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KRISTĪNE TOVMASJANA

DEPOSITIONAL ENVIRONMENT OF THE TIDALLY-DOMINATED TRANSGRESSIVE SUCCESSION: RĒZEKNE AND PÄRNU REGIONAL STAGES, BALTIC DEVONIAN BASIN

DOCTORAL THESIS

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Supervisor:
Ģirts Stinkulis, assoc. professor, *Dr. geol.* (University of Latvia)
Scientific advisor:
Piret Plink-Björklund, assoc. professor, *PhD.* (Colorado School of Mines, USA)

Reviewers:
Ervīns Lukševičs, professor, *Dr.geol.* (University of Latvia)
Juris Soms, docent, *Dr.geol.* (University of Daugavpils)
Andis Kalvāns, researcher, *Dr.geol.* (Tartu University)

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Vitālijs Zelčs, professor, *Dr.geol.* – chairman
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Laimdota Kalniņa, assoc. professor, *Dr. geogr.*
Valdis Segliņš, professor, *Dr. geol.*

Secretary:
Ģirts Stinkulis, assoc. professor, *Dr. geol.*

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The thesis is available at the Scientific Library of the University of Latvia, Rainis Blvd. 19 and Alberta Street 10, Riga.

Address for submitting the comments:
Dr. Ģirts Stinkulis, Department of Geology, University of Latvia, Raiņa Blvd. 19, LV-1586, Riga. Fax: +371 67332704, e-mail: Girts.Stinkulis@lu.lv

CONTENTS

Abstract

The thesis focuses on the interpretation of the depositional environments of Rēzekne and Pärnu siliciclastic succession of the Baltic Devonian Basin. Previous studies of this succession were based mainly on lithological, mineralogical and stratigraphical data, as well as to some sedimentological description of deposits in the eastern parts of the basin. To date no detailed overall synthesis of the depositional environment of the entire succession existed. Thus, the palaeoenvironmental setting of Rēzekne and Pärnu basins remained the subject of discussion and re-evaluation. In the past decade, the role of facies analysis in reconstruction of palaeoenvironments of Middle to lower Upper Devonian has been emphasised. The findings of the thesis, based on detailed facies analysis and sequence stratigraphic implications, indicate that the succession was deposited under the strong dominance of tidal processes in transgressive estuarine and tidal flat settings. High meso- to macrotidal ranges are suggested for this basin.

Keywords: facies analysis, sedimentary environments, sequence stratigraphy, tidal processes, estuaries, tidal flats, Baltic Devonian basin.

Anotācija

Disertācijas pētījumu tēma ir veltīta Rēzeknes un Pērnavas reģionālo stāvu nogulumu fāciju analīzei un to uzkrāšanās vides rekonstrukcijai Baltijas valstu teritorijā. Agrāk šai devona klastisko nogulumu slāņkopas daļai tika veltīti galvenokārt litoloģiski, mineraloģiski un stratigrāfiski pētījumi, kā arī nogulumu sastāva raksturojums baseina austrumdaļā. Detalizēti pētījumi ar mērķi rekonstruēt kopējo baseina attīstību līdz šim netika veikti. Tādēļ Rēzeknes un Pērnavas reģionālo stāvu nogulumu veidošanās apstākļi bija maz zināmi, kā arī diskutabli. Pēdējo desmit gadu laikā fāciju analīze tika plaši pielietota vidējā un augšējā devona klastiskās sedimentācijas baseinu rekonstrukcijās. Disertācijas darbs balstās uz fāciju analīzi un secību stratigrāfijas metodoloģijas pielietošanu, un pētījumu rezultātā tika secināts, ka Rēzeknes un Pērnavas laikposmos Baltijas devona baseinā nogulumu uzkrāšanos noteica plūdmaiņu procesi, turklāt kopumā transgresīvajā sistēmā veidojās estuāri un attīstījās plaši plūdmaiņu līdzenumi.

Atslēgas vārdi: fāciju analīze, plūdmaiņu procesi, sedimentacijas vide, secību stratigrāfija, estuāri, plūdmaiņu līdzenumi, Baltijas devona baseins.

Introduction

Devonian deposits are widespread in the territory of Baltic States and adjacent regions. The Lower, Middle and lowermost Upper Devonian succession consist mainly of siliciclastic rocks such as sandstones, siltstones and clays with a minor role of dolostones and gypsum (Kurshs, 1975; 1992). It forms more than 500 m thick tidally influenced complex that contains both transgressive and regressive tidally influenced and tidally dominated deposits (Plink-Björklund, Björklund, 1999).

This study focuses on Rēzekne and Pärnu Regional Stages (RS) of the Baltic Devonian Basin (BDB) that represents a complex transgressive succession of fluvial and tidal transition zone to outer estuarine deposits in tide-dominated estuaries associated with adjacent tidal flat complex that developed in a shallow, epicontinental sea.

The implications of tidal control to the deposition of Middle and lowermost Upper Devonian siliciclastic succession in general have been discussed by Plink-Björklund and Björklund (1999). The most recent studies by Pontén and Plink-Björklund (2007), Tänavsuu-Milkeviciene et al. (2009), Tänavsuu-Milkeviciene, Plink-Björklund (2009) and Pontén, Plink-Björklund (2009) discuss the depositional environment of the BDB in mid and early Late Devonian. The authors describe tide-influenced retrograding shallow marine sedimentary environment and prograding delta complex of the Narva Formation (Tänavsuu-Milkeviciene et al., 2009), delta plain and delta front deposits associated with a large southward prograding delta complex of the Aruküla Formation (Tänavsuu-Milkeviciene, Plink-Björklund, 2009) and Gauja Formation (Pontén, Plink-Björklund, 2007), as well as compare barforms from deltaic succession of the Gauja Formation with bars of estuaries in the Amata Formation (Pontén, Plink-Björklund, 2009).

The results of this study complement the results given by the above-mentioned authors on the overall development of the BDB in the Middle and early Late Devonian, which contributes to reconstruction of the sedimentary environments at different evolutionary stages of the BDB: from transgressive to regressive settings.

The motivation

Although recognized in ancient geological record (Rahmani,1988; Richards, 1994; Kvale et al., 1995; Lanier, Tessier, 1998; Plink-Björklund, 2005; 2008), as well as from modern environments (Dalrymple et al., 1990; Allen, 1991; Allen, Posamentier, 1993; Fenies, Tastet, 1998), differentiating ancient estuarine deposits from deltaic successions is often difficult and equivocal due to the similarity of the tide-dominated estuarine deposits with tide-dominated deltaic deposits in the rock record (*see discussions in* Walker, 1992; Richards, 1994; Boyd et al., 2006; Dalrymple, Choi, 2007). This is especially relevant to sand-rich estuarine successions, such as the Rēzekne and Pärnu RS, which have been previously interpreted as fluvial channel or distributary fills and shallow marine to delta deposits in the BDB (Kurshs, 1975; 1992; Kleesment, 1997). These interpretations, including "facies zonations" of deposits in Baltic States by Kurshs (1975), were based mainly upon lithological and mineralogical data, as well as on sedimentological studies of deposits in Eastern Latvia (Stinkulis, 1998). To date no detailed synthesis of the depositional environment of the Rēzekne and Pärnu RS exists. Thus, the larger palaeoenvironmental setting of the Rēzekne and Pärnu Basin remains the subject of discussion and re-evaluation.

Aim of the study

Compared with published data on wave-dominated estuaries, the data on tide-dominated estuaries remain relatively rare, primarily because tide-dominated estuaries are not that common (Tessier, 2012). In order to be able to recognize tide-dominated estuarine deposits in rock record and to differentiate tide-dominated estuarine deposits from tide-dominated deltaic deposits, detailed facies analysis is needed in the appropriate stratigraphic context. This especially concerns sand-rich estuarine systems, which are even more complicated to recognize due to their monolithic composition and which are up to date poorly described.

The aim of this study was to give a detailed description of facies of the Devonian Rēzekne and Pärnu succession in Baltic States, identify the major facies associations, describe their temporal and spatial relationship, provide a detailed facies model of the basin as a whole, and reconstruct the palaeoenvironments in the basin.

Main tasks

The main task of the thesis was to determine the overall trends of the deposition in the basin, therefore detailed facies analysis of deposits derived from drill-cores from various parts of distribution area of the Rēzekne and Pärnu RS has been carried out, in combination with studies of lateral and vertical variations of facies studies of the Pärnu Formation (Fm) in the outcrop area (Figure 1). All together fourteen drill-cores representing the Rēzekne and Pärnu RS have been described and the data from thirty additional drill-cores have been used from the previous works on the the Rēzekne and Pärnu succession in Baltic States (Sorokin, 1981; Kurshs 1992; Kleesment, 1997; Stinkulis, 1998). The data from the outcrop area with twenty two measured sections was combined with drill-core data that provides a 3-dimensional picture of the depositional environments, which is essential for understanding the changes in basin geometry and morphology. Such a detailed study, based on sedimentary facies analysis with sequence stratigraphic implications, provides a significant contribution not only to the interpretation of the sedimentary environments in the BDB, but also to the discussion on tidal influence in ancient basins.

In summary the tasks were as follows:

- To describe and interpret sedimentary facies of deposits in outcrops and from drill-cores (the Rēzekne and Pärnu RS);
- To study and describe in details the composition, facies and architecture of the beds in outcrops of the Pärnu Fm;
- To identify facies associations from described facies of the Rēzekne and Pärnu RS;
- To reconstruct the sedimentary environements and describe the evolution of the BDB in the Rēzekne and Pärnu time, including based on principles of sequence stratigraphy;
- Based on the above, to discuss the role of tidal processe in the BDB, including to provide with some criteria in recognition of sand-rich tide-dominated estuarine systems.

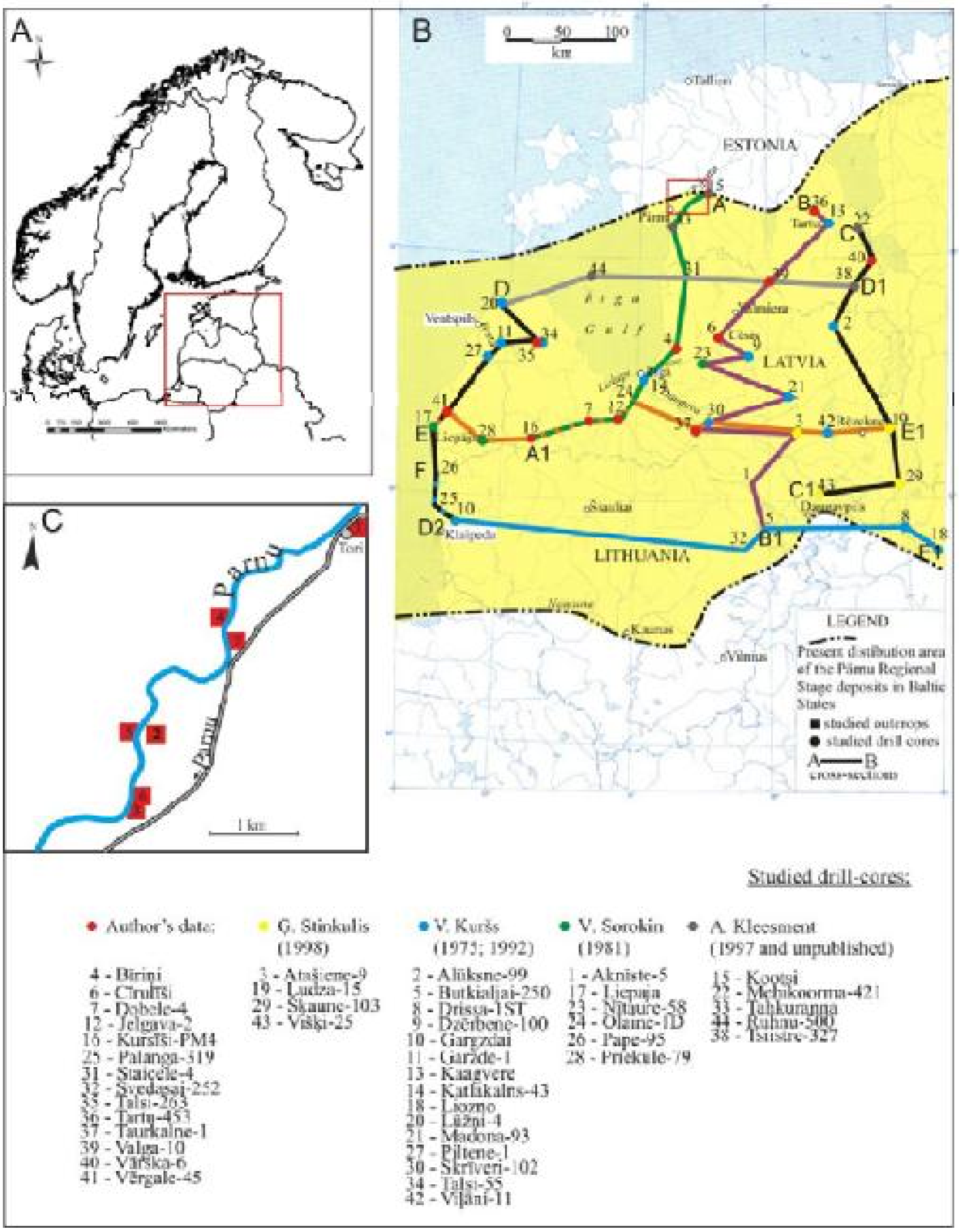

Figure 1. Location of the study area and present distribution of the Pärnu Regional Stage deposits

A – general map; B – map of present day distribution of Pärnu Regional Stage deposits. Note, that the location of drill-core 18 (Liozno) is out of scale, it is located 180 km to SE from drill-core 8 (Drissa-1ST). Liozno and Drissa-1ST drill-cores represent Rēzekne and Pärnu succession analogues in Belarus (Golubcov, Mahnach, 1961; Kurshs, 1975); C – location of the outcrops

1. attēls. Pētījumu teritorijas izvietojums un Pērnavas reģionālā stāva mūsdienu izplatības karte

A – pētījumu teritorijas novietojums pārskata kartē; B - Pērnavas reģionālā stāva mūsdienu izplatības karte. 18. urbums (Liozno) ir ārpus kartes mēroga, tas atrodas 180 km uz DA no 8. urbuma (Drissa-1ST). Liozno un Drissa-1ST urbumos ir sastopami Rēzeknes un Pērnavas slāņkopu analogi Baltkrievijā (Golubcov, Mahnach, 1961; Kurshs, 1975); C – atsegumu novietojums pētījumu teritorijā

Novelty of the research

In this study new interpretations on the depositional environments of lower part of Middle Devonian succession of Baltic States are presented. The deposits of the Pärnu Fm in the outcrop area (SW Estonia) are re-interpreted as a tripartite transgressive tide-dominated estuarine succession associated with tidal flat complex. The Rēzekne and Pärnu succession in the Baltic States records deposition in fluvial channels along the shoreline that become more tidally influenced basinward, separated by sandy to muddy and carbonate tidal flats, which fringe the shoreline, as well as tidal bars and sandflats with the superimposed smaller scale bedforms and elongate tidal ridges, that occur offshore from the flats.

