Clinoform growth in a Miocene, Para-tethyan deep lake basin: thin topsets, irregular foresets and thick bottomsets

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ABSTRACT

Late Miocene lacustrine clinoforms of up to 400 m high are mapped using a 1700 km² 3-D seismic data set in the Dacian foreland basin, Romania. Eight Meotian clinoforms, constructed by sediment from the South Carpathians, prograded around 25 km towards southwest. The individual clinothems show thin (10–60 m thick), if any, topsets, disrupted foresets and highly aggradational bottomsets. Basin-margin accretion occurred in three stages with changing of clinoform heights and foreset gradients. The deltaic system prograded into an early-stage deep depocenter and contributed to high gradient clinoforms whose foresets were dominated by closely (100–200 m) spaced 1.5–2 km wide V-shaped sub-lacustrine canyons. During intermediate-stage growth, 2–4 km wide canyons were dominant on the clinoform foresets. From the early to intermediate stages, the lacustrine shelf edges were consistently indented. The late-stage outbuilding was characterised by smaller clinoforms with smoother foresets and less indentation along the shelf edge. Truncated and thin topsets persisted through all three stages of clinoform evolution. Nevertheless, the resulting long-term flat trajectory shows alternating segments of forced and low-amplitude normal regressions. The relatively flat trajectory implies a constant base level over time and was due to the presence of the Dacian–Black Sea barrier that limited water level rise by spilling to the Black Sea. Besides the characteristic shelf-edge incision of the thin clinoform topsets and the resultant sediment bypass at the shelf edge, the prolonged regressions of the shelf margin promoted steady sediment supply to the basin. The high sediment supply at the shelf edges generated long-lived slope sediment conduits that provided sustained sediment transport to the basin floor. Clinothem isochore maps show that large volumes of sediment were partitioned into the clinoform foresets, and especially the bottomsets. Sediment predominantly derived from frequent hyperpycnal flows contributed to very thick, ca. 300–400 m in total, bottomsets. Decreasing subsidence over time from the foredeep resulted in diminishing accommodation and clinoform height, reduced slope channelization and smoother slope morphology.

INTRODUCTION

The term ‘clinoform’ was introduced by Rich (1951) to describe a synchronous, mud-dominated basinward-dipping stratigraphic surface that extends from the outer edge of a shelf (neritic water depth) through bathyal water depth to the basin floor. Steel & Olsen (2002) extended the above definition and proposed that shelf-margin clinoforms include the entire sigmoidal shape, composed of a relatively flat topset, basinward-dipping foreset and low-gradient bottomset. The clinoforms that bound the rock unit (clinothems) represent time lines within the basin stratigraphy. As the three different reaches of the clinoform tend to be sandy, muddy and sandy respectively, clinoforms highlight how timelines are not parallel to lithostratigraphic boundaries, and are commonly non-horizontal. Clinoforms form in any types of sedimentary basins (not only passive continental margins) that are sufficiently deep (generally deeper than 200 m) to develop a shelf-slope break regardless of tectonic setting (Steel & Olsen, 2002; Helland-Hansen et al., 2012). This study documents and describes late Miocene shelf-margin scale clinoforms ca. 400 m high (undecompressed) that developed in the deep lake Dacian Basin of Romania. The Dacian clinoforms are characterised by persistently thin or missing topsets, very irregular foresets and thick bottomsets (ca. 300–400 m total thickness). The continuously thin but characteristic topsets, seen to have minimal aggradation and a long-term flat shelf-edge trajectory, are uncommon but do exist in a marine setting of the Triassic
Finnmark Platform succession in the Barent Sea, of which the low-relief (ca. 150 m high) clinoforms are interpreted to form in a limited shelf accommodation condition (Handler-Jacobsen et al., 2005). The Dacian clinoforms provide an unusual record of (1) short- and long-term (100 ky and My) interplay between sediment supply and accommodation, controlled by tectonic subsidence and local climate, (2) shelf to basin-floor sediment delivery with attendant sand-mud-sand sediment partitioning, (3) forward accretion of a narrow shelf and minimal base-level fluctuation and (4) an evolving slope morphology from early to late stages of basin infill.

REGIONAL GEOLOGY OF THE DACIAN BASIN

The western Dacian Basin occupies the Late Miocene to Recent foreland of the South Carpathians Mountains. The Carpathians fold and thrust belt formed as a consequence of closure of the Neotethys Ocean that began during Late Jurassic and finalised in the Middle Miocene (Schmid et al., 2008). The Middle Miocene and younger foreland of the Southern Carpathians resulted from an event that the Paleogene to early Miocene pull-apart basin (i.e. Getic Depression) was inverted, partly detached and thrusted over the Moesia during the middle Miocene (Sarmatian) collision of the Carpathians (Rabaglià & Mačenko, 1999; Mačenko et al., 2003; Krézsek et al., 2013). In a strict sense, the term ‘Dacian Basin’ is applied only to the foreland formed by the Middle Miocene inversion and collision event and includes all parts of the Getic Depression but the thrust belts (Jipa & Olariu, 2009).

The Dacian Basin became isolated from the adjacent Tethyan lakes, Pannonian/lake Pannon (west-bounding) and Euxinian/Black Sea (east-bounding) basins and formed a reduced salinity lake by the end of Sarmatian (11 Ma) (Leever, 2007; Jipa & Olariu, 2009). The Pannonian, Dacian and Black Sea basins were separated by sills with the connections between the Pannonian–Dacian and Dacian–Black Sea basins named the Danube gateway (Leever, 2007) and Galați Seaway (Jipa & Olariu, 2009) respectively. The connectivity history of these basins is complex and has been addressed by a number of studies (e.g. Clauzon et al., 2005; Suc et al., 2011; Bartol et al., 2012). At the Danube gateway, the sill separated the Dacian from Pannonian basins by the end of Sarmatian (11 Ma); though the connection of the two basin might have been re-establish during Meotian (Leever, 2007; Leever et al., 2011). After closure, the eustatic effect on the Dacian Basin lake level was removed and thus the accommodation was then a function of tectonic subsidence and the lake water volume. Water volume in a lake is controlled by climate, whereby change is induced by Milankovitch or even in a shorter time-scale changes and the sill height (Carroll & Bohacs, 1999; Bohacs et al., 2000). Dacian Basin’s salinity decreased despite a, periodic, connection with the Black Sea through the Galați Seaway (Jipa & Olariu, 2009). During its 9–4 Ma evolution, the Dacian Basin underwent several environmental transitions between brackish and fresh water (Jipa & Olariu, 2009). There was also a transition from lacustrine to fluvial domination as the basin was filled, during the Dacian interval 4.9–4.1 Ma (Jipa & Olariu, 2013). The study of sediment dispersal patterns by Jipa & Olariu (2013) suggests that the Dacian Basin by itself was a separate source-to-sink system; with no significant amount of sediment transferred to Pannonian or Euxinian/Black Sea basins.

DATA SET AND METHODOLOGY

The study area (Fig. 1) is located in the western Dacian Basin. The data are imaged by three overlapping on-shore 3-D seismic volumes of 16 × 35 km and 30 × 35 km and 13 × 13 km (their exact locations cannot be shown since the data are proprietary to a petroleum company) with total coverage of ca. 1700 km². The frequency domain of the seismic volumes fall in a range of 20–30 Hz or an average value of 25 Hz. The best vertical resolution is at around 20–25 m when using average seismic velocity of 2200 m s⁻¹, derived from check shot data. A vertical unit of the seismic data shown in this paper is in two-way travel time (TWT, ms). Data from five exploration wells (Fig. 1) including check shots, spontaneous potential (SP) and resistivity (R) logs was used for time-depth conversion and lithologic calibration respectively. The clinoforms have been mapped based on a downlap stratal termination on seismic reflections that can be traced for tens of km in all data sets (regional scale). Clinoform morphology including slope length and gradient, and dimension of slope conduits were measured. Clinothem isochore maps based on their true vertical thicknesses (TVT) between the bounding clinoforms were produced to provide erosional and sediment depositional patterns on the clinoforms. Maps of amplitude patterns extracted on the mapped clinoforms and SP logs are used together to define relative grain size (coarser vs finer). In addition, root-mean-square (RMS) attribute extracted from the entire interval between the bounding surfaces (TVT), and well log stacking patterns are used for seismic geomorphology analysis to provide a depositional environment interpretation.

