and thin shell-rich storm deposits. There is very little evidence of sediment reworking, which we relate to low-energy in this protected environment. However, the absence of indications of wave or tide...
influence could also be related to the very strong river input entering the basin. The isolated nature of the basin caused lowered waters salinities (Popov et al., 2006; Leever et al., 2010; Stoica et al., 2013). Waters with lowered salinities present lowered water densities, which means that hyperpycnal flows are more likely to occur than in regular sea waters (Sturm and Matter, 1978; Mulder et al., 2003). Fine grained organic-rich sediments deposited by hyperpycnal plumes are very common in our section. The terrestrial organic material, occasionally found on foresets of cross-bedded sandstones, confirms a proximal low-energy depositional setting. Furthermore, the section records numerous ichnofossils, in particular in interdistributary bay and coastal plain deposits, which might be favored by low salinity and low-energy settings. Moreover, this depositional environment typically encouraged enrichment in situ mollusc fauna.

The delta prograded into a shallow depositional environment. Deltaic progradation formed thin sand bodies with an average thickness of 1 to 2 m, whereas they can be more than 10 m thick in the open ocean (e.g., Olariu and Olariu, 2015). Sediments were deposited in sand bodies with an erosive base and dm- to m-thick cross-bedded strata. Sharply-based sand bodies with relatively small-scale cross-bedding are often formed in shallow depositional settings (Fielding, 2010; Vincent et al., 2010). Due to reduced water depths, deltaic progradation generated numerous thin regressive parasequences. They are on average only 13.5 m thick in our section. Parasequences are known to be relatively thin in shallow environments (Böhacs et al., 2000; Sztanó et al., 2013), whereas they can become hundreds of m-thick in the open ocean (e.g., Olariu and Olariu, 2015). Moreover, the more distal parasequences at the base of our section are on average about 5 m thicker than the proximal ones at the top of the section. This upwards thinning highlights a gradual decrease in the rate of accommodation space available through time.

Sediment progradation occurred on a low-gradient slope. Numerous thin regressive parasequences were formed by frequent migration of multiple small distributary channels, covering a wide range of palaeocurrent directions. The formation of a wider distributary area can be enhanced by low-gradient slopes (Bhattacharya, 2006; Olariu and Bhattacharya, 2006). Distributary channels were affected by repeated delta-lobe switching, occurring due to recurrent avulsion. Delta-lobe switching seems to have been strictly controlled by autogenic processes. Neither the frequency of the parasequences, low-order sequences nor high-order sequences is in tune with astronomical climatic forcing (Fig. 17). Conversely, farther to the north, in the Râmnicu Sărat sections, more proximal deltaic deposits seem to have been affected by astronomical precession cycles (Vasiliev et al., 2004). This suggests Milankovitch cycles can be registered in river-dominated deltas (Sacchi and Müller, 2004; Li and Bhattacharya, 2013), but that climate forcing of river-dominated deltas may, in some cases, become overridden by frequent autogenic delta-lobe switching in more distal environments (Castelltort and Van Den Driessche, 2003).

As autogenic delta-lobe switching occur, relative water-level rise was noted on top of each abandoned delta lobe. The upper surface of abandoned lobes consequently recorded sediment starvation and winnowing, creating hardgrounds, enriched in shell fragments and glauconite. Such layers are often found in restricted basins (Cloud, 1955; Cattaneo and Steel, 2003). Along the studied section, these layers highlight post-diagenetic induration and oxidation only during the sedimentary interval corresponding to the mid-Pliocene Dacian deltaic system. High quantities of glauconite and organic material found in the deposits suggest increased runoff, which may have occurred in response to higher global temperatures during the Pliocene (Fedorov et al., 2013). Similar iron-rich sediments were formed in other locations around the Black Sea during the Pleistocene (Neveskaya et al., 2003; Fig. 16. Sedimentation rates calculated along the section. Comparison with the low-order sequences (dark gray) and the high-order sequences (light gray).
Krijgsman et al., 2010). Increased temperatures during the Pliocene could also have caused increased weathering of adjacent land areas, leading to increased iron content of the waters in the basins involved (Muratov, 1964). The depositional controls of these layers are still not well understood and additional research is needed.

6.3. Typical example of a river-dominated delta?

The studied deltaic system was interpreted, according to the classification of Galloway (1975), as a river-dominated delta. The sediments display a strong river influence, with a sedimentation rate of 152 cm/kyr on average. In the older part of the section, sediments were preferentially transported into the basin through density-driven and hyperpycnal flows, whereas towards the younger part, they were progressively transported through fluvial channels. The sediment supply fed a multitude of deltaic lobes and distributary channels, covering a wide distributary area. The deltaic system does not display any evidence for tide interference and reveals only minor wave activity. There is no evidence for sediment reworking except during flooding events. Additionally, no sand spit formation, symmetrical wave structures or clay draping on foresets were observed along the section. The absence of these features suggests a strictly river-dominated delta.

However, in enclosed basins, where wave and tide interferences are weak or even absent, this classification might not be the most appropriate. A more relevant classification could be the one proposed by Postma (1990), who considers the sedimentary architecture of river-dominated deltas depending on water depth and gradient of the basin. These two forcing factors seem to have played an important role on the sedimentary architecture of the studied delta. With regards to this classification, we could interpret it as a mouth bar-type delta with a Gilbert-type profile. Nevertheless, this classification does not include all the basin characteristics perceived along the studied section. Classifications established on deltas evolving in the open ocean might therefore not be applicable for deltaic system prograding into (semi-)isolated basins.

7. Conclusions

During the mid-Pliocene, the Dacian Basin formed an embayment of the Black Sea. The northern margin of this semi-isolated basin was supplied by a river-dominated delta prograding southwards east of the Carpathians. Deltaic progradation gradually infilled the basin throughout the Dacian regional stage (4.8–4.2 Ma), before being replaced by a predominantly fluvial environment at the onset of the Romanian regional stage (4.2–1.8 Ma).

The delta prograded into the restricted Dacian Basin, which formed a protected, brackish water, shallow depositional environment, with a low-gradient slope. This atypical depositional setting strongly influenced the sedimentary architecture of the deltaic system, which differs from a typical open ocean delta. By contrary to well-known river-dominated deltas, this deltaic system shaped a larger number of small terminal distributary channels, experiencing frequent delta-lobe


Wenz, W., 1942. Die Mollusken des Pliozäns der rumänischen Erdöl-Gebiete als Leitversteinerungen für die Aufschluß-Arbeiten (in German). Senckenbergiana 24 (293 pp.).