The study reveals the great impact of tidal processes in the sediment movement and deposition throughout the basin. It discusses the possible tidal ranges that prevailed in the basin, which contributes to the understanding of tidal processes in ancient epicontinental basins. Although the role of palaeotides for the ancient epicontinental basins is very persuasive, it has not really been resolved to this day (Flemming, 2012).

Examples of tide-dominated sand-rich estuaries in the rock record remain very rare (Tessier, 2012). Some criteria for recognition of sand-rich tide-dominated estuaries in the rock record, as well as the distinction of sand-rich tide-dominated estuaries with respect to facies and stratigraphy of sediment infill are proposed and discussed here.

Special attention is given to the description of the fluvial-marine transition in tidal zone, which is one of the most complicated areas on Earth in terms of sedimentary environments, as the large number of terrestrial and marine processes interact and determine the complex nature of the deposits (Dalrymple, Choi, 2004). The nature of the deposits in this zone, which is transitional between purely fluvial deposition beyond the tidal limit and almost purely tidal sedimentation at the seaward end, is not known in detail to date (Dalrymple et al., 2012). A comparison of sedimentary structures found in purely tidal environment with those found in fluvial-tidal transition zone is given in detail which is of crucial importance to the sedimentary reconstruction of ancient estuarine environments.

Study region and stratigraphy

The Rēzekne and Pärnu RS in the Baltic States consists of variable siliciclastic deposits with a total thickness from 15 to 70 m, which forms a part of 500 m thick siliciclastic succession of the Middle Devonian to lowermost Upper Devonian sequence in the region (Kurshs, 1975; 1992; Kleesment, 1997). In eastern Latvia and Estonia the Pärnu RS is divided into two formations: the Rēzekne and Pärnu, based on lithological data, whereas for the rest of the Baltics only Pärnu Formation is distinguished (Kurshs, Stinkulis, 1998, Figure 1.2). It needs to be noted that the correct stratigraphic position of the lowermost part of the succession - the Rēzekne Fm - is still under discussion, and the opinions if it belongs to the Lower or Middle Devonian are still controversial (Paškevičius 1997; Kleesment, Mark-Kurik 1997; Savvaitova 2002; Mark-Kurik, Põldvere 2012). In this study it is assumed that both Rēzekne and Pärnu Fm in Latvia, as well as the Rēzekne and Lemsi Fm of the Rēzekne RS in Estonia belong to a lower part of a transgressive cycle in the Middle Devonian succession and are thus described together.

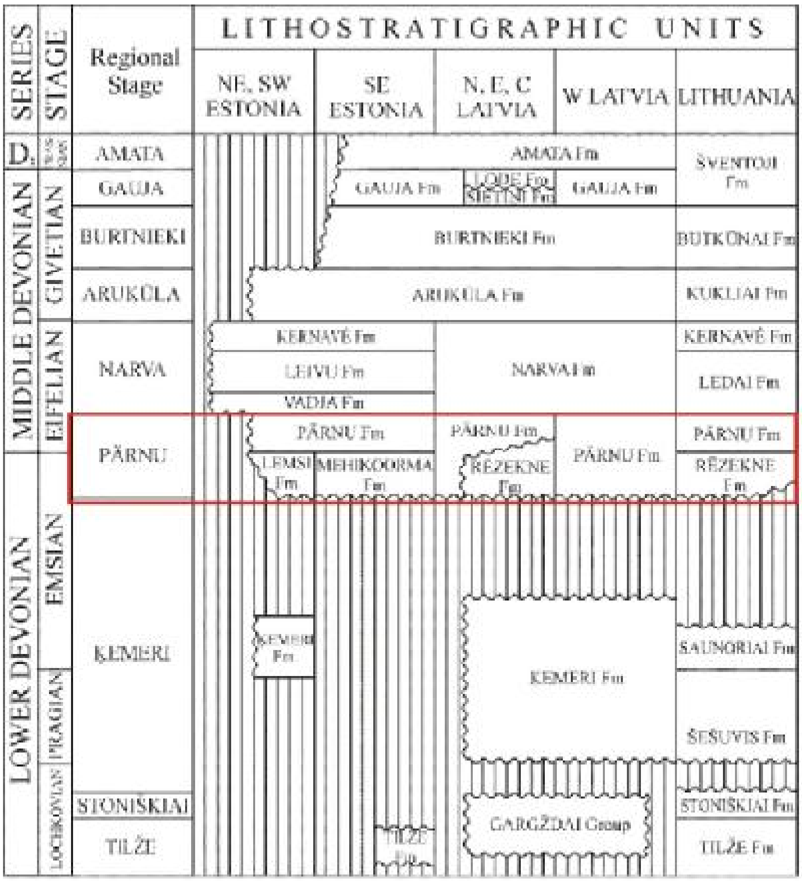

Figure 2. Stratigraphy of Lower to lowermost part of Upper Devonian in Baltic States (Lukševičs et. al., 2012). The studies succession is marked in red

2. attēls. Apakšējā līdz augšējā devona apakšdaļas slāņkopas stratigrāfiskais iedalījums Baltijas valstīs (Lukševičs et. al., 2012). Pētījumu intervāls iezīmēts ar sarkanu krāsu

The exposures of the deposits of the Pärnu RS form a small narrow wedge-shaped area in the NW part of its distribution area (see Figure 1.1.). The deposits crop out on the banks of the Pärnu River near Tori village in SW Estonia. In these outcrops Pärnu Fm is represented by coarse-grained deposits and provides an excellent opportunity for detailed analysis of facies and their lateral and vertical variations. For the rest of the territory the deposits are found only in drill-cores.

Approbation of results

The results of the research are presented in the following scientific papers:

1. Ivanov A., Lukševičs E., Stinkulis Ģ., **Tovmasyan K.**, Zupiņš I., Beznosov P., 2006. Stratigrafija devonskih otlozhenij Andomskoj gori (Stratigraphy of Devonian deposits in Andoma hill). *Problemi Geologiji i mineralogij*. Siktivkar, pp. 382-396 (in Russian).
2. Stinkulis Ģ., **Tovmasjana K.**, 2004. Rēzeknes un Pērnavas svītu nogulumi Baltijas austrumos (Deposits of Rēzekne and Pärnu Formations in Eastern Baltics). *Latvijas ģeoloģijas vēstis, 12,* Valsts ģeoloģijas dienests, Rīga, lpp. 6-13.

3. **Tovmasjana K.**, Stinkulis G., 2012. Klastisko un karbonātisko plūdmaiņu nogulumu sedimentācija Austrumlatvijā: vidusdevona Rēzeknes un Pērnavas svītas (Siliciclastic and carbonate tidal sedimentation: Middle Devonian Rēzekne and Pärnu Formations). *Latvijas Universitātes raksti. 789. sēj. Zemes un vides zinātnes.* Rīga, Latvijas Universitāte, lpp. 66-86.
4. **Tovmasjana K.**, Stinkulis G., Karušs J., Pipira D., Ostašovs M., Zupiņš I., 2011. Sandstones of the Devonian Sietiņi Formation and their sedimentary environment in the Bale-II sand pit. *In* Stinkulis Ģ., Zelčs V. (eds.), *The Eighth Baltic Stratigraphical Conference – Post-conference Field Excursion Guide*, Riga, pp. 33-36.
5. **Tovmasjana K.**, Stinkulis Ģ., Krakopa E., Zupiņš I., 2011. Sandstones, clayey deposits and dolocretes of the devonian Burtnieki Formation in the Veczemji cliff. *In* Stinkulis Ģ., Zelčs V. (eds.), *The Eighth Baltic Stratigraphical Conference – Post-conference Field Excursion Guide*, Riga, pp. 54-57.

Others:

1. **Tovmasjana K.**, 2003. Ģeoloģiskie dabas pieminekļi: Gūtmaņa ala. Sietiņiezis (Geological monuments: Gūtmaņa cave. Sietiņi cliff). *Latvijas Ģeoloģijas Vēstis, 11*, Riga, Valsts ģeoloģijas dienests, lpp. 20-23.

The results of the research are presented in the following international scientific conferences:

1. **Tovmasyan K.**, 2003. Sedimentary structures and origin of deposits of the Sietiņi Formation, Upper Devonian, Latvia. *22nd IAS Meeting of Sedimentology*, Opatija, Abstracts, p. 208.
2. **Tovmasyan K.**, 2004. Tidal signatures in deposits of Pärnu Formation (Middle Devonian) in Estonia. *Tidalites-2004, 6-th International Conference on Tidal Sedimentology*, Copenhagen, Abstracts, pp. 183-185.
3. Ivanov, A., Luksevics, E., Stinkulis, G., **Tovmasyan, K.**, Zupins, I., Zabele, A., 2005. Devonian Stratigraphy and Vertebrate Fauna of the Andoma Hill Section (Onega Lake, Russia). *In* Ivanov A., Young G. (eds.), *Middle Palaeozoic vertebrates of Laurussia: Relationships with Siberia, Kazakhstan, Asia and Gondvana. Ichthyolith Issues Special Publication 9*, pp 17-21.
4. Kreišmane D., **Tovmasjana K.**, Saks T., 2012. Sedimentary environments and evolution of the Middle Weicselian Basin, SE part of the Baltic Sea depression. *30th Nordic Geological Winter Meeting*. Reykjavik, Abstracts, p. 185.
5. Lukševičs E., Ivanov A., Stinkulis G., Zupiņš I., Beznosov P., **Tovmasyan K.**, 2002. *In* Satkūnas J., Lazauskiene J. (eds.), *Sedimentology and vertebrate fauna of the Gauja Regional Stage (Late Givetian) of the Main Devonian Field. 5-th Baltic Stratigraphical Conference,* Vilnius, Extended Abstracts, pp. 110-113.
6. Sevastjanova I., **Tovmasyan K.**, 2006. Composition and facies analysis of Middle Jurassic Sandstones, SW Latvia. *Sediment 2006, the 4th Annual Conference of the Central European Section of SEPM (Society for Sedimentary Geology)*, Gottingen, Abstracts, p. 154.
7. **Tovmasyan K.**, Stinkulis G., 2008. Implications of fluvial, tidal and wave processes to the deposition of siliciclastic succession of Devonian, Andoma hill, NW Russia. *In* Hints O., Ainsaar L., Männik P., Meidla T. (eds.), *7-th Baltic Stratigraphical Conference*. Tallinn, Abstracts, pp. 71.