In Fig. 2, mapped surfaces in the seismic data are tied with a well log that has biostratigraphic makers. The markers, published as an internal report of the petroleum industry, indicate the boundaries of Sarmatian–Meotian and Meotian–Pontian times. In this study, the ages of these previously defined boundaries are constrained by magnetostratigraphy and micropaleontology study of Vasiliev et al. (2005) and they correspond to 9.02 and 5.8 Ma respectively. The late Miocene western Dacian Basin clinothems are defined between seismic mapped
Bounding surfaces A and I. Since age dating of surface A or any surfaces in between the defined boundaries is not available, for age estimation purpose, it is assumed here that (1) sediment supply and accommodation rates were constant throughout this interval (Y–I) and so (2) surface A which is located around mid-section, marks the time between surfaces Y (9.02 Ma) and I (5.8 Ma) or ca. 7.4 Ma and the A–I interval has an estimated total time span of 1.6 My. At least, eight clinothems are observed within the interval but only seven of them can be mapped because of data coverage limitation. Using eight clinothem numbers, the individual clinothems are therefore approximated to be of 200 ky duration. The clinothems lithostratigraphically correspond to the Middle to Upper Meotian rock unit. The unit is named after Meotian Stage, the Eastern Paratethys time domain, which is younger and older than Sarmatian–Pontian stages respectively. It is notable that since subsurface biostratigraphy from well data relies on recognition of the fauna that is tied to the lithologic units, the current subsurface chronostratigraphic framework follows a lithologic-based scheme (lithostratigraphy).

**Clinoform mapping**

The subdivision of the clinoformal succession follows the genetic sequence approach (Galloway, 1989) because the sedimentary wedge geometries in the seismic data set allow maximum flooding surfaces (MFSs) to be more easily identified and mapped than extensive erosive surfaces. Clinoforms that are defined as maximum flooding surfaces are therefore used as the bounding surfaces (BS) of the clinothems in this study. The MFSs produce both muddy and sandy horizons, depending on positions on the depositional profile. In the seismic data, the MFS surfaces are recognised as a widespread surface of downlap stratal termination. Clinoform geometry including height, foreset length and foreset gradient were measured on depositional dip-oriented cross-sections (Fig. 5). The foreset gradients are averaged values for the entire length and are calculated from the measured heights and lengths (Table 1). Due to variability in clinoform morphology along strike, the values were collected along northern and southern profiles that are ca. 10 km apart (the locations are shown in Fig. 1 as Fig. 5a, b) for comparison purposes. The sections were carefully placed within intercanyon areas because it is assumed here that slope gradients in these areas define more stable slope condition or the condition before the canyons were formed. This prevents distorted geometric measurement due to erosional processes. However, because the clinoforms migrated laterally as they prograded basinward, canyons might have shifted into the measured profiles. To systematically describe clinoforms, the basin fill is subdivided into nine seismic facies based on the seismic reflection amplitude, geometry and lateral continuity as shown in Fig. 3.

Since the absolute age control of the studied clinoforms is not available, progradation and aggradation rate cannot be accurately calculated. However, for comparison purposes, amounts of clinoform progradation (P) and aggradation (A) are measured along the northern and southern depositional cross-sections (Fig. 5) and P/A ratios are calculated for individual clinothems and the entire system (Table 2).
Shelf-edge trajectory analysis

Shelf-edge trajectory concepts (Helland-Hansen & Martinsen, 1996; Steel & Olsen, 2002 and Helland-Hansen & Hampson, 2009) are widely used to analyse clinoform evolution. A common approach of shelf-edge trajectory analysis is to monitor the vertical and lateral (landward/basinward) migration of a point, shown in a 2-D cross-sectional profile as a first break in slope basinward of the shelf (Olariu & Steel, 2009), which defines the boundary between the topset and foreset as a mean to interpret clinoform aggradation and progradation. The shelf-edge trajectory reflects a dynamic interplay between sediment supply and accommodation rates that contribute to sediment partitioning into each depositional system on a clinoform. Since shelf-edge trajectory generally reflects an interaction between sediment supply and accommodation that can vary along a shoreline, multiple depositional dip profiles should be utilised in order to capture the lateral variability (e.g. Henriksen et al., 2009). Shelf-edge trajectory classes of Helland-Hansen & Hampson (2009), namely ascending, flat and descending trajectories are used in this study.

Table I. Summary of clinoform foreset geometry (Time to depth conversion is based on seismic velocity of 2200 m s⁻¹)

<table>
<thead>
<tr>
<th>Clinoform</th>
<th>Height (m)</th>
<th>North</th>
<th>South</th>
<th>Foreset length (m)</th>
<th>North</th>
<th>South</th>
<th>Foreset gradient (degrees)</th>
<th>North</th>
<th>South</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>434.5</td>
<td>396</td>
<td>8660</td>
<td>D</td>
<td>445.5</td>
<td>309.4</td>
<td>7305</td>
<td>2.94</td>
<td>2.43</td>
</tr>
<tr>
<td>C</td>
<td>335.5</td>
<td>352</td>
<td>6520</td>
<td>E</td>
<td>337.7</td>
<td>258.5</td>
<td>6600</td>
<td>2.93</td>
<td>2.51</td>
</tr>
<tr>
<td>G</td>
<td>258.5</td>
<td>242</td>
<td>7165</td>
<td>H</td>
<td>396</td>
<td>146.3</td>
<td>6600</td>
<td>1.81</td>
<td>1.55</td>
</tr>
</tbody>
</table>

Fig. 2. Mapped seismic horizons with age controls from well 2 (see location in Fig. 1). Well logs comprise spontaneous potential (SP) and laterolog deep resistivity (LLD) curves. Sandstone lithology is differentiated from shale or mudstone by the yellow colour fill with dots. The mapped horizons, except for horizon I, are shown with colours that will be systematically used in this paper. The focused interval is in Stage 3 basin infill and contains seven clinoths (C).
Canyon and channel terminology

Shepard (1965) defined a submarine canyon as a V-shaped erosional feature that cuts into an underlying rock, whereas a valley is a feature with steep walls and flat base. In the past decades, abundant submarine conduits with varying morphology have been imaged in seismic data sets from different deepwater margins around the world. These conduits are typically characterized by seismic reflection amplitude, geometry, and lateral continuity.
the world but the naming criteria have not been consistently used. The terms submarine canyon and valley are often used to describe submarine valleys (e.g. Prather, 2003) or are differentiated by the feature’s depth, e.g. Saller & Dharmasamadhi (2012). Here, the term ‘valley’ will be discarded and a focus will be on two types of submarine or sub-lacustrine conduits; canyons and channels. The classification is based on the feature’s width, depth of the topographical expression and cross-sectional geometry. The submarine canyons are generally defined as ‘dip-oriented V-shaped to U-shaped erosional features on submarine slopes typically filled with a variety of channel deposits’ (Prather, 2003). A submarine channel is defined as a feature with negative relief created by erosion of sediment gravity flows and serves as a sediment transporting conduit into the deepwater environments (Fildani et al., 2013). In the western Dacian data set, the sizes of canyons and channels are distinctive. Sub-lacustrine canyons are typically more than 800 m wide and more than 50 m deep (Fig. 3c). The canyon bases truncate the underlying seismic reflections. In strike oriented cross-sections, they have a V-shaped appearance on the upper and middle slope and gradually become wider and shallower U-shaped features towards the toe-of-slope. In comparison, the observed slope channels are mostly <200 m wide and 25–40 m deep (Fig. 3c, g). The channels appear as spots of high-amplitude reflections (HARs) (Deptuck et al., 2003) with levees (Fig. 3g) or without levees observed (in the canyons) (Fig. 3c). These HARs have high acoustic impedance presumably due to presence of coarse lag deposits on an erosional surface (Deptuck et al., 2003). The HARs are, in the Dacian Basin data set, most likely to be equivalent to a ‘channel complex’ or a stack of erosional and filling events according to architectural hierarchy of Fildani et al. (2013).

### RESULTS

#### Large-scale basin infill architecture

Infill architecture at basin scale (Fig. 4) is described and interpreted prior to the detailed analysis of clinoforms themselves. A simplified sketch over a regional transect across the western Dacian Basin (Fig. 4a) shows structural components of the South Carpathians, Ceahlau nappe, Burdigalian Wedge or Sub–Carpathian nappe, Moesia and Dacian Basin. The Dacian Basin fill is an unshaded area above and south of Burdigalian wedge (Fig. 4a). The oldest clinoforms in Stage 2 have mainly north–south progradation direction whereas the younger, Stage 3, clinoforms prograded in northeast-southwest direction (Fig. 4b). From seismic reflection geometry and terminations patterns in the depositional dip profile, the basin evolution is subdivided into five stages from the oldest to youngest (1–5) shown by line-drawings over the dip-oriented seismic cross-section in Fig. 4b. Based on biostratigraphic age control of the tied wells (e.g. Fig. 2), boundaries between stages 1–2 (horizon Y) and 4–5 (horizon I) correspond to Sarmatian–Meotian (9.02 Ma) and Meotian–Pontian (5.8 Ma) ages respectively.

Stage 1 was characterised by a complex and uneven palaeobathymetry with the presence of east-west oriented topographical highs created by the Burdigalian Wedge (Krőzsek et al., 2013). In the northeast of the study area, a sub-vertical large tectonic structure is observed associated with fan-shaped inclined reflections on its north side. This growth structure has internal reflection geometry that changes from offlap to onlap on top or ‘overflow’ the structure, and later becomes offlap again towards the end of Stage 1. Southward of this structure, the internal reflections are nearly horizontal and onlap onto its irregular basal surface (Fig. 4b).

At the beginning of Stage 2 or early Meotian age, strata continuously buried the pre-existing structures. Chaotic sub-horizontal reflections are found toward the north. Internally, this stage contains reflections with large-scale downlap stratigraphic termination particularly in the north where the sediment package is thicker. The clinoforms in this stage dominantly prograded from north to south.

The shelf-margin clinoforms in Stage 3, the focus of this study, are separated from the underlying unit by a high-amplitude reflection, which is recognised as a clinoform downlap surface named Bounding Surface A (BS A, Figs 2 and 4b). Stage 3 clinoforms are 300–400 m high, and display well-developed breaks in slope defining the geometry of the topset, foreset and bottomset. They can be distinguished from the underlying southward migrating Stage 2 clinoforms by their dominant southwest prograding direction. The clinoform height progressively decreases toward deposition of the smaller scale (100 m) Stage 4 strata of which the base downlaps onto the top of Stage 3 (Fig. 4b). The top of Stage 4 appears as a conformable surface and corresponds to the Meotian–Pontian

### Table 2.

Progradation (P) and aggradation (A) measurements and P/A ratios of the individual clinothems. The P/A ratios calculated from the total progradation and aggradation are used as averaged values to compare with continental margins listed in Carvalhal et al. (2009) for characterization purpose, see text for discussion (time to depth conversion for aggradation is based on seismic velocity of 2200 m s\(^{-1}\)).

<table>
<thead>
<tr>
<th>Clinothem</th>
<th>North</th>
<th>South</th>
<th>North</th>
<th>South</th>
<th>North</th>
<th>South</th>
<th>P/A</th>
<th>P/A</th>
</tr>
</thead>
<tbody>
<tr>
<td>C1</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>n/a</td>
<td>5270</td>
<td>29</td>
<td>184</td>
<td></td>
</tr>
<tr>
<td>C2</td>
<td>n/a</td>
<td>n/a</td>
<td>1575</td>
<td>61</td>
<td>26</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C3</td>
<td>4870</td>
<td>9</td>
<td>553</td>
<td></td>
<td>2220</td>
<td>2</td>
<td>1009</td>
<td></td>
</tr>
<tr>
<td>C4</td>
<td>3470</td>
<td>–3</td>
<td>–66</td>
<td>2540</td>
<td>29</td>
<td>89</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C5</td>
<td>3540</td>
<td>46</td>
<td>77</td>
<td>5970</td>
<td>–17</td>
<td>–362</td>
<td></td>
<td></td>
</tr>
<tr>
<td>C6</td>
<td>1640</td>
<td>11</td>
<td>149</td>
<td>3260</td>
<td>33</td>
<td>99</td>
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<tr>
<td>C7</td>
<td>4370</td>
<td>55</td>
<td>79</td>
<td>4190</td>
<td>50</td>
<td>85</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Total</td>
<td>17</td>
<td>890</td>
<td>68</td>
<td>25</td>
<td>025</td>
<td>186</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Average 262 135
boundary (5.8 Ma) or horizon I (Figs 2 and 4b). The basin fill is represented by horizontal seismic reflections in Stage 5 that is Dacian–Romanian in age.

**Interpretation of the basin infill architecture**

The architecture of the Dacian Basin infill can be classified into; (1) basin deepening, (2) clinoform filling and (3) filled-up events. Stage 1 occurred during basin deepening and was strongly influenced by laterally variable tectonic uplift and subsidence (Rabăgia & Mațenco, 1999). The subsidence rate during this time was ca. 1100 m My$^{-1}$ (Mațenco et al., 2003). According to Leever (2007), the pre-existing topography on the southwest of the study area was the Lower Cretaceous carbonate platform of the underlying Moesian plate. The carbonate platform was formed during early stages of the Tethys closure and is capped by the ‘pre-Paratethys denudational surface’ onto which Stage 1 reflections on-lap. The structure observed in the northeast (Fig. 4b) is part of the Sarmatian–Early Meotian (Late to Middle Miocene) imbricated thrust system formed by reactivation and inversion of pre-existing extensional faults (Rabăgia & Mațenco, 1999). The thrust-bounded piggyback basins were being progressively filled during Stage 1 (Krészek et al., 2013) which resulted in the variable stratal termination styles. Although parts of the inner South Carpathians were uplifting, the contemporaneous high subsidence rate in the western Dacian Basin outpaced sediment supply in stage 1 (Rabăgia & Mațenco, 1999). Clinoform filling, including Stages 2–4, reflects sediment supply being able to keep up with subsidence. The rate of thermal subsidence decreased to around 150 m My$^{-1}$ (Mațenco et al., 2003) in Stage 2. A larger
unit thickness adjacent to and above the thrust wedge implies sediment supply from the north or northeast (Fig. 4b). The syn-depositional structural uplift and relief induced mass-transport events represented by the chaotic seismic reflections on its west side (Fig. 4b).

The sediment wedge in this stage is interpreted to be the distal part of southward prograding clinoforms (Fig. 4b). The reason of westward switching of clinoform progradation during transition from Stage 2 to 3 is not well understood. The sediment depocenter migration may have been caused by rearrangement of the drainage systems by tectonic influence as exemplified by the Early Eocene tectonics of the Western Interior, central United States (Galloway et al., 2011). Deflection of the fluvial drainage systems in the Western interior due to the Laramide intraplate-stress had affected the Gulf of Mexico shoreline by causing southward shift of the Colorado fluvial axis, formation of the Houston-Brazos axis and the Upper Wilcox Rio Grande deltaic depocenter (Galloway et al., 2011). However, Rabgá & Mațenco (1999) and Mațenco et al. (2003) report Meotian time of the Dacian Basin as a period of relatively tectonic quiescence and Rabgá & Mațenco (1999) have suggested an autogenic avulsion of the fluvio-deltaic system as a cause of westward clinoform shifting. The deltaic progradation may have followed the depositional gradient as an example from lake Texoma, USA (Olariu et al., 2012). The authors have shown that the hyperpycnal flow-dominated delta progradation, more common in a lake setting, has been deflected from the Red River mouth as a function of the river discharge and lake-floor gradient. The hyperpycnal flows tend to follow the steepest gradient (Olariu et al., 2012). The depositional gradient would be continuously decreasing as the Stage 2 clinoforms progressively filled up the basin in north–south direction and with time the deltaic system shifted westward for a steeper bathymetry. Since the Stage 3 clinoform construction relied on sediment from a fluvio-deltaic system, clinoform migration would have followed its feeder location.

The clinoforms prograded and progressively filled the basin until it became too shallow for clinoforms to develop in late Stage 4. In Stage 5, the basin became nonmarine/nonlacustrine and its infill is characterised by horizontal seismic reflections. The depositional environment was previously interpreted to have been a transition from lacustrine to fluvial environments (Jipa & Olariu, 2009).

**Clinoform characteristics**

The ca. 600 m thick clinothem stack is defined by the first mapable clinoform in Stage 3 or bounding surface (BS) A at the base and a gently inclined and upward concave horizon I (Figs 2 and 4b) which marks the transition to an interval of flat-lying reflections with lower amplitude. Seismic facies (SF) are delineated based on reflection amplitude, continuity, geometry and distinctive seismic geomorphology on attribute maps as summarised in Fig. 3.