The main results of the research are presented in the following local scientific conferences:

1. **Tovmasjana K.**, 2001. Gaujas reģionālā stāva smilšakmeņu tekstūras (Sedimentary structures of sandstones of Gauja Regional Stage). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 59. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 159-160.
2. **Tovmasjana K.**, 2002. Devona Sietiņu svītas smilšakmeņu tekstūras Bāles karjerā (Sedimentary structures of sandstones of Devonian Sietiņi Formation in Bāle sand pit). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 60. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 77-78.
3. **Tovmasjana K.**, 2003. Devona Sietiņu svītas nogulumu fācijas (The facies analysis of deposits of Upper Devonian Sietiņi Formation). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 61. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 194-195.
4. **Tovmasjana K.**, 2004. Plūdmaiņu pazīmes Pērnavas svītas nogulumos Igaunijā (Tidal signatures in deposits of Pärnu Formation in Estonia). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 62. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 181-183.
5. Lukševičs E., Stinkulis Ģ., **Tovmasjana K.**, Zupiņš I., 2005. Andomas kalna (Krievija, Oņegas ezera DA krasts) ģeoloģiskā uzbūve (The geological structure of Andoma hill, SE coast of Onega lake, Russia). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 63. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 130-132.
6. Sevastjanova I., **Tovmasjana K.**, 2006. Latvijas juras Papiles svītas smilšaino nogulumu sastāvs un to faciālā analīze (Composition and facies analysis of sandstones of Papile Formation, Jurassic, Latvia). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 64. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 202-204.
7. Stinkulis Ģ., **Tovmasjana K.**, Lukševičs E., Zabele A., 2005. Devona nogulumu sedimentācijas apstākļi Andomas kalna apkārtnē (Depositional environment of the Devonian succession, Andoma Hill, SE coast of Onega lake, Russia). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 63. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 155-156.
8. **Tovmasjana K.**, Plinka-Bjorklunde P., 2010. Estuāru un deltu nogulumu faciālās atšķiribas un to salīdzinājums griezumā griezumā (The main differences in the facies of tide-dominated estuaries and tide-dominated deltas and their comparison in rock record). *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 68. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, lpp. 374-376.
9. **Tovmasyan K.**, Stinkulis Ģ., 2008. Facies analysis of tidal and wave-influenced siliciclastic shallow marine succession of Devonian, Andoma hill, NW Russia. *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 66. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, pp. 236-238.
10. **Tovmasyan K.**, Stinkulis Ģ., Plink-Björklund P., 2008. Depositional environment in the tidally-influenced transgressive succession, Pärnu Regional Stage, Baltic Devonian Basin. *Ģeogrāfija. Ģeoloģija. Vides zinātne. Latvijas Universitātes 66. zinātniskā konference. Referātu tēzes*. Latvijas Universitāte, Rīga, pp. 238-239.
11. **Tovmasyan K.**, Stinkulis Ģ., Plink-Björklund P., 2008. Facies architecture and deposition in estuarine fluvial-tidal transition zone: example from Middle Devonian Pärnu Regional Stage, SW Estonia. *Ģeogrāfija. Ģeoloģija. Vides*

zinātne. Latvijas Universitātes 66. zinātniskā konference. Referātu tēzes. Latvijas Universitāte, Rīga, pp. 240-241.

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1. Study area and geological background

The BDB developed in the western part of the East European Platform (EEP) on the Euramerica (*Baltica* until the Middle of Silurian) plate. From Ediacaran times up to the Late Ordovician, the Baltica was a separate plate that drifted from high to moderate latitudes on the southern hemisphere towards the equator (Cocks, Torsvick, 2005). The equatorial position of Baltica was reached by Silurian-Devonian times (Cocks, Torsvick, 2005; 2006). Following the Ediacaran to Early Cambrian opening of the Tornquist and Iapetus oceans and the break-up of the Rodinia supercontinent, the Baltica began to converge during Middle and Late Cambrian times with Avalonia as well as the Laurentia-Greenland plates, to which it was finally sutured during Late Ordovician and Silurian times, respectively, along the Trans-European Suture Zone and the Arctic-North Atlantic Caledonides (Nikishin et al., 1996; Torsvik, Rehnstrom, 2003; Roberts, 2003; Cocks, Torsvick, 2006). The BDB has been "squeezed" in between two collision belts: the Scandinavian Caledonides (closure of Iapetus Ocean) and the German-Polish Caledonides (closure of Tornquist Ocean) (Figure 1.1.).

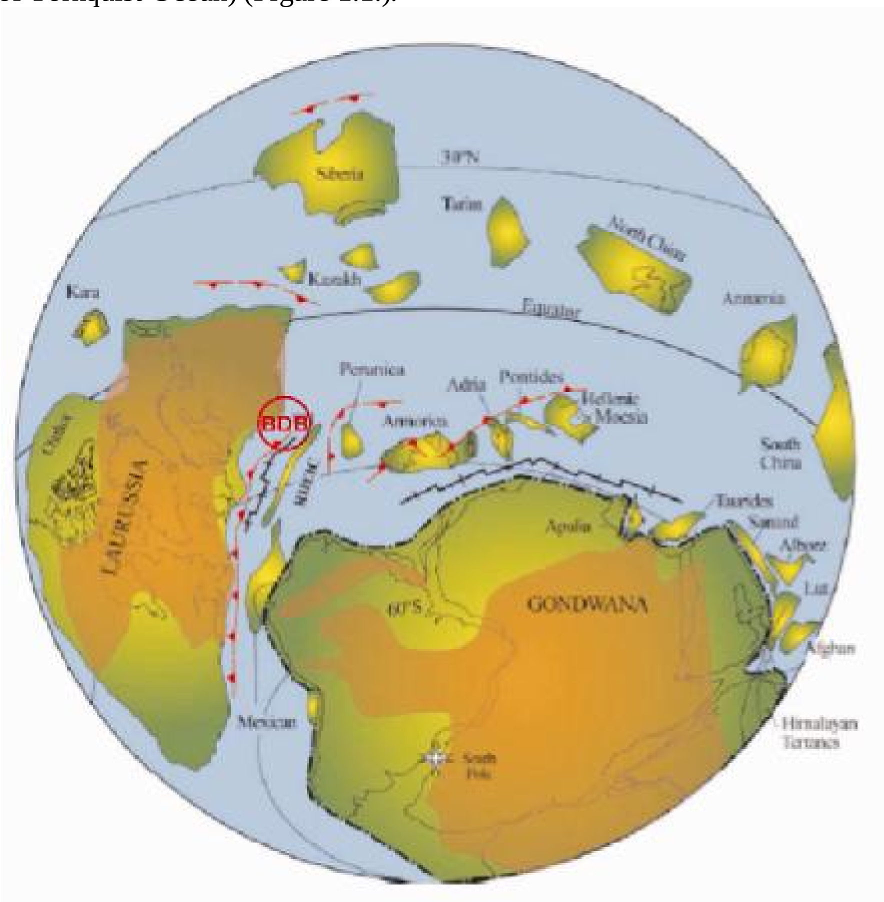


Figure 1.1. Palaeogeography of the Early Devonian (Cocks, Torsvik, 2006)
Baltic Devonian Basin (BDB) is situated in between the collision belts related to closure of Iapetus and Tornquist oceans

1.1. attēls. Agrā devona paleoģeogrāfija (Cocks, Torsvik, 2006)
Baltijas devona baseins (BDB) atradās starp divām kolīzijas joslām, kas izveidojās aizveroties Iapetusa un Torkvista okeāniem

Preceding the BDB, in Early and Middle Ordovician, the western part of the EEP, as far as the Moscow Basin was slowly subsiding and covered by shallow, epicontinental sea with extremely low sedimentation rate. The basin was an evenly and weakly tilted ramp where carbonate and argillaceous sediments accumulated. Pulsatory input of the fine terrigenous material, probably from the Iapetus margin of the Baltica, started in the early Caradoc. During the middle Caradoc the sedimentary succession was interbedded with several volcanic ash layers, derived from the northwest - Iapetus Ocean (Cocks, Torsvick, 2005).

At the end of the Ordovician, since the late Caradoc and especially during the Silurian, the upheaval of the northwestern part of the craton occurred (mid Wenlock), caused by the closing of the Iapetus ocean. At the same time on the southwestern margin of the craton, the subsidence of the basin floor intensified, influenced by overthrusting and subduction of the Avalon plate (related to the closure of Tornquist Ocean). As a result a relatively narrow, deep, “starved”, basin formed where hemipelagic argillaceous deposits accumulated (Plink-Björklund, Björklund, 1999). The sea gradually retreated from the northwest and central parts of the EEP, and the basin evolved from an epicontinental sea into a pericontinental gulf. The Late Ordovician and earliest Silurian is the time of the tectonic inversion in the western part of the EEP, characterised by periods of low influx of siliciclastic mud alternated with intensive supply of a fine clastic material (late Caradoc - middle Llandovery). The early - middle Llandovery ended with a regional denudation in the western Estonia and eastern Lithuania, probably caused by the beginning of the collision between the Laurentia and Baltica plates. During the late Llandovery - early Ludlow influx of the fine clastics started to infill the “starved” depression (Plink-Björklund, Björklund, 1999). During the late Llandovery - early Wenlock the sedimentary succession was interbedded by multiple volcanic ash layers, derived from Scandinavian Caledonides. During the late Ludlow - Pridoli (up to the Til ∩e Age) intense influx of siliciclastic material from the Scandinavian Caledonides filled the basin depression and diluted carbonate sedimentation (Plink-Björklund, Björklund, 1999). By the beginning of the Devonian, only a remnant lagoon was preserved in NW Latvia and south-western Lithuania. During the Early Devonian, the northern part of the BDB was uplifted and most of the basin experienced subaerial erosion (Kurshs, 1992; Kleesment, 1997). The uplift has been attributed to propagation of stress from the Baltica to Laurentia-Greenland plate collision (Plink-Björklund et al., 2004).

The Early and Middle Devonian sedimentary succession in the BDB is strikingly different from the described Ordovician to Silurian, as the accumulation of carbonate and clayey material changed into deposition of sandy to conglomeratic siliciclastics (Kurshs, 1992; Kleesment, 1997; Plink-Björklund, Björklund, 1999). A phase of coarse siliciclastic deposition occurred in the BDB from the Early Devonian until the beginning of the Late Devonian (Frasnian). Such an appearance of coarse clastics in a basin fill is noted typically as the time of tectonically rejuvenated relative source-area uplift. At the end of the Early Devonian and during the Emsian, the basin started to subside again (Plink-Björklund, Björklund, 1999).

The interpreted maturity development (see Kurshs, 1975) together with the suggested sediment input from northwest suggest that the Middle to lowermost Upper Devonian deposits in Estonia and Latvia are the product of a cannibalisation of the Scandinavian Caledonian foreland basin (Plink-Björklund, Björklund, 1999; Figure 1.2.). In the Early Devonian, Estonia and part of Latvia were an erosion and bypass area, while contemporaneous siliciclastic deposition occurred in southern portions of the Baltic Basin. A forebulge axis migrating with time may have constrained sediment transport southward along the axis of the foredeep (Plink-Björklund, Björklund, 1999). During the Emsian time extensional collapse and uplift occurred in the Scandinavian Caledonides. The uplift exposed the foredeep sediments to erosion and caused southeast transport across the foreland to Estonia and Latvia (Plink-Björklund, Björklund, 1999). This redeposition continued until the Baltic Basin was filled and the siliciclastic depositon replaced by evaporite deposition during the Late Devonian.

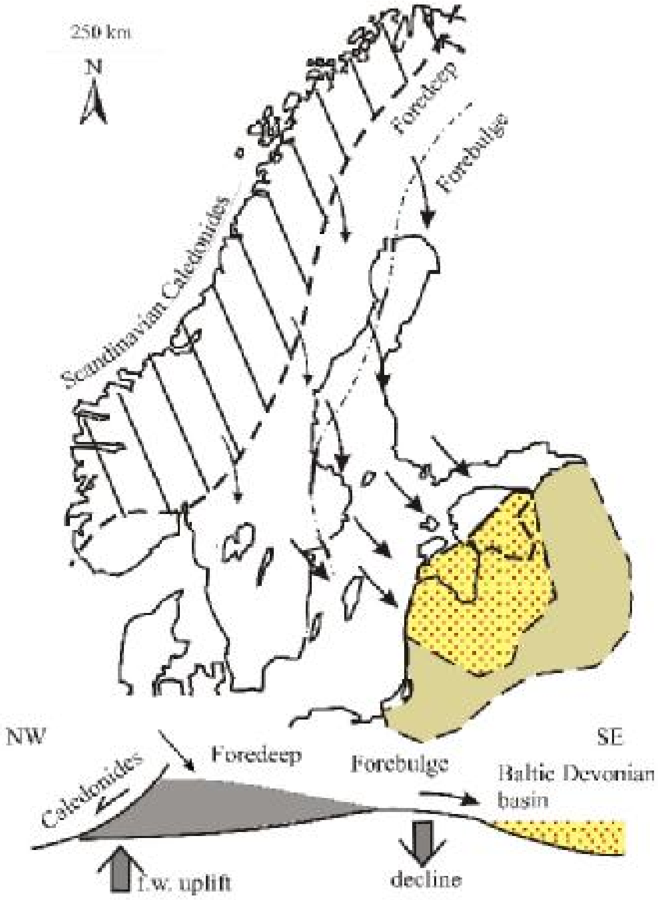


Figure 1.2. Scandinavian Caledonides and the Baltic Devonian Basin (BDB)
The dotted area illustrates the Baltic Devonian basin, while the homogeneous colour represents the adjacent to BDB epicontinental sea territory. Note, that the boundaries of the basins are schematically illustrated (unpublished data by Plink-Björklund, Björklund, 2000; data from Kurshs (1992) used)
1.2. attēls. Skandināvijas Kaledonīdu kalni un Baltijas devona baseins
Punktētais laukums apzīmē Baltijas devona baseinu, viendabīgās krāsas iecirknis –epikontinentālā baseina teritorijas daļu. Baseinu robežas ir ilustrētas shematiski, uz ko norāda raustītās līnijas (pēc Plinkas-Bjorklundes un Bjorklunda, 2000, nepublicēts materiāls, kā arī pēc V. Kurša, 1992)

The Rēzekne RS deposits are spread in southern Estonia and are covered by the Pärnu RS (Kleesment, Mark-Kurik, 1997). The succession overlies erosively Ordovician and Silurian deposits of different age. The total thickness of the stage varies from 0,7 to 51,5 m, and it is thickest in eastern Estonia (Kleesment, Mark-Kurik, 1997). It is characterized by sandstone mainly and dolomitic marl in the upper part of the section in SE Estonia. The lower part of the section consists of sandstone, often conglomeratic, with dolomitic matrix (Kleesment, Mark-Kurik, 1997; Figure 1.3.).