**Bounding surface A**

Characterised by very strong positive seismic amplitude, this surface separates basin infill Stages 2 and 3 in Fig. 4b. It is locally discontinuous and overlain by bright chaotic seismic reflections that are encased in more transparent reflections, particularly toward NE of the basin margin. Where tied with well logs, the surface is represented by upward fining (Fig. 6a, d), blocky (Fig. 6b) and upward coarsening (Fig. 6c) sandstone packages. At basin-wide scale, successive seismic reflections downlap onto this surface (Figs 4b and 5). On the amplitude map, linear high-amplitude features of SF 6 are present on the basin slope and lobe (SF 8) and chaotic patches (top of SF 9) are found on the basin floor (Fig. 7a).

**Clinothem 1**

Bound by the BSs A and B, this oldest clinothem represents an initial stage of margin accretion in Stage 3 (Fig. 4b). Internally, on the slope the clinothem is filled by SF 2 (Fig. 3b) that comprises a package of 150–200 m high reflectors steeply dipping (4°–5°) (Fig. 5b). At some locations along the toe-of-slope and on the basin floor, the bottom part of the clinothem is dominated by chaotic seismic reflections of SF 9 (Figs 3i and 5b). In depositional strike cross-sections, high-amplitude reflections of SF 5 (Fig. 3e) are found associated with SF 9 on the toe-of-slope (Figs 5 and 6b). Clinothem isochore map (Fig. 8a) shows thickest accumulation, around 300 m, of sediment on the slope close to the basin margin and the sedimentary wedge becomes thinner westward. The strongest amplitude on the RMS extraction map is lobate on the basin floor (Fig. 8a).

**Bounding surface B**

BS B is characterised by a fairly smooth profile in dip cross-sections with its strongest amplitudes on its bottomset (Figs 5b and 7b). On the time structure map (Fig. 7b), the clinoform rollover is deeply incised by 0.5–2 km wide sub-lacustrine canyons. These canyons are closely spaced with inter-canyon area of around 200 m wide (Fig. 6a) and they are wider and deeper toward the north. Amplitude extraction on the surface (Fig. 7b) reveals high amplitude, strike elongate belt at the rollover and a linear pattern within the canyons (SF 5) that gradually widen up to multiple lobate features on the bottomset. Straight and bright amplitudes are observed to overlap the lobate pattern on the bottomset. Bright lineaments on the ridges between canyons (SF 6) are also observed on the forest.

**Clinothem 2**

Clinothem 2 (C2) is thickest in the northern area and tapers out toward the south (Fig. 8b). Note that none of the cross-sections in Fig. 6 intersects the thickest part
of the clinothem. The clinothem is dominated by steeply incline SF 2 that downlaps onto BS B (Fig. 5b). Along the depositional strike cross-sections, the clinothem is dominated by a vertically stacking SF 3 (Fig. 3c), large concave reflections or V-shaped canyons (Fig. 6a). High amplitude, both concave and convex spots of SF 5 and 6 (Fig. 3e, f) are observed in canyon thalwegs and ridges respectively (Fig. 6a) and further downdip SF 5 is more
Fig. 6. Depositional-strike oriented cross-section from northwest to southeast or most proximal to distal basin (a–d). The locations of the cross-sections are shown in Fig. 1. Profile (a) is from a different seismic volume and thus its amplitude and resolution appear slightly dissimilar to the others. Multiple normal faults present on the profiles have offsets around 30–40 m. Seismic facies (SF) described in the text are labelled with yellow letters. (a) Base of slope of clinoform B and mid-slope of clinoform C are characterised by prevalent closely spaced V-shaped canyons. (b) Multiple high-amplitude spots are present on canyon ridges as well as their thalwegs. (c) Varying slope morphology on clinoform E is shown by the presence of a channel-levee system (also adjacent to the canyons) in the southeast whereas canyon systems is dominant in the northwest. (d) C5 is highly aggradational and has bright amplitude on the basin floor which is a stark contrast with the underlying and overlying clinothems.
dominant (Fig. 6b). Corresponding to the cross-sections, SF 5 and SF 6 appear as linear high amplitudes within and on the ridges of the canyons on the RMS map (Fig. 8b). The latter is continuous downdip and spreads out to lobate amplitudes on the basin floor (Fig. 8b).

Bounding surface C

High-amplitude patterns highlight the clinoform topset toward its indented rollover and bottomset (Fig. 5). Well-developed V-shaped sub-lacustrine canyons are prevalent on the foreset (Figs 6a and 7c). The shelf-edge indenting canyons are 0.5–1 km wide with a spacing of 100–200 m. The amplitude surface (Fig. 7c) shows high-amplitude distribution on the topset and on the bottomset downdip of the canyons.

Clinothem 3

Clinothem 3 (C3) is bounded by BSs C and D (Figs 5 and 6) and it appears to be thickest in the northern region (Fig. 8c). The clinothem contains dipping reflections of SF 2 (Fig. 3b) internally downlapping onto BS C and at smaller scale, their underlying reflections (Fig. 5a, b). In strike oriented cross-section (Fig. 6a), 200–300 m spaced broad V-shaped (1–2 km wide) canyons, SF 3 (Fig. 3c), are still present but not as well developed as those within C2. Low-amplitude sub-horizontal facies, SF 4 (Fig. 3d), and high-amplitude concave spot, SF 5 (Fig. 3e), occur between and within SF 3. Where lithology is available from well logs, in the thalweg, a stacking of SF 4 corresponds to a succession of ca. 100 m upward fining sandstone package (Fig. 6a) which is interbedded with mudstone toward the top. Toward the toe-of-slope in the north, the clinothem is dominated by SF 5 that is observed within the canyon (SF 3) (Fig. 6b). At multiple locations, the clinothem is cut by syn-depositional normal faults that have orientation normal to the shoreline and are commonly observed adjacent to high-amplitude reflections of SF 5 (Fig. 6a–c). C3 RMS map displays high amplitudes within the canyons (SF 5) in places where some of them continue to lobate patterns on the basin floor (Fig. 8c). One of the SF 5 can be traced updip.

Fig. 7. Time-structure (left), amplitude (middle) and geomorphological interpretation (right) maps of the clinoforms (bounding surface, BS). The dashed lines present on both maps represent shelf edges and toes-of-slope. Locations of the dip and strike cross-sections are also shown on the time structure maps. The amplitude is extracted on the clinoform surfaces. The high amplitude or warm colour tones represent coarser-grained deposit than the low amplitude or cool tones. (a) BS A (b) BS B and (c) BS C. Note that besides the presence of Moesian platform underneath, the southernmost part of the maps appear to be shallow on the time structure maps due to foreland basin back-tilting effect (see also Fig. 5). (d) BS D, (e) BS E, and (f) BS F. (g) BS G and (h) BS H.
to where the canyon indents the shelf edge in the seismic volume (Fig. 8c).

**Bounding surface D**

On the northern cross-section (Fig. 5a), SF 1 (Fig. 3a), a group of subparallel reflections, is wider and brighter than on the southern section, especially close to the rollover (Figs 5b and 7d). The northern topset has no observable incision on the seismic profile (the incision on the foreset is a deep erosion on BS E). Conversely, the southern rollover, where the foreset is longest and gentlest (Table 1), is incised by a canyon of ca. 4 km wide (Figs 5b and 7d). The sub-lacustrine canyons become wider (2–3 km) and more asymmetrical than those on BS B and BS C (Figs 6b and 7b–d). High amplitude on the topset has a broad lobate feature on the northeast corner of the map and small erosional features with dim amplitude are observed incising the topset (Fig. 7d). Linear amplitude pattern highlights the canyon thalwegs (SF 5) on the foreset and at least at one location, it shows a connection with the erosional feature on the shelf where the canyon head incises the rollover (Fig. 7d). High-amplitude flat spots, SF 6 (Fig. 3f), is observed on the canyon ridges (Fig. 6b) and they display linear amplitude pattern on the amplitude map (Fig. 7d). SF 5 linear amplitudes link to lobate features observed at the transition to the bottomset and those that occur farther basinward on the bottomset (Fig. 7d).