During a transgression at the Rēzekne time, the epicontinental see flooded a great part of the EEP (Kleesment, 1997). The northen part of the basin extended to Estonian territory and sandy sediments of nearshore shallow sea formed there, while in SE Estonia sandy-silty sediments deposited (Kleesment, 1997).

The present-day distribution of the Pärnu RS comprises the southern part of Estonia, Latvia, mid-Lithuania, Northern Belarus in the south and NW Russia to the east (see Figure 1B). The deposits of the Pärnu RS are distributed in wider areas than more ancient Devonian deposits and spread in the whole territory of Latvia, western, northern and central Lithuania, southern Estonia, as well as northern Belarus and further to the east. The total thickness of these formations increases up to 70 m in the Gulbene deep, but in other regions it usually does not exceed 20-40 m. The rocks crop out only along the Pärnu River within 6 km small area in SW Estonia (see Figure 1C). It unconformably overlies the Ordovician and Silurian carbonate rocks, mainly in the eastern part of its distribution area, or the lower Devonian siliciclastic deposits. Pärnu RS forms a lower part of a 500 m thick dominantly siliciclastic Middle Devonian (Kleesment, Mark-Kurik, 1997; Stinkulis et al., 2011), probably Middle to lowermost Upper Devonian (Brangulis et al., 1988) succession. It is overlain by a shallow-marine transgressive succession (Plink-Björklund, Björklund, 1999; Tänavsuu-Milkeviciene et al., 2009), followed by a thick tide-dominated (Tänavsuu-Milkeviciene, Plink-Björklund, 2009) to tide-influenced deltaic system (Pontén, Plink-Björklund, 2007) and transgressive estuarine succession at the top of the succession (Pontén, Plink-Björklund, 2009).

During the Pärnu time (Eifelian), deposition of coarse to fine grained siliciclastics prevailed with minor role of carbonate deposition in the eastern part of the basin. An eastward expansion of the BDB to the Moscow Basin (Narbutas, 1984; Kurshs, 1992) has been interpreted to have occurred in the beginning of Eifelian, in Pärnu time, for the first time in the Devonian it penetrated eastwards into the Moscow syneclise. Open sea, like before, was situated west of the epicontinental basin and was partly separated from it by the northwest-southeast barrier – northern part of the Belarus-Mazurian anteclise (Kurshs, 1975). Strong river influx from north lowered the salinity of water and provided large amount of sandy clastics, therefore the section in northern part of eastern Baltics is predominated by sandstones (Kurshs, 1992; Kleesment, 1997). According to Kleesment (2009), an increased roundness of quartz grain in the belt from Kihnu to Värska marks a possible position of the ancient coastline. Freshwater influx did not reach regularly the eastern parts of the basin, therefore carbonate deposits prevail in this part of the basin. Further to south-east in the territory of Belarus deposits contain less siliciclastic admixture, and are represented by the clayey dolostones (Kurshs, 1975). The gypsum lenses and intercalations often occur in the clayey carbonate deposits. Eastwards from Latvia in the upper part of the sequence, gypsum layers appear and thicken to southeast, but further to east, in Moscow Basin, even beds of halite occur (Tikhomirov, 1995).

The subdivision of the Pärnu succession into formations is unclear until today and is based on a combination of palaeontological, mineralogical and lithological data (Kleesment, 1997; PaΓkevičius, 1997; Kurshs, Stinkulis, 1998; Savvaitova, 2002; Mark-Kurik, Põldvere 2012; LukΓevičs et al., 2012; see Figure 2). The main opinion prevails on subdivision of two regional stages: Rēzekne RS (upper part of Emsian) and Pärnu RS corresponding to the lower part of Eifelian (Kleesment, Mark-Kurik, 1997; PaΓkevičius, 1997). These regional stages can be distinguished from each other by lithology and cyclic pattern in eastern part of the south-east Baltics and mineral composition in south-western Estonia and north-western Latvia (Kurshs, 1992; Kleesment, Mark-Kurik, 1997; Kurshs, Stinkulis, 1998). In other areas due to predominance of siliciclastics and poor palaeontological characteristics these regional stages are nearly impossible to subdivide, therefore they are mentioned as the Rēzekne-Pärnu undivided RS (Kurshs, Stinkulis, 1998).

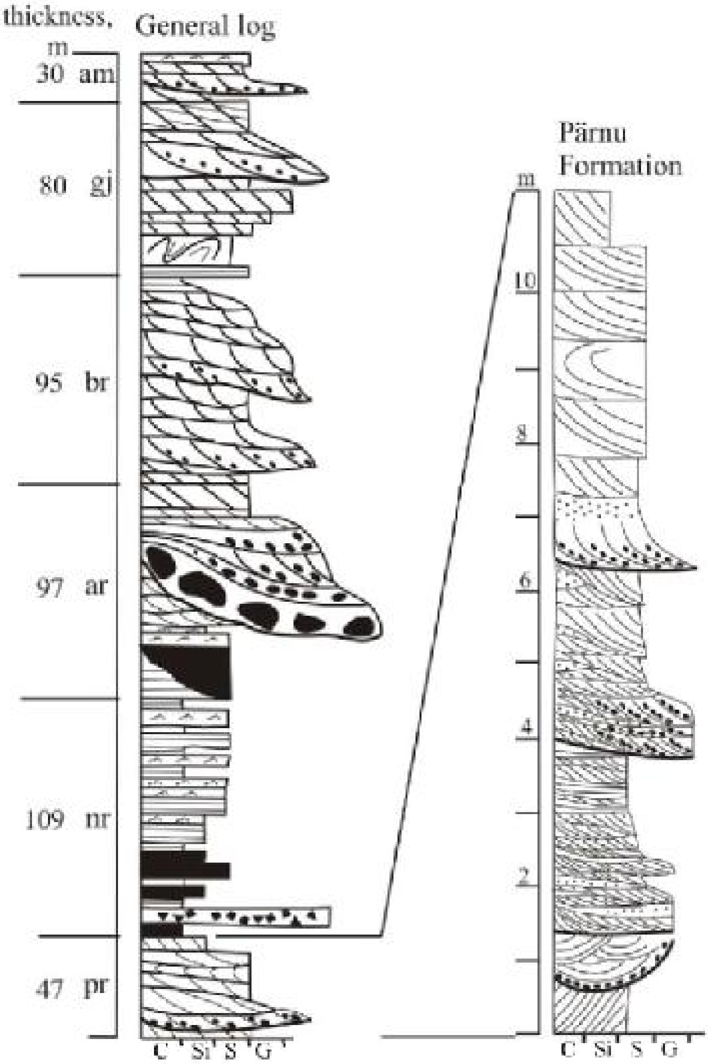


Figure 1.3. Generalized sedimentary log of the Middle to lower Upper Devonian succession in Baltic States with a generalized sedimentary log of Pärnu Formation at the outcrop area

For the horizontal scale: C – claystone, Si – siltstone, S – sandstone, G – gravel; for the vertical column: pr – Pärnu Regional Stage, nr – Narva Regional Stage, ar – Aruküla Regional Stage, br – Burtnieki Regional Stage, gj – Gauja Regional Stage, am – Amata Regional Stage (Plink-Björklund, unpublished data)

1.3. attēls. Vienkāršots vidus- un augšdevona apakšdaļas slāņkopas griezums Baltijas valstīs un vienkāršots Pērnavas svītas nogulumu griezums atsegumu teritorijā

Pie horizontālā mēroga: C – māls, Si – aleirolīts, S – smilšakmens, G – grants; Pie vertikālās kolonnas: pr – Pērnavas reģionālais stāvs, nr – Narvas reģionālais stāvs, ar – Arukilas reģionālais stāvs, br – Burtnieku reģionālais stāvs, gj – Gaujas reģionālais stāvs, am – Amatas reģionālais stāvs (Plinka-Bjorklunde, nepublicēts materiāls)

The same problems complicate the subdivision of lithostratigraphical units corresponding to the Rēzekne and Pärnu RS. The Rēzekne Fm corresponds to the Rēzekne RS in the eastern part of the south-east Baltic States, but the Lemsi Fm is attributed to the same RS in the south-western Estonia (Paškevičius 1997; Kleesment, Mark-Kurik 1997).

The Pärnu Fm is attributed to the Pärnu RS in all the distribution area of these deposits (Paškevičius, 1997) or to the Pärnu RS in eastern and northern part of the south-east Baltics and to both Rēzekne and Pärnu RS (in original work defined in the same stratigraphic volume as the Pärnu RS) in central and western parts of the south-east Baltics, where separation of two formations is problematic (Kurshs 1975; Gailīte et al., 2000; Lukševičs et al, 2012).

According to Lyarskaja (1974; 1981), the Rēzekne Fm and its analogues by geological age are not found in the western part of the Baltics. According to this opinion, the deposits of the Rēzekne RS (upper part of the Emsian) are not found in the above-mentioned area. However, the Lemsi Fm (age-equivalent of the Rēzekne Fm) has been identified by mineralogical data in the Lūžņi drill-core, and two cycles composed by sandy deposits in lower parts and clayey-carbonate deposits in upper parts resembling the Rēzekne and Pärnu fms have been documented in the Talsi-55 and Gārzde-1 drill-cores (Kurshs, 1975). All these drill-cores are located in the north-western Latvia. According to Kurshs (1992) the age-equivalents of the Rēzekne Fm and Lemsi Fm are distributed also outside the above-mentioned eastern and north-western parts of the south-east Baltics, but due to its composition it's not possible to distinguish two formations. Thus, it was proposed to use the name Pärnu Fm for the siliciclastic beds of both Rēzekne and Pärnu RS in the areas where more detailed subdivision is problematic and for the deposits of only Pärnu RS where it's possible to divide the Rēzekne or Lemsi Fm in the Rēzekne RS (Kurshs, 1975; 1992; Gailīte et al., 2000; Savvaitova, 2002; Lukševičs et al, 2012).

Moreover, the Pärnu Fm in Estonia is further subdivided into a lower, coarser-grained Tori Member and an upper, fine-grained Tamme Member (Kleesment, Mark-Kurik, 1997). This subdivision is done by lithological data: the Tori Member consists of light-grey, loose, mainly cross-stratified coarse to medium-grained sandstone that becomes finer upward, while the Tamme Member consists of fine and very-fine grained, mainly cemented sandstone that contains thin interlayers of siltstone and mudstone, and is dominantly plane-parallel stratified with sandy dolostone at the topmost part of the Pärnu succession (Kleesment, Mark-Kurik, 1997).

In this study, the Rēzekne and Pärnu units are considered together as one single succession representing the basal part of a transgressive cycle in the Middle Devonian and thus are studied together. Moreover, the entire succession is subdivided into several retrogradational stratigraphic units based on the detailed facies analysis (see Appendixes). Due to uncertain formation boundaries, described above, the correlation between the core sections is based on lateral and vertical changes of facies. It must be noted that this may seem somewhat controversial to the stratigraphic subdivision and can be argued, however, in numerous places, the formation boundaries follow changes in depositional environments and sediment input, and thus are diachronous, instead of representing actual timelines.

2. Earlier studies of the Middle to lower Upper Devonian siliciclastic succession in the Baltic States

The origin of the Devonian siliciclastics in the Baltic States has been re-interpreted several times. In the very 19-th century an opinion on the aeolian origin of Devonian siliciclastics prevailed. Later, alluvial cycles have been distinguished in the siliciclastic succession of the Devonian by Poļivko (1977). After that, main points of the discussion were related to dominance of marine versus continental sedimentary environment during the Early Devonian - beginning of Late Devonian as a whole, as well as different episodes of this time. Abundant cross-stratification structures, as well as fining-upwards cycles composed of sandstones in lower parts and clayey deposits in upper parts (interpreted as alluvial cycles by Poļivko, 1977) have been used as the main arguments to attribute the siliciclastic deposits to alluvial environments. Several parts of the Devonian sequence such as the Ķemeri, Aruküla and Burtnieki formations sometimes have been interpreted as the lacustrine deposits. However the major works by V. Kurshs give evidence about dominant role of the marine processes during sedimentation of the whole Devonian siliciclastic sequence (Kurshs, 1975; 1992). He suggested that the Devonian siliciclastic deposits in the present area of Baltic States most possibly were accumulated in a wide and shallow epicontinental sea (Kurshs, 1975, 1992). However, he did not exclude that alluvial settings dominated along the northern margin of the basin, which is largely removed due to the post-Devonian erosion processes. According to Kurshs (1992), the presence of deltaic zone in the northern part of present distribution area of the Devonian deposits points to the existence of continental settings further to the north.

A great contribution to the study of the Middle to early Late Devonian BDB was done in the timeframe from 1999 to 2003 by P. Plink-Björklund and L. Björklund and an international team of specialists and students within the framework of the project "Tectonic and sedimentologic history of the Devonian Baltic Basin: Relation to development of Scandinavian Caledonides and German-Polish Caledonides". P. Plink-Björklund and L. Björklund described first sedimentology and sequence stratigraphy of the BDB in the northern Baltics, and the interaction of sediment input from the Scandinavian Caledonides and the Caledonian foredeep (Plink-Björklund, Björklund, 1999). Later the studies also focused on sedimentological and sequence stratigraphic correlation across the whole BDB, with incorporation of basin analyses, reconstruction of palaeoenvironments in various parts of the Middle-lower Upper Devonian succession in Baltic States. According to these studies, the origin of deposits of some intervals of the Middle to lower Upper Devonian has been re-interpreted. Thus, the upper part of the Narva RS has been interpreted as a progradational, siliciclastic-rich deltaic system (Tänavsuu-Milkeviciene et al., 2009), Aruküla RS is interpreted as tide-dominated delta (Tänavsuu-Milkeviciene, Plink-Björklund, 2009), Gauja RS as a tide-influenced delta plain and delta front (Pontén, Plink-Björklund, 2007), and more recently the Amata Fm has been interpreted as tide-dominated estuarine system (Pontén, Plink-Björklund, 2009).