**Clinothem 4**

Clinothem 4 (C4) shows rare SF 1 (Fig. 3a) and well-developed steeply incline SF 2 (Fig. 3b) on its uppermost part (Fig. 5a). On the foresets, truncational V-shaped features of SF 3 (Fig. 3c) associated with bright spots of SF 5 (Fig. 3e) are dominant facies (Fig. 6b). These canyons (SF 3) are 1.5–2 km wide. The strongest amplitude with laterally continuous expression, SF 8 (Fig. 3h), occurs at the toe-of-slope (Figs 5 and 6d). At the toe-of-slope, a transition from SF 3 to laterally continuous reflections of SF 8 is expressed by wide v-shaped high-amplitude reflections with high-amplitude spots of SF 5 and SF 6 (Fig. 3e, f). The reflections
appear aggradational at the base but are truncated by BS E at the top (Fig. 6c). All of the SF 4 (Fig. 3d) reflections within the C4 appear to laterally migrate to the SE (Fig. 6c) and multiple small normal faults, continuations of those in C3, are more abundant in the north. Although the clinothem is thickest in the north above BS D foreset, thick deposits are observed on the southern part as well (Fig. 8d). The RMS map (Fig. 8d) reveals high-amplitude deposit at and below BS D rollover (SF 2), linear amplitude within the canyons (SF 5) and lobate pattern at multiple locations on the basin floor (SF 8).

### Bounding surface E

The proximal topset and upper foreset are incised (Figs 5 and 6b). On the well logs (Fig. 6b, c), BS E corresponds with most positive SP log and lowest R log values inferring a muddy lithology. Right below the shelf edge, the muddy interval is underlain by ca. 10 m thick upward fining sandstone (Fig. 6b). Along the strike profiles and on a map view (Figs 6b, c and 7e) the clinoform morphology varies. Toward the north where the water depth in front of the margin is greater, the foreset is dominated by multiple 2–4 km wide shelf-edge indenting canyons (SF 3). Although towards the south ca. 200 m wide channel with levees, SF 7 and shallower canyon are present. The levees are taller on the northern than the southern sides (Fig. 6c). On the amplitude map (Fig. 7e) lobate amplitude pattern is found on the topset, close to the rollover. On the foreset, linear features of SF 5 and SF 7 highlight the canyon and channel thalwegs respectively (Fig. 7e). The linear high amplitude within the channel continues on the bottomset for around 4 km before transforming to a lobate feature. Downdip of the canyons, the amplitude pattern shows lobes at the transition to the bottomset. These lobate features correspond with a 10–15 m thick sandstone interval that is, on the well logs interpreted to represent slightly upward coarsening at the base and upward fining to the top (Fig. 6d).

### Clinothem 5

Clinothem 5 (C5) lies between BS E and BS F and is characterised by bright and smooth amplitude reflections with strongest amplitudes on the lower foresets and bottomsets (Fig. 5). On the strike profile (Fig. 6c), toward the north, a mid-slope segment of the clinothem comprises moderate amplitude reflectors of SF 3 (Fig. 3c) associated with high-amplitude concave spots.
of SF 5 (Fig. 3e) and rare chaotic reflections SF 9 (Fig. 3i). It is notable that the clinothem at this location corresponds to a succession of upward fining and blocky sandstone on the well log (Fig. 6c). Along the southern margin the clinothem is characterised by sub-horizontal SF 4 (Fig. 3d) highly aggradational asymmetry small concave reflections of SF 7 (Fig. 3g). C5 is thickest above the BS E lower foreset and bottomset in the north (Fig. 8e). However, the RMS map shows that larger bright amplitude lobes are observed on the basin floor in the south.

**Bounding surface F**

The rollover is less indented than the older clinoforms but shelf-indenting canyons are still present (Figs 5 and 7f). These canyons are asymmetrical and 1–2 km wide around the rollover (Fig. 6c, d). An ca. 4 km wide canyon occurs downdip of the southern rollover where the topset deposits are thin or missing (Fig. 7f). At this location, the surface amplitude map reveals a lobate feature on the topset close to the rollover and some high-amplitude lobes in front of the canyon mouths on the lower foreset and bottomset.

**Clinothem 6**

Clinothem 6 (C6) is a package of moderately high seismic reflections with locally high amplitude on the bottomsets. It is bounded by BSs F and G (Figs 5 and 6). On the depositional dip profile (Fig. 5), parallel SF 1 (Fig. 3a) is thin but SF 2 (Fig. 3b) or steep inclined reflections downlapping onto BS F and low amplitude, sub-horizontal SF 4 (Fig. 3d) are dominant. In addition to SF 4 that is well imaged along strike sections (Fig. 6d), the divergent and laterally downlapping SF 7 (Fig. 3g) is observed near the toe-of-slope position (Fig. 6d). The thickest sediment accumulation occurs on the northernmost and southeastern areas (Fig. 8f). The high amplitudes on the RMS map are mostly lobate and located above BS E upper foreset and bottomset in the southeast and on the bottomset in the northwest.
Bounding surface G

The foreset of BS G appears to be relatively smooth (Fig. 5) with few sub-lacustrine canyons (Figs 6d and 7g). The rollover is slightly indented but not clearly connected with the canyons that occur at the mid-lower foreset (Fig. 7g). The clinoform has nearly uniform foreset
gradient of ca. 2° (Table 1) along the entire margin. The most prominent amplitude pattern forms a strike elongate belt at the rollover whereas the foreset shows uniformly low amplitude. A high-amplitude lobate pattern aligning perpendicular to the strike elongate belt is also observed, where the rollover is most protruding in the northern and southeastern areas (Fig. 7g).

**Clinothem 7**

The internal architecture of C7 is composed of locally interrupted parallel SF 1 (Fig. 3a) that caps gently inclined seismic reflections SF 2 (Fig. 3b) and the low-amplitude sub-horizontal SF 4 (Fig. 3d) that laterally change to SF 3 (Fig. 3c) in the northwest (Figs 5 and 6d). The low to moderate amplitude SF 4 reflections are, interestingly, abnormally thick (Fig. 6d). The clinothem is thickest toward the northwest and southeast parts of the study area (Fig. 8g) corresponding to the high lobate amplitude on the RMS map. The lobate features in C7 tend to extend from the upper to lower slope (Fig. 8g); in contrary to most clinothems that the lobes are found on the lower slope and basin floor.

**Bounding surface H**

Bounding Surface H is the youngest mapable clinoform in the study area with its height over 200 m on the southern margin (Table 1). On dip-oriented cross-sections (Fig. 5) the northern foreset is smoother than the southern foreset which appears to be incised by high-amplitude SF 3 (Fig. 3c), a sub-lacustrine canyon. It is the first clinoform of which the topset can be traced throughout the study area (Figs 5 and 7h). On the topset, an east–west elongate erosional feature is observed (Figs 6c and 7h). The feature has dim amplitude but a high-amplitude pattern appears on its north side (Fig. 7h).

**Summary of clinoform features**

In summary, clinoform rollovers are arcuate in a map view (Fig. 7). In the depositional strike cross-sections (Fig. 6), the clinoform foresets appear to be highly irregular with prominent incisions of around 50–100 m deep. Multiple syn-depositional faults are also observed orientated perpendicular to the shelf edge and have approximated maximum offset of 30–40 (Fig. 6b–d). At multiple locations, a slight thickness change in the hanging-wall stratigraphy is observed, e.g. Fig. 6d. From the summarised clinoform geometry in Table 1, the clinoforms and their foreset gradients are plotted in Fig. 9. It is notable that except for the youngest clinoform G, the northern segments of all clinoforms have steeper foresets than on the southern segments (Table 1; Fig. 9). However, slope gradients of both areas along with degree of incision at the clinoform rollovers decreased during the progradation and by the time clinoform G formed, the slope of the entire margin became almost uniform (ca. 2°).

From Table 2, the overall P/A ratios of the Dacian clinoforms in the northern and southern margin are 262 and 135, respectively, with entire margin average of 198. When comparing these values with the marine clinoform progradation and aggradation database collected by Carvajal et al. (2009), the Dacian Basin clinoforms can be classified as rapidly prograding moderately deep margin type, similar to clinoforms of the Lewis-Fox Hills and Pletmos margins (Carvajal et al., 2009 and reference therein). Carvajal et al. (2009) interpreted the progradation of these margins to be mainly driven by high sediment supply, i.e. ‘supply dominated’ systems.