Few works in relation to the Rēzekne and Pärnu RS exist focusing mainly on lithological, mineralogical, palaeontological and stratigraphical issues (Kalamees, 1988; Kurshs, 1975, 1992; Kleesment, Mark-Kurik, 2007; Kleesment, 2007; Kleesment, 2009). However no detailed facies analysis exist on these deposits. V. Kurshs describes the lithology of the Rēzekne and Pärnu successions in Baltics, divides major "facies zones" (Kurshs, 1975) and provides an insight into prevailing palaeoenvironments and palaeogeographical reconstructions of the basin (Kurshs, 1975; 1992). A. Kleesment describes the lithology and mineralogy of these deposits in Estonia (Kleesment, Mark-

Kurik, 1997; Kleesment, 2009) and interprets them as near-shore shallow sea (Kleesment, 1997). In her most recent study, A. Kleesment (Kleesment, 2009) gives data on sand grain roundness and corrosion, and suggests that a line from Kihnu to Värska marks a position of the palaeocoastline in Pärnu time.

The most detailed recent facies studies have been made by G. Stinkulis (1998) in the framework of his dissertation project. He studied in details the mixed carbonate-siliciclastic facies of the Pärnu RS (Rēzekne and Pärnu Fm) in eastern parts of Latvia based on detailed lithological and mineralogical composition studies of deposits in four drill-cores. The study allows to precise the main facies assemblages that formed in the eastern part of the Pärnu basin in Baltics. According to Stinkulis (1998) a special role during this time in the eastern parts of Pärnu basin in Baltics was played by synsedimentary positive tectonic structure - Viļaka arch, that created very shallow settings and partly hindered strong currents, which carried large amount of siliciclastic material across the central part of basin. The description of typical feature of these sediments – the dolomite ooids and peloids, is given in detail. Internal fabric of ooids, including concentric laminae of sulphide minerals, suggests that these grains formed under changeable low and high agitation (Stinkulis, 1998). Mixed-composition clayey carbonate and sandy deposits exhibit wave ripples and whirl-like structures interpreted to be formed in strong winds and possibly even storms. The dominant type of sediments - dolomitic marls and sandy dolomitic marls with inclusions of gypsum, are interpreted to be accumulated in lagoonal settings. According to more recent studies, dolocretes that occur in many intervals of the Devonian succession, suggest to indicate frequent episodes of subaerial exposure during the Devonian basin development (Stinkulis, Spuņeniece, 2011).

3. The role of tides to the clastic deposition and tidal signatures in siliciclastic sediments

Origin of tides

A tide is any periodic fluctuation in water level that is generated by the gravitational attraction of the Moon and Sun. The Moon, because it is closer to the Earth, exerts a tide-generating force which is twice as large as that of a Sun. Tides represent the vector sum of two forces: 1) Gravitational attraction of the Moon, which is strongest on the side of the Earth facing the Moon, and 2) the centripetal force caused by the revolution of the Earth-Moon system about the common centre of mass. On the side of the Earth facing the Moon, the gravitational attraction of the Moon is greater than the oppositely directed centripetal force, while the reverse is true on the other side of the Earth. The water in the oceans therefore piles up in two bulges, one underneath the Moon, and the other on the opposite side of the Earth (Dalrymple, 1992). Because the Earth rotates, the bulges travel around the Earth as two tidal waves causing water levels to rise and fall regularly. Rising water levels are known as flood tides, whereas the fall is the ebb tide. The most commonly observed period for one complete tidal cycle is the semidiurnal (twice a day) period of 12.42 hours. Because the rotational axis of the Earth is inclined with respect to the orbital plane of the Moon most of the time, any given point on the Earth surface passes closer to the crest of one tidal bulge than the other, thereby adding a diurnal (once a day) component to the tidal spectrum (Dalrymple, 1992).

The interaction of the Moon and Sun produces a still longer periodic variation in tidal range – the difference in water level between successive high-tide and low-tide levels. When the Sun and Moon lie in a straight line relative to the Earth, their effects add to produce greater than average tidal ranges – spring tides. When the Sun and Moon are at right angles, their forces counteract each other and the tidal range is smaller than average – forming the neap tide. For semidiurnal tides, the spring-neap cycle has a period of 14.77 days, and contains 28 tidal cycles. Diurnal tides have a neap-spring period of 13.66 days that contains 14 tidal cycles (Dalrymple, 1992; Kvale, 2012)

Many areas experience a net or residual transport of sediment in the direction of the stronger (dominant) current. The weaker current which flows in the opposite direction is termed the subordinate current.

Tidal range

Because the tide-generating forces are small, only the open oceans develop significant tides, and even these typically have a range of less than 1 m. Smaller bodies of water (including enclosed seas) cannot develop an appreciable tide of their own. The tides which are observed on continental shelves are due to the forcing action of the oceanic tide. As the tidal wave moves from the open ocean onto a continental shelf, shoaling and convergence associated with coastal embayments concentrate the energy within the tidal wave into a smaller cross sectional area, and the tidal range increases (Dalrymple, 1992).

Tidal dominance over other processes is most common in areas where the tidal range is large, because this results in strong tidal currents. Consequently, most microtidal (0-2 m tidal range) and mesotidal (2-4 m tidal range) areas are wave (or storm) dominated, whereas some mesotidal and most macrotidal (>4 m tidal range) areas are tide-dominated (Johnson, Baldwin, 1996). However, if wave action is limited due to topographic sheltering, or the tidal current speeds are increasing by a topographic construction, tidal dominance can even occur in microtidal areas (Dalrymple, 1992).

Main tidal sedimentary structures

The periodic changes in the current speed and direction which characterize tidal systems produce several sedimentary structures which are diagnostic of tidal deposition.

Cross bedding contains some type of regularly spaced, internal discontinuities formed in areas where the maximum speed of the dominant current is capable of producing dunes. If the subordinate current is capable of eroding the lee face of the dunes formed by the preceding dominant current, ***reactivation surfaces*** are produced (Figure 3.1.).

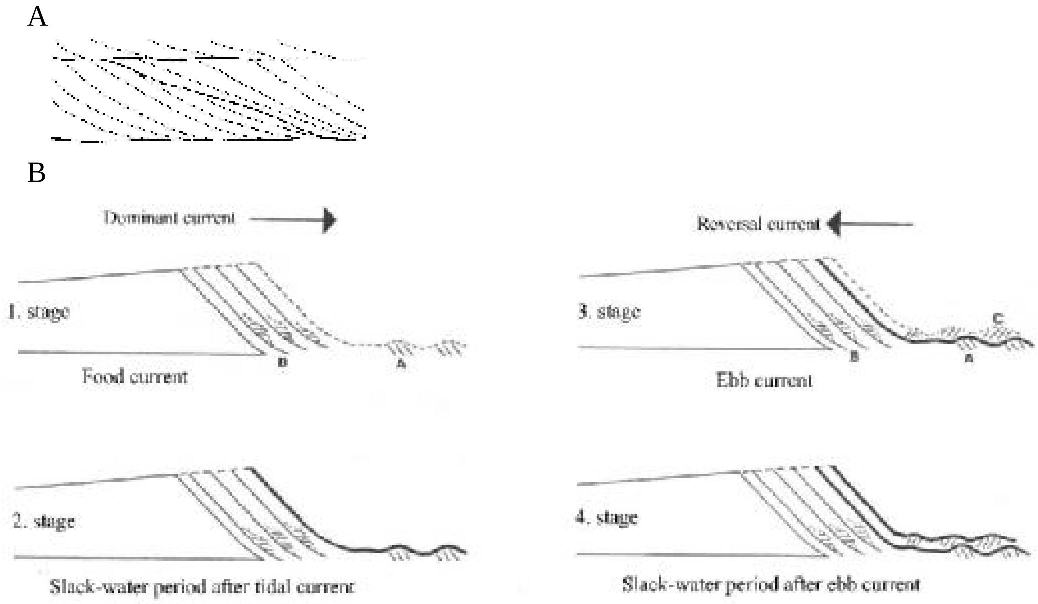


Figure 3.1. A - reactivation surfaces, B – formation of reactivation surfaces in cross-strata (Nichols, 1999)
3.1. attēls. A – reaktivācijas virsmas, B – reaktivācijas virsmu veidošanās slīpslāņotās sērijās (Nichols, 1999)

Mud drapes may also be deposited on the lee face during one or both slack-water periods if suspended sediment concentrations are high enough. Typically the amount of sand deposited by the subordinate current is small, and hence the mud drapes deposited after the dominant and subordinate tides are closely spaced (Figure 3.2).

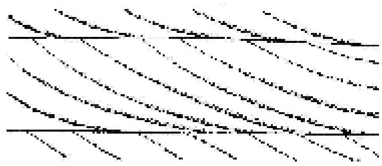


Figure 3.2. Mud drapes in cross-strata (Nichols, 1999)
3.2. attēls. Māla kārtiņas uz slīpajiem slānīšiem (Nichols, 1999)

The deposits of a single, dominant tide are called ***tidal bundles***, whether bounded by reactivation surfaces or mud drapes. Because of the variation in tidal current speed associated with the neap-spring cycles, sequences of tidal bundles commonly show cyclic variations in thickness, with thicker bundles forming during spring tides when currents are stronger and the dune migrates further (Figure 3.3.).

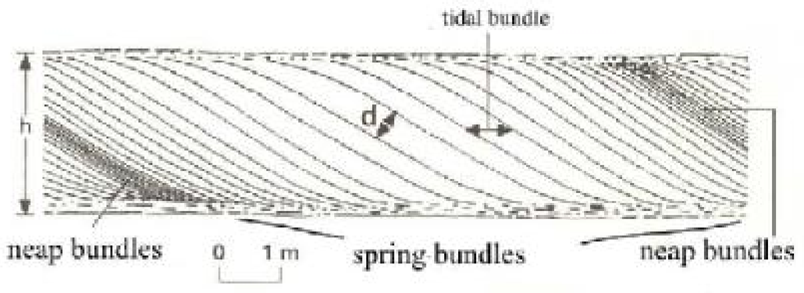


Figure 3.3. Illustration of tidal bundles and neap-spring cycles in cross-strata (Nichols, 1999)
3.3. attēls. Plūdmaiņu kārtas un plūdmaiņu cikli slīpslāņotās sērijās (Nichols 1999)

Herringbone cross stratification which is marked by opposed (bipolar) dip directions in adjacent sets of cross-bedding, is widely used as an indication of tidal deposition. However, bipolar cross-bedding is not universally developed in tidal settings, because the development of either flood or ebb dominance commonly produces a unidirectional palaeocurrent pattern.

Flaser, wavy and lenticular bedding reflect a unit of ripples formed by the maximum currents of the dominant tide, followed by deposition of a mud layer during the ensuing slack-water period (Figure 3.4). If the subordinate current is strong enough, a second rippled unit with an opposite palaeocurrent direction will be deposited on top of the first mud drape, followed by another mud drape during the second slack-water interval. Such beds, which show cyclic changes in layer thickness due to neap-spring variations, in tidal current speed are called ***tidal rhythmites*** (Kvale, Archer, 1990; Dalrymple et al., 1991).

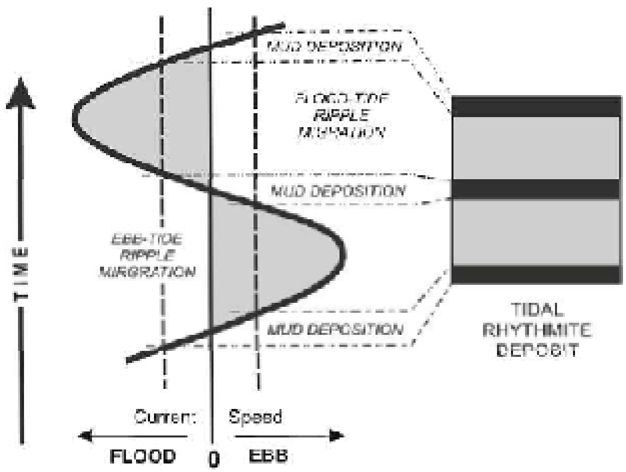


Figure 3.4. Formation of a tidal rhythmite in tidal current (Dalrymple, 1992)
3.4. attēls. Plūdmaiņu ritmīta veidoľanās plūdmaiņu straumēs (Dalrymple, 1992)

4. Materials and methods

This study is based on the detailed field data analyses of the Pärnu Fm that is exposed in a wedge-shaped outcrop area, ca. 6 km long, oriented NE to SW along the Pärnu River, as well as from drill-cores comprising the Rēzekne and Pärnu sedimentary succession in the Baltic States (see Figure 1). The outcrops are from 2 to 10 m high and usually 30-50 m long. The largest outcrop, Nr 1, stretches continuously for more than 350 m. The exposures are long and high enough for studying architectural elements of deposits and determining their orientation. Twenty two vertical lithostratigraphic logs were measured at seven outcrops at the scale of 1:25. Major bounding surfaces and stratigraphic packages were documented on photomosaics in order to reconstruct the internal architecture of sandbodies. These high resolution photomosaics have been used to map changes of sedimentary structures, main bedding surfaces, bedding geometry, as well as to reconstruct palaeocurrent orientation for each cross-section and for the entire outcrop belt, both laterally and vertically, and to reveal their combinations.

Palaeocurrent measurements (n=481) were primarily derived from cross strata, of which dip azimuths rose diagrams were compiled. Dip directions of master surfaces, such as channel scour surfaces, main bedding surfaces, as well as smaller scale bedding surfaces and reactivation surfaces, ripple cross-lamination, were also measured in order to identify the local palaeocurrent directions in different parts of the palaeochannels and bars. Separate rose diagrams were created to reflect both the regional and local variations of the palaeocurrent directions, as well as the variations through the measured sections of individual depositional units.

The preserved thicknesses of cross-stratified bedsets were recorded in order to calculate the subaqueous dune heights and subsequently the channel depths within the palaochannel thalwegs, using equations of Leclair and Bridge (2001) and Dalrymple and Rhodes (1995), (for details see Chapter 5.1.2).

Water depth for the facies has been calculated from cross-set thicknes according to Leclare and Bridge (2001):

Mean dune height is approximately 2.9(+-0.7)X mean cross set thickness:

$$Hs=2.9(+-0.7)X\ Sm,$$

Where *hm* is the mean dune height and *sm* is the mean cross-set thickness.

According to Leclare and Bridge (2001), the dune height fall within the range of:

$$3<d/hm<20$$

Where *d* is the water depth and *d/hm* is usually between 6-10:

$$6<d/hm<10$$

Water depth was also calculated according to Dalrymple and Rhodes (1995), where dune height (Hd) should be approximately 17% of the water depth:

$$Hd=0.167h$$

Trough cross-stratified mudstone conglomerate (facies 2), trough cross-stratified sandstone (facies 3) and cross-stratified sandstone with mud and mica drapes (facies 4) were chosen for the calculations, as these facies represent deposition by migration of dunes, which scale with flow depth. Large scale cross-stratified sandstone (facies 10), as well as compound cross-stratified sandstone (facies 11) are interpreted to represent deposition in bars, which do not scale with flow depth. The same concerns to smaller scale bedforms, such as ripple laminated sandstone (facies 7) and climbing ripple laminated sandstone (facies 8). In its turn sigmoidal cross-stratified sandstones (facies 12) are very scarcely represented in the study area, that no water depth calculations have been attempted. For calculations of the water depth for facies 2, 3 and 4 the most representative parts of these facies, taking into account their location, have been selected.

In addition to the outcrop data, drill-core data from different parts of the Rēzekne and Pärnu RS distribution area have been collected by lithostratigraphic logging (see Figure 1B). In total fourteen drill-cores with data of the Rēzekne and Pärnu successions have been described from core storages of the Estonian Academy of Science, the Estonian Geological Survey, the Department of Geology at the University of Tartu in Estonia, the Institute of Geology and Geography, Lithuania, and the Latvian Environment, Geology and Meteorology Centre. Thirty additional drill-core data have been used accordingly from other authors' works and data on the Rēzekne and Pärnu succession in Baltics (Sorokin, 1981; Kurshs, 1975; 1992; Kleesment, 1997; Stinkulis, 1998).

Based on the above data, the sedimentary facies were identified by sedimentary structures, textures, and composition and were grouped into facies associations based on the lateral and vertical relations of sedimentary facies.

4.1. The conceptual framework of facies analysis: from processes to sequences

The general aim of facies analysis is to recognize a hierarchy of stratigraphic elements in the sedimentary succession: from sedimentary facies to facies associations and their sequences, in order to identify the depositional processes and environments (systems). It is necessary to be able to further reconstruct the depositional systems tracts and the history of their behaviour, particularly progradational (regressive) and retrogradational (transgressive) phases reflecting relative base-level changes. Based on this, correlative time-stratigraphic surfaces, such as sequence boundaries, are recognized and cross-basinal stratigraphic correlations can be made. Facies analysis is thus the principal tool of sequence stratigraphy and sedimentary basin analysis (Nemec et al., 2003).

In the 1970s, the log concept combined with the Walther law ("*The deposits of the same facies area and similarly the sum of the deposits of different facies areas were formed beside each other in space, but in a lithostratigraphic profile we see them lying on top of each other*") gave rise to the concept of facies analysis and the development of profile-like environmental facies models (Middleton, 1973). The assemblages and the vertical stacking pattern of sedimentary facies shown by a log have become the basis for the analysis of depositional environments, that is, for the reconstruction of the coeval facies belts of an ancient environment, their depositional processes and their relative shifts or lateral migration with time. Facies models are generalizations based on a comparative study of the local facies successions of a particular ancient environment, commonly supported by studies of analogous modern environments (Walker, 1976; 1992). In constructing the facies model, geologists try to abstract from the local profiles only those features or facies association characteristics, that seem to be essential to an understanding of the nature of the depositional environment and its processes.

A short conceptual framework of facies analysis, used in this study, is outlined below.

Lithostratigraphic logging

Lithostratigraphic log is a graphical representation of the stratigraphic succession of rocks based on their systematic, unit-by-unit and bed-by-bed description from the drill-core or an outcrop sections (for details see Figure 1.3.). Lithostratigraphic logging is the basic technique of data acquisition and representation in sedimentary geology, especially for such main purposes as the palaeoenvironmental analysis, sequence stratigraphy and basin analysis.

The logging procedure itself is a systematic, unit-by-unit description of the sedimentary succession or its selected portions. The outcrop section or drill-core is divided completely into more or less uniform units or beds, which are described successively, using a standard set of conventional criteria. The log itself is a graphical representation of descriptive sedimentological information collected through the logging procedure. The vertical (thickness) scale adopted for the log should be sufficient to show the thinnest unit that needs to be distinguished, the appropriate scale for the log used in this study is 1:25, that is a 1 cm thickness in the log thus corresponds to 25 cm thickness in the lithostratigraphic section for measured sections in outcrops, while for drill-cores scale of 1:125 (1 cm thickness in the log corresponds to 1 m 25 cm thickness in the lithostratigraphic section).

The logs contain all the descriptive details that may be relevant to the sedimentological interpretation of the sedimentary succession. This information is coded with the use of various graphical symbols (explained in the legend), but some additional brief annotation can be given at the right-hand margin of the log.

Depositional processes and environments are interpreted from the logs. Therefore, it is critical that the information is collected carefully and systematically in the logging procedure, and that the facies descriptions are subsequently separated from the interpretations.

The distinction, description and interpretation of facies

When logging the sedimentary succession is subdivided into units, which differ macroscopically from the adjacent ones and represent some genetically coherent episodes of deposition. These units are the "building blocks" of the sedimentary succession (Figure 4.1.). Units of the same type are classified as one facies. A facies is a particular type of sedimentary deposit, or specific variety of "building block", and a sedimentary succession may contain many units of any particular facies (Nemec et al., 2003).

The interpretation of facies normally pertains to the depositional processes, or physical model of sediment emplacement. Different facies have different origin. Only few facies are environmentally diagnostic when considered in isolation. A term *lithofacies* is also used, however since the greek-derived prefix *litho-* may imply rocks or lithified deposits, a term sedimentary facies is usually used, which is clearer and applicable to both sedimentary rocks and non-consolidated deposits (Nemec et al., 2003).

Sedimentary facies are the basic types of sedimentary deposits, distinguished in descriptive macroscopical terms as the elementary "building blocks" of a sedimentary succession (Harms et al., 1975; Reading, 1986, 1996; Walker, 1992; Miall, 2000). The terms "facies" was introduced into geology by Nicholaus Steno in 1669, to mean the entire aspect of a part of the Earth surface during a particular interval of geological time. The modern geological usage of the term is closer to its meaning when introduced in 1838 by Gressly, who referred to the bulk of the lithological and palaeontological aspects of a stratigraphic unit, rather that Earth surface. However, Gressly's paper remained largely unknown until reprinted with comments by Wegmann (1963) and the same term had meanwhile been used with a number of different meanings (Teichert, 1958; Markevich, 1960; Middleton, 1973; Reading, 1986). The present-day usage of the term differs somewhat from Gressly's and derives rather from the definition given by Moore (1949): "Sedimentary facies is defined as any areally restricted part of a designated startigraphic unit which exhibits characters significantly different from those of other parts of the unit". According to this definition, facies are restricted in extent both stratigraphically and geographically, although the same facies may obviously be found repeated many times, at several levels, within the same stratigraphic unit (e.g., within a particular formation). The sedimentary facies were given brief descriptive names and informal symbol designations (see table 5.1.).

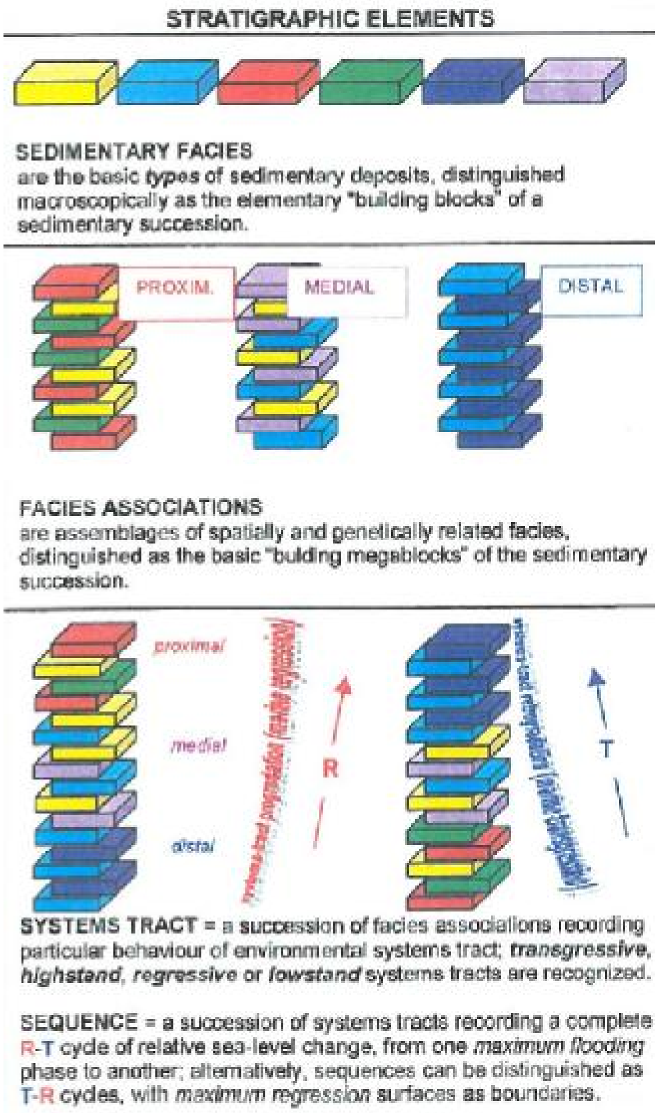


Figure 4.1. The building blocks of the facies, facies associations, systems tracts and sequences (After Nemec et al., 2003)
4.1. attēls. Fāciju, fāciju asociāciju, sistēmu kopu un secību uzbūve (pēc Nemec et al., 2003)

The distinction of facies associations

A *facies association* is an assemblage of spatially related and genetically coherent facies that overlie conformably one another. Facies associations are the "building megablocks" of a sedimentary succession, representing different depositional environments (see Figure 4.1.). Special attention is given to the lateral and vertical organization of facies within a particular facies association, by analysing jointly all its occurrences. This provides clues to the style of sedimentation or depositional processes operation in a particular ancient environment.

The distinction and analysis of facies associations lead to the identification of depositional systems (environments) which taken together indicate as to what kind of systems tract the sedimentary succession represents. *System* – although originally defined as a sedimentary rock (facies) assemblage by seismic stratigraphers, the term generally means a sedimentary environment. Based on knowledge of the environments, they are organized conceptually into depositional systems tract (facies association tract) by arranging them from relatively proximal to distal according to the general direction of seaward sediment transfer and inferred relative water depth. *Systems tract* – a spatial array, or "proximal-to-distal" belt of coeval environments through which the general transfer of sediments occur from land to sea; a three-dimensional environmental linkage of contemporaneous depositional systems (Nemec et al., 2003). Depending upon the relative sea-level behaviour, a systems tract may be transgressive or regressive, or may represent the sea-level highstand or lowstand.

Sequence stratigraphy

The vertical stacking of the stratigraphic succession indicates whether the depositional system tract is prograding (and its shoreline advancing) or retrograding (and its shoreline retreating) with time. This step of facies analysis brings to the topic of relative sea-level changes, where sequence stratigraphy concepts are used.

In the 1980s, the concept of a depositional systems tract emerged from seismic stratigraphy (Payton, 1977) and became the basis of the modern concepts of sequence stratigraphy (Posamentier et al., 1988; Van Wagoner et al., 1990). In sequence stratigraphy, the stratigraphic sequence of sedimentary facies associations are used to reconstruct the relative changes of the sea level (or base level) and the interplay through time, between the basin's accommodation space and the sediment accumulation. For this purpose, the lithostratigraphic profile with detailed facies log is used, which is further subject to facies analysis.