**Shelf-edge trajectory**

Shelf-edge trajectory is plotted on the flattened dip-oriented northern (Fig. 10a) and southern (Fig. 10b) profiles and migration path of the shelf edges and toes-of-slope associated with trajectory patterns is show in a map view (Fig. 11). The cross-sections are flattened on horizon I whose geometry appears to be relatively flat throughout the data set and thus presumably was deposited close to horizontal. The northern cross-section does not intersect the shelf edges of clinoforms A and B, whereas the southern profile cap-
tures all of the clinoforms. Since the vertical change in the shelf-edge positions are often minimal and difficult to observe from the cross-sections in Fig. 10, measurements of progradation and aggradation shown in Table 2 should be used for comparison. Flat to slightly ascending and ascending trajectory patterns during deposition of C1 and C2, respectively, are observed on the southern cross-section (Fig. 10b; Table 2). The shelf-edge trajectory has a similar pattern during development of C3. During deposition of C4, the shelf-edge trajectory is descending on the northern but is ascending on the southern segments. Corresponding to the flat to descending trajectory, the isochore maps of C1–C4 (Fig. 8a–d) show thicker deepwater deposits in the northern part of the basin. The contrast pattern of the shelf-edge trajectory along the strike is continuous through the deposition of C5–C7. On the northern section, the shelf-edge trajectory is ascending through the formation of C5–C7 whereas on the southern segment it is descending during formation of C5 and then is ascending during deposition of C6 and C7. On the plan view plot (Fig. 11), a greater distance of shelf-edge progradation is generally observed with flat or descending trajectories. However, when a shelf-edge indenting canyon is present, e.g. where the southern profile intersects C3, the shelf edge with flat trajectory appear to prograde less than the northern location of which the trajectory is slightly ascending. In addition, the toes-of-slope in the south migrate basinward at higher rate, especially downdip of the shelf-edge indenting canyons.

**INTERPRETATION AND DISCUSSION**

**The evolution of basin-margin clinoforms**

The clinoform foreset morphology is observed to spatially and temporally change (Figs 6 and 7). Thus, steeper foresets due to an increase in water depth in front of the shelf edges have induced slope failure and likely generation of erosive turbidity currents that carved out the slope canyons on clinoform B and C. A decrease in the foreset gradient marks a transition from deposition of clinoforms C to D (Fig. 9). Concurrently, sub-lacustrine canyons on clinoform D became less symmetrical than those V-shaped and equal spacing canyons on clinoform C (Figs 6a, b, and 7c, d). In addition, the degree of shelf-edge indentation and slope length varies along clinoform D strike (Fig. 7d). Clinoform E is a spectacular example of along strike variability in slope conduit styles and degree of erosion (Fig. 7e). Canyons are dominantly present in the north in contrast to channel-levee systems in the south (Fig. 6c). The northern segment slope is dominated by larger slope-confined canyons and remains at 3° (Table 1) through the deposition of clinoforms D to F (Figs 6 and 9); in contrast, the southern margin, where the slope is longer and gentler, ca. 2.5° (Table 1) smaller canyons and channel-levee systems are observed. During the formation of clinoform G, the slope gradient of the entire margin decreased to around 2°, causing there to be smoother slope morphology (Fig. 7g, h). According to the change in slope morphology associated with foreset gradients, three stage of basin fill evolution are interpreted as Early, Intermediate and Late stages.
Early stage of margin accretion (BS A to D)

The Stage 3 clinoforms (Fig. 4b) began to migrate on top of the inherited topography (the thrust wedge and Stage 2 north–south migrating clinoforms and significant mass wasting processes was dominant on the basin margin and generated erosional scars (Figs 7a and 12b). This is evident by the presence of chaotic relatively high-amplitude seismic reflections that are interpreted as mass-transport deposits on the basin floor (Figs 5a and 12b). Due to the high supply and the narrow shelf setting, subsequent rapid shoreline progradation towards the proto-shelf edge provided terrestrial derived sediment to the deepwater environment in the form of sediment aprons draping down onto the dipping sea floor. Once the delta systems arrived at the proto-shelf edge coarser/sandier sediment was transported to the upper slope in the form of shelf-margin delta fronts (Porębski & Steel, 2003) or unconfined delta-fed aprons (Galloway, 1998) (Fig. 12c). The term ‘apron’ is defined by Galloway (1998) as laterally extensive deposit fed by a line source system. In this case, distributary channels of a deltaic system are considered as a line source or more specifically ‘arcuate line source’ of the delta-fed aprons by Galloway (1998). Accumulation of the sandy delta-fed aprons might have been extensive along the depositional strike possibly due to the presence of multiple small mountainous rivers. Failure scars of mass-transport processes on the upper slope might be a focus of the sediment accumulation. The continuous deposition of delta-fed aprons at the shelf edge and on the upper slope subsequently oversteepened sedimentary wedges and triggered retrogressive failure on the slope (Fig. 12d). The failures produced turbidity currents that incised into the underlying slope deposits. The closely spaced slope-confined canyons (SF 3, Fig. 3c) were formed by this process and the recycled sediment was deposited as slope aprons (considering that the failure occurred at multiple locations along the strike and thus behaved as a line source) and basin-floor fan (Figs 6a, b and 12d). Retrogressive failure is believed to be one of the common canyon forming processes especially on margins, where closely spaced canyons are present such as the Miocene New Jersey shelf (Pratson & Coakley, 1996; Fulthorpe et al., 2000) and Plio-Pleistocene Ebro margin (Bertoni & Cartwright, 2005; Kertznus & Kneller, 2009). Pratson & Coakley (1996) proposed a retrogressive failure model to explain the New Jersey slope-canyon formation. In their model, oversteepening of sediment on the slope created a chain of collapse leading to retrogressive failure especially at the canyon head where sedimentation rate is high (headward erosion). Collapse of proto-canyon head and wall would have generated turbidity currents that in turn, by erosion, steepened the canyon head and walls to further promote retrogressive failures. In the Dacian Basin, sediment-laden stream flows were capable to form hyperpycnal flows (Mulder & Alexander, 2001; Parsons et al., 2007) that could be the dominant process incising the V-shaped slope canyons because: (1) slope channels, some are within shelf-edge indenting canyons, are observed connecting to fluvial channels at the shelf edge (Fig. 7c–e) and this implies that the turbidity currents were directly generated from the river-derived flows; (2) sustained and repetitive flows tend to be more efficient at
incising and maintaining canyons, especially those located in front of river mouths (Mulder et al., 2003). Considering erosional processes through slope aprons, the high-amplitude slope ridges (SF 6) (Figs 3f, 6a and 7b-d) might be erosional remnants of the pre-existing delta-fed aprons or perched mouth bars (Porebski & Steel, 2003; Fig. 26) or collapsed semiconsolidated blocks from the shelf edges. This stage of shelf-margin construction is comparable to the erosional shelf-margin style of Ryan et al. (2009). Clinoform geometry in this stage is largely controlled by mass wasting processes and slope healing occurred at a much slower rate than erosion (Ross et al., 1994). In addition to erosional and depositional processes, the small normal faults that are most likely syn-deposi-

Fig. 12. Block diagrams illustrating the Dacian basin infill evolution that is associated with the clinoform foreset morphology change from the Early (a–d) to Intermediate (e) and Late (f) stages (see text for detailed discussion). Note that the colour codes resemble those used in Figs 7 and 8.

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tional, commonly occur within C3 (Fig. 6b), could controlled the position of channels and canyons particularly on the lower slope.

The Ebro Margin slope (Horizon 1 of Kertznu & Kneller, 2009) is dominated by V-shaped canyons similarly to the early-stage Dacian cliniforms and then the width of the canyons and the inter-canyon area increase on the younger cliniform (Horizon 2). The authors linked the cliniform morphology with the process change from turbidity current to mass wasting processes as the basin deepened/gradient increase. Closely spaced slope-confined canyons with sharp depositional ridge found in the Dacian Basin slope are also characteristic of the Late Pliocene Ebro margin (Bertoni & Cartwright, 2005). Since incised valley or shelf-edge incision by fluvial channel is not observed, the authors suggest that retrogressive mass failure was due to the rapid muddy slope progradation as the dominant process. Moreover, inherited topography of the underlying Burdigalian Wedge in the Dacian Basin (Fig. 4) could have played a role by creating differential compaction and consequent instability of the slope profile.