Sequence is a stratigraphic sequence of facies associations recording the complete cycle of the relative sea-level (or base-level) rise and fall. Sequence can be conceptually divided into systems tract and the choice of sequence boundaries may depend upon the requirements of a particular study (such as, unconformities, correlative conformities or transgressive surfaces).

Transgressive and regressive shoreline systems

An important key concept of the base-level change is that deltas and strandplains generally represent regressive shoreline condition, where fluvial, tidal, storm or wave influence respectively is dominant. On the other hand, estuaries and barrier/lagoon systems are developed under transgressive conditions at the shoreline (Figure 4.2.). This is how the base-level change affects the shorelines. This doesn't imply however that short segments of strandplain cannot develop during transgressions, or short lived barrier bars and lagoons cannot be present during regression of the coastline.

The above hierarchical order in the facies analysis and palaeoenvironment reconstruction is used in this study and a model of transgressive systems by Boyd et al. (1992) is applied to the studied succession.

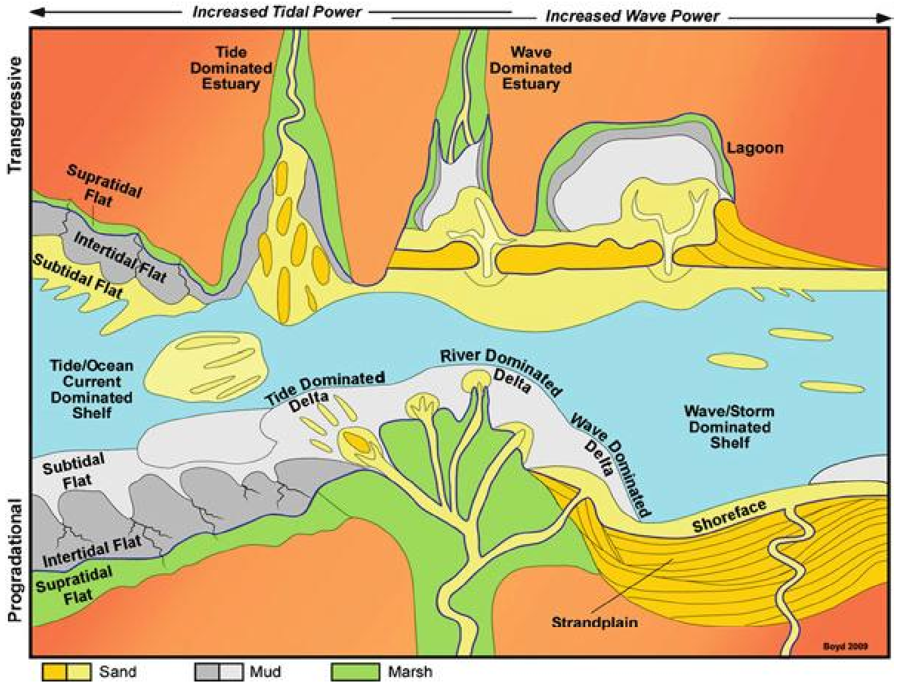


Figure 4.2. The distribution of major coastal depositional features (After Boyd et al., 1992).

4.2. attēls. Galvenās piekrastes sedimentācijas vides un to izvietojums (pēc Boyd et al., 1992)

5. Results and interpretation

The results of study and their interpretation are discussed in two chapters. The Chapter 5.1. focuses on facies and facies associations of deposits of the Pärnu Fm documented in outcrop belt. This part of results and interpretation deals with the deposits studied in most detail, as the wide outcrop belt provides excellent opportunities for facies analysis and detailed bed architecture studies.

The Chapter 5.2. describes facies associations in whole studied area – the Baltic States (see Figure 1). Therefore this chapter combines data previously discussed in Chapter 5.1., but is built mainly on the data collected from the drill-cores.

5.1. Facies architecture and deposition in sand-rich estuary and fluvial-tidal transition zone in the northern part of the Baltic Devonian Basin during the Pärnu time

5.1.1. Sedimentary facies

Thirteen sedimentary facies have been distinguished based on their sedimentary structures and lithology deposits of the Pärnu Fm documented in outcrop belt. See Table 6.1.1.1. for summary on the sedimentary facies (F), containing facies description and interpretation, distinguished from the whole study area, including the outcrop belt. Below detailed description and interpretation of facies, found only in outrop belt, is given.

Facies 1: Mudstone and siltstone conglomerate

Mudstone and siltstone conglomerate is composed of crudely shaped mudstone and siltstone pebbles of variable sizes, up to 30-60 cm in diameter. Matrix of the conglomerates consists of sandstone similar to the under- and overlaying sandstone beds. In the lens-shaped, more angular varieties of siltstone pebbles, numerous very well preserved fragments of plant remains, such as *Psilophytites* sp. and *Hostinella* sp. occur (Kalamees, 1988). These plant remains occur in several sites of the study area (outcrops 1, 4, 2) and are abundant only in F1. In some places fossil fish remains occur (Kleesment, Mark-Kurik, 1997). F1 has in all places an erosional and irregular base. Bed thickness is most commonly up to 0.5 m. The conglomerates can be traced laterally for tens of meters. In most places the lateral pinch-out of the conglomerates is gradational. Thickness of beds decreases two times over a distance of 3-5 m. Upper contact of the conglomerates is usually gradational, and F1 normally grades into the trough cross-stratified mudstone conglomerate (F2). F1 occurs in all the studied outcrops, and is commonly associated with F 2, 3 and 6.

Interpretation:

The mudstone and siltstone conglomerate is interpreted to be deposited as rip-up clasts on the channel scours and as lags from high energy currents (Collinson, 1996). The mudstone and siltstone clasts are locally derived.

Facies 2: Trough cross-stratified mudstone conglomerate

Trough cross-stratified mudstone conglomerate is matrix supported and consists of rather rounded mudstone clasts and a medium- to coarse-grained sandstone matrix. The mudstone clasts are of granule and pebble size, in most cases 0.5-2 cm, rarely up to 4 cm in diameter. The clasts concentrate mostly on toesets of the cross strata. Fish bones, quartz pebbles, as well as *Psilophytites* sp. and *Hostinella* sp. plant remains typically occur together with mudstone pebbles. The thickness of cross-sets varies from 4 to 40 cm, in depositional units 0.43-1.2 m thick that are based by erosion surfaces in most places and form lenticular units. The lower bed contacts are usually sharp and erosional, the upper bed contacts are gradational and F2 grades into the trough cross-stratified sandstones of F3. Palaeocurrent directions derived from cross-strata are very variable, mainly in a range of 45-165°. Most of the palaeocurrents are towards 135-165° and 45-95° (Figure 5.1.). The calculated water depth for F2 according to Leclair and Bridge (2001) equations is ca. 1-7 m; and according to Dalrymple and Rhodes (1995) equations ca. 1-4 m.

F2 is widely distributed in all the studied outcrops. It is associated with F 1, 3 and 6, but most commonly underlies F3.

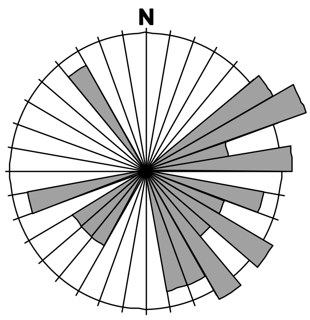


Figure 5.1. Palaeocurrent directions derived from Facies 2 cross-strata (N=32, outer circle represents 10 %)

5.1. attēls. Slīpo slānīšu krituma azimutu rozes-diagrammas 2. fācijai (N=32, ārējā loka malas vērtība ir 10 %)

Interpretation:

Trough cross-stratified mudstone conglomerate is deposited from traction currents by migration of 3-D dunes in channels in a high energy environment. The presence of rounded clay clasts indicate that current was strong enough to erode and transport the clayey material. The angular form of the clay clasts, as well as the occurrence of the F 2 in association with erosional surfaces and scours, indicate high energy bedload deposition (Collinson, 1996; Miall, 1996).

Facies 3: Trough cross-stratified sandstone

Trough cross-stratified sandstone is one of the most common facies in the studied outcrops. It consists of fine- to coarse-grained sandstone, in places with clay clasts on the toesets of the cross strata. Two variations of this facies exist: variably grained (F3a) and well-sorted (F3b). In some places overturned and convolute cross-strata occur. These deformation structures are typical only in parts of the cross-sets, rarely in the whole sets. Usually the upper parts of the individual cross-sets are deformed. In some cases within the individual cross strata grain size decreases away from the concave-up scours. Trough cross-stratified sandstones occur as erosionally based, thick lenticular units with mostly gradational upper contacts. The thickness of cross-sets is 4-50 cm, the depositional units are up to 3.3 m thick. Palaeocurrent directions derived from cross strata vary between 75-255°, with most of paleocurrent measurements between 105° and 165° and also from 225° to 255° (Figure 5.2). The calculated water depth for F3 according to Leclair and Bridge

(2001) equations is ca. 0.7-5 m, and according to Dalrymple and Rhodes (1995) equations is ca. 0.9-4 m.

F3 is widely distributed in all the studied outcrops. It is associated mostly with F2, as well as with F4, 9 and 11.

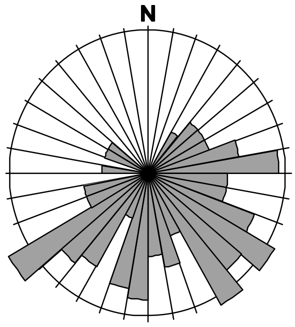


Figure 5.2. Palaeocurrent directions derived from Facies 3. cross-strata. (N=111, outer circle represents 10 %)
5.2. attēls. Slīpo slānīšu krituma azimutu rozes-diagrammas 3. fācijai (N=111, ārējā loka malas vērtība ir 10 %)

Interpretation:

Trough cross-stratified sand beds are deposited from traction currents by migration of 3-D dunes in fluvial channels. Overturned cross lamina formed as a result of post-depositional deformation due to drag under overpassing currents (Collinson, 1996). The fining upwards grain size in individual cross strata suggests waning currents during deposition of each of the individual strata, thus indicating unsteady current regime, probably due to occasional tidal influence (Middleton, 1991; Nio, Yang, 1991, Pontén, Plink-Björklund, 2007).

Facies 4: Cross-stratified sandstone with mud and mica drapes

Cross-stratified sandstone with mud and mica drapes is one of the most common F in the studied outcrops. F4 consists of very fine to fine-grained cross-stratified sandstone with mud or mica laminae that drape almost every individual cross strata. The mud drapes are typically single drapes with thickness of several mm. The alternating mud and sand forms bundles, associated with small erosional surfaces within the cross sets – the reactivation surfaces. In some places, in the toesets of cross strata, flame structures occur as small darker lenses. Current ripple lamination and reactivation surfaces are common. The thickness of cross-sets is 3-35 cm, and the depositional units are in most places 0.3-2 m thick. In places the thickness of individual cross strata changes systematically along the sets. F4 forms erosionally or sharply based lenticular units. In places it occurs as single, small unit within F3. Palaeocurrent measurements derived from cross strata show bimodal directions between 105-165° and 195-255°, with majority of the directions between 195°-225° (Figure 5.3.). The calculated water depth for F4 according to Leclair and Bridge (2001) equations is ca. 1.5-9 m, and according to Dalrymple and Rhodes (1995) equations ca. 1.5-7 m.

The cross-stratified sandstone with mud and mica drapes occurs almost in all the studied outcrops, and is associated with F 7, 8, 11 and 12.

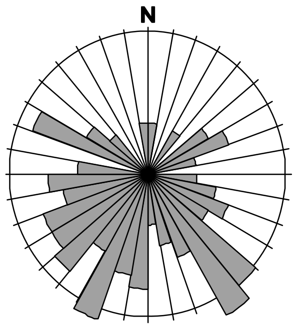


Figure 5.3. Palaeocurrent directions derived from Facies 4 cross-strata. (N=96, outer circle represents 10 %)
5.3. attēls. Paleostraumju virzieni 4. fācijai. (N=96, ārējā loka malas vērtība ir 10 %)

Interpretation:

Cross-stratified sandy beds with mud and mica drapes are deposited from traction currents by migration of 3-D dunes (Collinson, 1996), where currents were systematically accelerating and decelerating with slack-water periods. Such cyclic changes of flow velocity are common in tidal environments (Nio, Yang, 1991). Mud and mica drapes on the cross strata deposited from suspension during slack water periods (Dalrymple, 1992). The lack of double mud drapes and abundance of single mud drapes may indicate that subordinate currents were strong enough to erode the first mud drape (Visser, 1980). The alternating mud and sand, as well as the systematic changes in the bundle thicknesses associated with reactivation surfaces, also indicate tidal influence. A systematic change in bundle thickness along the cross-sets, suggests deposition from alternating neap and spring tides (Nio, Yang, 1991; Dalrymple, 1992).

Facies 5: Plane-parallel stratified sandstone

Plane-parallel stratified sandstone is composed of very fine to medium-grained sandstone with parallel stratification. F5 occurs rarely and does not form separate units, it occurs as interbeds within the other facies. The thickness of sets is variable, 4-37 cm, depositional units are only 9-65 cm thick. The lower contact of the F5 is commonly sharp, the upper contact is gradational. The parallel stratification is lined by mud or mainly mica. Mica is a typical compound of this facies. Mud laminae are usually non-continuous.

Plane-parallel stratified sandstone is distributed very scarcely in the studied outcrops. It is associated with F 3 and 4, as well as in several places it underlies F7. The contact with F3 is erosional, with F4 and 7 mainly sharp and in places erosional.