Intermediate stage of margin accretion (BS D to F)

At the beginning of the intermediate stage (BS D to F), the cliniform foresets became gentler (Fig. 9). V-shaped slope canyons were still present but had lost their symmetry and became wider due to erosion and collapse of their walls (Patson & Coakley, 1996). The shelf edges became most indented on the southern margin, where however, the slopes were gentler and longer (Fig. 7d, e). The locations of most incised shelf edge correspond to where a connection of the fluvial and slope channels is observed on cliniforms D and E (Fig. 7d, e). This suggests that hyperpycnal flows play an important role on cutting and maintaining the canyons (Mulder et al., 2003) in the lacustrine Dacian Basin. Thus, the sediment supply became an important factor in addition to slope gradient in controlling, especially along strike variability, the slope morphology and deepwater sediment delivery (Fig. 12e). The contrast of slope conduit systems is well illustrated on BS E (Figs 6c and 7e) on which, the foreset was dominated by sub-lacustrine canyons except for the southeast margin, where a single large channel-levee system was present. Along-strike variation in slope conduit styles, canyon vs channel-levee systems, is also observed on the Pleistocene East Kalimantan margin (Saller & Dharmasamadhi, 2012). The central East Kalimantan slope in front of the Mahakam Delta is dominated by an unconfined channel-levee system although the northern slope is dominated by mid-slope canyons and valleys. The latter is interpreted to have been sediment starved, except for when the lowstand delta sat at the shelf edge during the Pleistocene time, thus hindering the development of basin-floor fans. The canyons were mainly created by failure while the channel systems continuously received sediment from the delta (Saller & Dharmasamadhi, 2012).

Another example of variable slope morphology along depositional strike is on the Equatorial Guinea where two canyon types have been characterised by Jobe et al. (2011). The “Type I canyons” are typically erosional active slope conduits whereas their ‘Type II canyon’ does not indent the shelf edge, lacks basin-floor fans and is filled with smooth draping hemipelagic or dilute turbidity current deposits. Type II canyons are interpreted as formed under low or lack of sediment supply.

Although the Dacian Basin slope morphology is similar to the East Kalimantan slope, a lack of sediment supply cannot explain the Dacian Basin slope morphology variability because most canyons indent the shelf edges whereas the Type II canyon of Jobe et al. (2011) and the East Kalimantan canyons occur at middle slope or do not indent the shelf edges. Sub-lacustrine channels (SF 5; Fig. 3e) are observed in most canyons (Fig. 6). In addition, the high-amplitude patterns on the slopes and basin floors imply that the Dacian cliniforms were not preferentially filled by dilute turbidity currents with fine grained sediment because coarse grained deposits are present within the canyons or channels and on the basin floors (Figs 7e and 8d, e). The contrast in turbidite systems and slope conduit styles is thus likely due to a combination of the slope gradients and sediment supply. Since during the intermediate stage, the slight thrust fault movement might result in the steeper slope (by 0.5°) in the north. Thus, the margin might have been more failure prone. The sub-lacustrine channels on cliniform E and within C5 are connected to the fluvial channel at the shelf edge (Figs 6c and 7e) would imply direct fluvial sediment input with constant and probably larger sediment volume (sustained turbidity current, i.e. hyperpycnal flow and minimal sediment loss from dispersion by shelf processes) to the deepwater. With more sediment supply relative to accommodation, the southern slope is longer and gentler (Table 1). Similarly to the Early stage, multiple syn-depositional normal faults exert control on the locations of the slope conduits in conjunction with the sediment input points. Overall, this stage of margin accretion involves a high degree of shelf-edge incision, slope erosion, sediment recycle and sediment bypass to the deepwater environments (Fig. 8d, e).

Late stage of margin accretion (BS F to H)

The last stage is characterised by low subsidence rate and low accommodation space. The basin was being filled up by the cliniforms (Fig. 12f). Slope and basin-floor aggradation (Fig. 8f, g) reduced the margin slope-to-basin floor relief and gradient (Figs 9 and 11) and the upper slope environment became more stable (Ross et al., 1994). The Dacian Basin during this stage was characterised by lower than 250 m of cliniform relief. The relatively smooth depositional profiles and dim amplitude pattern on the slopes and most of the basin floors of C7 (Figs 5 and 8g) suggest that little coarse grained sediment was transported to the deepwater environments. Rareness of sub-
lacustrine canyons and channels led to less-confined turbidite systems on the slope (Fig. 8f, g) and the slope was being healed during evolution from BS F to G. Above BS F, the seismic reflections gradually change from smaller but steeper dipping clinoforms downlapping onto the underlying deposits interpreted as shelf-edge delta clinoforms (Fig. 5b).

**Shelf-edge trajectory analysis and the clinoform configuration**

The slight contrasts between the shelf-edge trajectories along different parts of the basin margin could have resulted from local sediment supply variability (see Olariu & Steel, 2009 for strike variability in the modern shelf edges, Sanchez et al., 2012a for an example from Miocene northern Carnarvon Basin, and Olariu et al., 2012 for a change in depocenter when a successive deltaic system reached the shelf edge in Maastrichtian Fox Hills Formation, Washakie Basin) or inherited basin physiography associated with local subsidence and process regime at the shelf edge as discussed in the case of Waterford Formation in Karoo Basin, South Africa (Jones et al., 2015). In the Dacian Basin case, a stark contrast of sediment supply along the shelf edge on the northern and southern sections that are 10 km apart is less plausible. Particularly during deposition of C4, when the trajectory was descending in the north but ascending in the south (Fig. 10), if sediment supply was to be lower in the northern area and could not keep up with the base-level rise as suggested by the ascending trajectory on the southern segment, the significant progradation and descending trajectory should not be observed. In addition, most clinothems are thicker in the north and this implies ample sediment supply (Fig. 8a–d). Since parts of the clinoforms formed in the piggy-back depocenters, the difference in the trajectory along the shelf edge is interpreted to result from locally variable basin subsidence caused by the thrust movements as illustrated in Fig. 13. On the northern section, an E-W trending thrust fault is present under the clinoform interval (Figs 4 and 5a). Although the maximum tectonic activity occurred during Sarmatian time (Rabăgia & Mățenco, 1999), the fault movement could have persisted through the Meotian, but with much slower rate as suggested by the growth structure above the thrust wedge and the associated mass-transport deposits in Stages 1 and 2 (Fig. 4b). During formation of C1–C3, the early clinoforms were accreting over the faulted northern area as well as on the relatively undeformed southern area (Fig. 13a). The uplift along the thrust fault gradually ele-

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**Fig. 13.** A sketch of along strike shelf-edge trajectory variability. The red solid line schematically shows how the subsidence pattern changes with time. The flat line, also shown as a dashed line for comparison, represents uniform thermal subsidence. A change in the shape of the line denotes altered subsidence. (a) During C1–C3 deposition, part of the individual clinoforms was accreting above the thrust fault in the north and on the sedimentary wedge in the south. (b) During deposition of C4, the fault uplift elevated the northern shoreline and this resulted in relative base-level fall even when the actual base level, recorded in the south, was rising. (c) In the north, C5 was built out into a flexure zone in front of the thrust, and with the higher rate of subsidence than overall falling lake level, the northern shoreline experienced a base-level rise. (d) The shelf edge on both profile recorded a consistent pattern of relative base-level rise during the deposition of C6 and C7.
vated the northern shoreline and resulted in relative fall of base level whereas the southern shoreline, under influence of the regional thermal subsidence and climate, was experiencing base-level rise during the deposition of C4 (Fig. 13b). A shelf incision is not observed on the northern seismic profile (Figs 5a and 10a) during this base-level fall. The absence of incised valleys here may be that they are too shallow to be resolved by the seismic resolution or they are truly absent due to equal or steeper fluvial gradients than on the shelf (Posamentier et al., 1992; Muto & Swenson, 2005). In contrast to the previous episode, the basin floor southwest of the thrust is influenced by locally higher flexural subsidence and when the shoreline approached the region, the northern shoreline would ‘feel’ that the base level was rising, even though the base level was falling at the basin scale in the time of C5 (Figs 10 and 13c). Later during formation of C6–C7, the base level was rising and the trajectories on both sections display a consistent ascending pattern (Figs 10 and 13d).

In summary, although the Dacian Basin lake level was controlled by regional tectonic subsidence, climate and sediment supply, local tectonic movement along smaller structures could have been a primary parameter that influenced base-level signature preserved in the stratigraphy. Dominance of tectonic control on the Dacian Basin base level has also been suggested by Rabagia & Matei (1999) as well as in the lake Pannon by Sztanó et al. (2013). Moreover, shelf-indenting canyons might have exerted some control on shelf-edge migration in the Dacian Basin. For example, the shelf edges appear to prograde less during the formation of C2 and C3 where the canyon heads are present close to the southern dip profile (Fig. 11) and despite the flat trajectory of C3. At these locations, the sediment could be captured by the canyon.

Fig. 14. (a) Late Meotian–Early Pontian paleogeographic map modified from Leever et al. (2009). The map shows locations of the gateways that are connections between the Para-tethyan basins and the Mediterranean. (b) A schematic diagram showing the Dacian Basin as a hydrologically closed entity due to the presence of a sill/barrier to the Black Sea Basin, namely Galați seaway. Note that the clinoform migrating direction is not drawn based on real data (northeast-southwest for the Dacian Basin). In addition to tectonic subsidence and climate, the basin base level was controlled by the sill elevation, i.e. lake water could not rise above the sill. This limited the lake-level fluctuations and the subsequent clinoform topset aggradation, in contrary to the neighbouring Black Sea. (c) Left, a vertically exaggerated shelf-edge trajectory from Fig. 10b. The individual Dacian clinothems have topset aggradations <60 m. The base-level curve is ‘clipped’ at the sill level and remains constant until it turns into a falling stage. Right, a eustatic sea level curve during 9–5.8 Ma from Miller et al. (2005). Larger sea level fluctuations by eustasy are reported for the study interval and might have influenced the Black Sea base-level change but not the isolated Dacian Basin.

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heads and bypassed to the deepwater environments as thick sediment accumulation found downdip of the indented shelf edges (Fig. 8b, c).

Sill control on topset thickness

The Meotian–Dacian clinoforms display a characteristically and remarkably thin to absent topsets, whereas their bottomsets are highly aggradational (Fig. 5). Furthermore, the clinoforms demonstrate a consistent oblique progradational pattern of toplap, suggesting erosion and degradation at times. However, limited shelf accommodation is not the entire story of the persistently meagre topsets and overall flat shelf-edge trajectory over the time periods of >2 My (Figs 5 and 10). Within a time interval of some 200 ky, cycles of base-level rise and fall should have been observed from the stratigraphic record. The relatively constant base-level signal is therefore interpreted to have been a consequence of a topographical barrier or ‘sill’ named Galați Seaway between the Dacian and the Black Sea basins (Leever, 2007; Jipa & Olariu, 2009) (Fig. 14a). During humid periods, water discharge from the surrounding rivers and precipitation would have caused lake level to rise, but once the water rose above the sill, the positive water balance would have been systematically drained off, spilling to the Black Sea, i.e. overfilled lake type of Carroll & Bohacs (1999) and Bohacs et al. (2000) (Fig. 14b). This condition caused constant base level at the sill elevation instead of rising according to a normal base-level cycle and resulted in a ‘clipped’ base-level curve (Fig. 14c). The lake level could fall below the sill level for example when the evaporation exceeded water input in drier periods or the tectonic subsidence rate outpaced water input rates from the rivers and precipitation. With the water level stable at or occasionally falling below the sill, the Dacian Basin was in a ‘balanced-fill’ lake type (Bohacs et al., 2000) and consequently, the accommodation that allowed 10–60 m topset aggradation was then a function of the tectonic subsidence and climatic cycles (Fig. 14c). In contrary to the Dacian Basin, the shelf edge of the Black Sea clinoforms is observed to have vertical variability in a range of 100 m (Fig. 14 and Fig. 4 of Munteanu et al., 2012). In addition, the eustatic sea level curve during 9–5.8 Ma shows sea level fluctuations of 30–70 m (Miller et al., 2005). If the Dacian was to be connected to an open marine system, the similar magnitude of base-level fluctuation, 30–70 m of sea level changes plus an influence of tectonic subsidence, should have been recorded.

Thick Bottomsets

The aforementioned limited shelf accommodation with relatively high sediment supply caused prolonged normal and forced regression with flat and descending trajectories. These conditions facilitated rapid shoreline migration to the shelf edge and promoted sediment transport to the deepwater areas of the basin. With the normal regression alone, it is known that deepwater fans can form provided high supply (Carvajal & Steel, 2006). The forced regression could enhance sediment gravity flow and mass wasting activities and subsequently, canyon cutting processes (Posamentier et al., 1992). Besides, lower suspended-sediment concentrations are required to generate hyperpycnal flows in a reduced salinity lake Dacian Basin than in a marine basin (Mulder & Svivitski, 1995). Thus, a higher frequency of hyperpycnal flows could be expected and would have enhanced river-derived sediment transport to the slope and basin floor as illustrated by the connection of slope conduits with the delta distributors channels at the shelf edge on clinoforms C, D and E (Fig. 7c–e). In summary, the persistent sediment bypass and transport to the deepwater environment through sub-lacustrine canyons and channels during prolonged regressions contributed large volumes of sediment to the basin floor, thus making very thick clinoform bottomsets (Figs 5 and 14a).

Is a very thin topset characteristic of lake clinoforms?

Most of studied shelf-margin clinoforms of various tectonic settings such as Eocene Spitsbergen (piggy-back basin) (Johannessen & Steel, 2005); Paleogene Porcupine and Plio-Pleistocene Ebro (rifted basins) (Kertznus & Kneller, 2009; Ryan et al., 2009) and Washakie Foreland basin (Carvajal & Steel, 2006) developed in marine settings and they all show significant amount of topset aggradation. Therefore, a key question is whether remarkably thin topsets associated with flat shelf-edge trajectory a characteristic of lake basins? In the adjacent Pannonian Basin, the Mio-Pliocene (11.6–6.8 Ma) lake clinoforms are up to 700 m high and they have 100–200 km wide topsets with significantly aggradation (>1000 m maximum thickness in the Derecske trough, Juhasz et al., 2007). These Pannonian clinoforms resemble other marine clinoforms more than the lake Dacian clinoforms. The fundamental difference of the Dacian and Pannonian basins is their tectonic settings. Lake Pannon was of backarc origin and the lake-level change was influenced by fast tectonic subsidence and back-tilting (Juhasz et al., 2007) whereas the Dacian Basin is a foreland basin with a slow, post-tectonic subsidence profile (Matenco et al., 2003). Hence, it is concluded here that the thin topset configuration is not typical for deep lake basin clinoforms and clinoform geometry in any environment setting primarily depending on interplay between sediment supply and accommodation space (Helland-Hansen & Martinsen, 1996; Steel & Olsen, 2002). Marine clinoforms with thin topsets, despite of rare occurrence, have been document in the Triassic formation in the Barent Sea by Hadler-Jacobsen et al. (2005). Nevertheless, due to a lake basin’s closed environment and presence of a sill, a significant base-level rise with low tectonic subsidence rate is limited and thus clinoforms in such lake basin have higher poten-
tial to achieve consistently thin topset geometry than marine clinoforms.

**CONCLUSIONS**

(1) The Dacian clinoform growth geometry is primarily a function of sediment supply and accommodation space created by the subsidence and lake-level changes. Mi- Pliocene limited shelf accommodation due to the presence of the Dacian–Black Sea sill and slow thermal subsi- dence contributed to the clinoforms with persistently thin topsets. Large amount of sediment budget was bypassed to the depocenters of the deepwater environments. High deposition rates on the upper slope triggered retrogressive failure that formed the prevalent slope canyons that later became the focus of gravity-induced sediment transport including turbidity currents, hyperviscous flows and mass wasting processes. The prolonged regressions facilitated steady sediment supply at the shelf edges and significant amount of sediment was transported through the slope conduits to the basin floor and contrib- uted to the thick bottomset configuration.

(2) In the Dacian Basin, the Stage 3 clinoform development associated with varying foreset morphology can be divided into Early, Intermediate- and Late stages. The early-stage clinoforms (A–D) are characterized by steep, 3°, slope gradient that is dominated by closely spaced canyons that are uniformly present along strike. During the Intermediate stage or clinoform s-D- F, the slope morphology varies along the margin. The steeper northern slope is dominated by sub-lacustrine canyons whereas the channel-levee system is observed on the gentler southern slope. In this study, the presence of the sub-lacustrine canyons is not an indicator of a lack of sediment supply but continuously active canyon erosion. In the Late stage or clinoform F–H, the foreset gradients decrease due to progressively reduced water depth. Degree of shelf-edge indenta- tion was minimal and the slope tended to be less dis- rupted by the sub-lacustrine canyons or channels.

(3) Variation in shoreline and shelf-edge trajectories along strike are proved common but is still underappreciated. Local tectonic movement could cause relative base-level change along the basin margin and subsequent variation in shelf-edge trajectory patterns. To constrain the base-level behaviour at times, local sediment supply and tectonic movements should be taken into account with shelf-edge trajectory analysis that is done on multiple profiles along the shoreline.

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