Interpretation:

Plane-parallel stratified sandstone formed from unidirectional currents by movement of plain beds (Collinson, 1996). The association of these fine- to medium grained sandy deposits with cross-stratified variously grained, generally more coarse grained sandstones, and their occurence in channels suggests that F5 deposited in the upper flow regime (Reineck, Singh, 1980). The close association with F3, as well as the presence of mud and mica drapes on the parallel laminae suggests the tidal influence in the deposition of this facies.

Facie 6: Structureless sandstone

Structureless sandstone is composed of very fine to coarse-grained sandstone, with no visible structures. F6 occurs scarcely, it usually forms interlayers or small-scale lenticular units. The thickness of sets is 9-23 cm, but depositional units are only 0.2-2 m thick. Two varieties of F6 have been distinguished according to its occurence in the different parts of the section: 1) very fine to fine-grained, well-sorted, cemented sandstone and 2) medium to coarse-grained, poorly-sorted and loose sandstone. The first type occurs in the very upper part of the section and has a gradational contacts with both over- and underlying facies. It is associated with F 5, 7 and 13. The latter type of the F usually occurs in the lower part of the section in sharp-based lenticular beds. The upper contact is usually sharp, rarely gradational. The material is poorly sorted, flame structures occur in places.

F6 is typically associated with F3 and 9.

Interpretation:

The first variety of F6 is interpreted to be deposited as plane beds in lower flow regime, and only seems to be ungraded. Due to the homogenous sand grain size the sedimentary structures are not visible. The second variety of F6 is interpreted to have formed from rapid deposition (Collinson, 1996), where no traction occured and thus no structures formed (Miall, 1996), due to very high sedimentation fallout rates. The flame structures indicate deformation caused by pore water escape and high sediment load (Reineck, Singh, 1980).

Facies 7: Current ripple cross-laminated sandstone

Current ripple cross-laminated sandstone consists of very fine to fine-grained sandstone that is cross-laminated, and rich in mud and mica. The thickness of cross-laminae is several mm, the thickness of depositional units varies from 2 to 70 cm. Cross-laminae are in most places lined by mud or mica, draping almost every individual laminae. The contacts with other facies are sharp and gradational, rarely erosional. F7 does not form separate depositional units, it occurs normally as interlayers and in toesets of cross-stratified sets. Most often it occurs on toesets of cross-stratified sandstones (F4) with ripple migration directions opposite to the direction of cross-strata. Measurements derived from cross laminae show paleocurrent directions towards 195-255°, with the majority of the measurements between 225-255° (Figure 5.4.).

F7 is most commonly associated with F4 and F5, as well as F6 and F8.

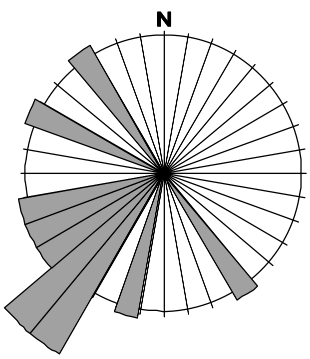


Figure 5.4. Palaeocurrent directions derived from cross-laminae of Facies 7. (N=11, outer circle represents 10 %)

5.4. attēls. Ripsnojuma slīpo slānīšu krituma azimutu rozes-diagrammas 7. fācijai (N=11, ārējā loka malas vērtība ir 10 %)

Interpretation:

Current ripple cross-laminated sandstone formed from low energy currents by migration of ripples (Reineck, Singh, 1980). Current ripples may coexist with 3-D dunes and occur on the toes of the dune lee-sides which are subjected to lower flow velocities and flow separation cells. F7 may have formed as back-flow ripple sets deposited on the bottomsets of the dunes from unidirectional currents, or be deposited from the subordinate current of assymetrical tidal currents. However, the occurrence of reverse directions of ripple cross-lamination, as well as mud and mica drapes on the cross laminae, deposited from suspension during slack water periods, suggest tidal origin (Nio, Yang, 1991; Dalrymple, 1992).

Facies 8: Climbing ripple laminated sandstone

Climbing ripple laminated sandstone consists of very fine to fine-grained sand material that is cross-laminated, and mica or mud rich. The thickness of cross-laminae is from few mm to 1 cm, the thickness of the units varies from 10 to 130 cm. Lamination is in most places lined by mica drapes, and in some places by mud drapes. The lower contact with other facies is sharp, in places even erosional, the upper contacts are mainly gradational. F8 forms rather small units and occurs usually on the toesets of cross-stratified sandstones. However in places, climbing ripple laminated sandstone forms thicker (up to 1-1.5 m) and more extensive (about 10 m) units, where the shape of climbing ripple lamination changes repetitively upwards. Two varieties of climbing-ripple lamination and transitional forms between them have been distinguished: 1) The angle of climb of the ripple laminae changes from vertical to low angle (ripple laminae in-phase and ripple laminae in-drift of Jopling and Walker, 1968). Where the angle of climb is vertical, the ripple crests occur directly above each other, rarely with a slight downcurrent shift of crests. Where the angle of climb is lower, the ripple laminae have low-angle sub-parallel bounding planes which delineate the beds and climb in downcurrent direction. The bounding surfaces in F8 dip in the opposite directions. There is vertical transition and repetition of vertically climbing and downstream migrating climbing ripples lamination.

Palaeocurrent measurements derived from cross laminae show highly variable directions between 75-315°. Two main directions are towards 105-135° and 165-195° (Figure 5.5.).

F8 is associated with F4, 11 and 6.

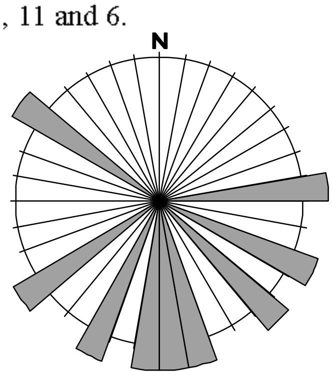


Figure 5.5. Palaeocurrent directions derived from cross-laminae of Facies 8. (N=9, outer circle represent 10 %)

5.5. attēls. Ripsnojuma slīpo slānīšu krituma azimutu rozes-diagrammas 8. fācijai (N=9, ārējā loka malas vērtība ir 10 %)

Interpretation:

Climbing ripple laminated sandstone formed by migration of ripples in a high suspended sediment load with high sedimentation rate (Reineck, Singh, 1980). Mud and mica drapes deposited during the slack water regime due to tidal currents (Nio, Yang, 1991). Paleocurrent directions derived from cross-laminae, indicate a reverse or oblique deposition to the main bedding surfaces, possibly due to reverse tidal currents. Vertical repetitive transition of two types of climbing ripple-lamination indicates cyclic increase and decrease of deposition rates due to changes in the ratio between sedimentation rates and downstream ripple migration rates. The deposition is interpreted to have occurred from periodically waning tidal currents (Lannier, Tessier, 1998).

Facies 9: Sandstone with deformation structures

Sandstone with deformation structures consists of fine-grained sand material with overturned cross-strata, convoluted and buckled cross-strata, as well as flame structures The thickness of the overturned units is 5 to 50 cm, similar to the set thickness in F 3 and 4, where the overturned cross strata commonly occur. The deformation affects only the topmost 10-12 cm of the cross-sets. In contrast, flame structures are closely associated with F 6 and form thicker, in places more than 1 m thick units. The latter sandstone has sharp lower contacts, whereas the lower contact of the overturned cross-stratification is gradational.

F9 occurs very scarcely. It is closely associated with F3, 4, as well as F6 and 12.

Interpretation:

The overturning of the upper parts of the cross sets indicates deformation due to shear stress caused by overriding traction currents on soft sediments (Owen, 1996). The flame structures indicate deformation caused by pore water escape and high sediment load (Reineck, Singh, 1980).

Facies 10: Large scale cross-stratified sandstone

Large scale cross-stratified sandstone consists of fine to coarse-grained sandstone, in places with clay and quartz pebbles and fossil fish remains as lags. It is formed by inclined master bedding surfaces with superimposed low angle cross-strata, dipping down the master surfaces. It forms thick, lenticular beds. Both upper and lower contacts of F10 are sharp and erosional. Only in a few places the upper contact is gradational. Superimposed cross-stratified sets are 10-30 cm thick, and the large-scale units are up to 2 m thick. F10 can be traced laterally for more than 30 m. Palaeocurrent directions derived from cross strata measurements show bimodal distribution and vary between 105° to 165° and 195° to 255° with majority of the directions between 135-165°. The master bedding surfaces dip towards 195-225° (Figure 5.6.).

F10 is very common and associated with F2, 3, 7, 11 and 12.

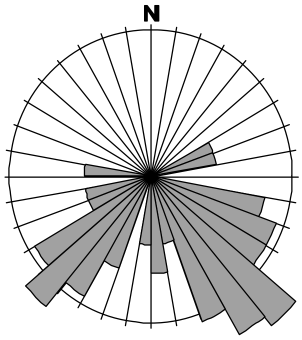


Figure 5.6. Palaeocurrent directions derived from cross-strata and master bedding surfaces of Facies 10. (N=56, outer circle represents 10 %)
5.6. attēls. Slīpo slāņīšu un slāņojuma virsmu krituma azimutu rozes-diagrammas 10. fācijai (N=56, ārējā loka malas vērtība ir 10 %)

Interpretation:
Large scale cross-stratified sandstone is interpreted to have formed by migration of 3-D dunes (Miall, 1996) on tidally influenced channels or tidal bars (Dalrymple, 1992). The master surfaces represent bar accretion surfaces and the dunes migrated superimposed on the barforms. The close association of F10 with other facies, rich in tidal signatures, indicates deposition from tidal currents.

Facies 11: Compound cross-stratified sandstone
Compound cross-stratified structure is known also as inclined cross-bedding (Dalrymple, 1984) and has been called double-cross stratification by Kurshs (1975, 1992). F11 consists of fine to coarse-grained sand, draped in places by non-continuous mud or mica laminae. The most typical feature of F11 is inclined erosional surfaces superimposed by cross-sets – reactivation surfaces. Compound cross-stratified sandstone consist of decimetre-scale cross-sets, separated by inclined reactivation surfaces. The inclined surfaces and the cross-strata dip in the same direction in general. Foresets of the reactivation surfaces vary from low to high angle and dip at 9 to 28°. The reactivation surfaces separate cross-sets into a series of sigmoidal sets, several meters wide in lateral extent. A variation in the lateral spacing of reactivation surfaces is typical. Mud drapes rarely extend to the top of the cross-strata, they pinch out on the mid-foreset by the truncating reactivation surfaces. It is typical for F11 that the grain-size decreases upwards from the concave-up scours, similar to F3. The compound cross-stratified sandstone form 20 to 220 cm thick lenticular units, and the cross-sets are 5-40 cm thick. The contact with other facies is in most places sharp, the lower contact is erosional. Palaeocurrent directions derived from cross strata show unimodal distribution 195-285°, but the inclined reactivation surfaces are oriented towards 255-285° (Figure 5.7.).

F11 is widely distributed in the studied outcrops, and is usually associated with F3, 4 and 10.

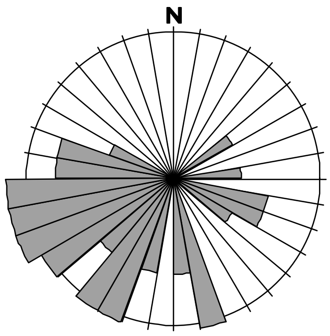


Figure 5.7. Palaeocurrent directions derived from cross-strata and reactivation surfaces of Facies 11. (N=58, outer circle represents 10 %)
5.7. attēls. Slīpo slānīšu krituma azimutu un reaktivācijas virsmu rozes-diagrammas 11. fācijai (N=58, ārējā loka malas vērtība ir 10 %)

Interpretation:

The compound cross-stratified sandstone is interpreted to be formed by migration of 3-D dunes on tidal or tide-influenced bars or compound dunes (Dalrymple, 1984; Dalrymple, Choi, 2007). The cross-stratification is formed by the dominant current and the reactivation surfaces are formed by subordinate current (Visser, 1980; Willis et al., 1999). Variations in the lateral spacing of reactivation surfaces within the cross sets indicate significant changes in the rate of bedform migration during successive tidal periods. The fining-upwards grain size on the individual laminae scale suggests distinctly waning currents during the deposition of each individual cross strata, thus suggesting tidal influence (Middleton, 1991; Nio, Yang, 1991, Pontén, Plink-Björklund, 2007). The mud/mica drapes indicate deposition in slack water periods (Dalrymple, 1992). Their indistinctive character suggests subsequent erosion by subordinate tidal current.

Facies 12: Sigmoidal cross-stratified sandstone

Sigmoidal cross-stratified sandstone consists of very fine to fine-grained sandstone, in places rich in mica and mud. The trough cross-sets have characteristically a sigmoidal shape. Mud and mica drapes occur on the entire cross-strata, or occur only on the toesets of the cross-strata, forming a repetitive alternation of mudstones and sandstones. Reactivation surfaces are common. F12 occur in all places as interbeds within other F and forms small lenticular units. The cross sets are only of 8-17 cm thick, the bed contacts of the sigmoidal cross-stratified sandstone are in most places sharp, but in places gradational. The paleocurrent measurements derived from cross-strata dip azimuths are towards 195-255° (Figure 5.8.). F12 is closely associated with F4, 11 and less with F3.

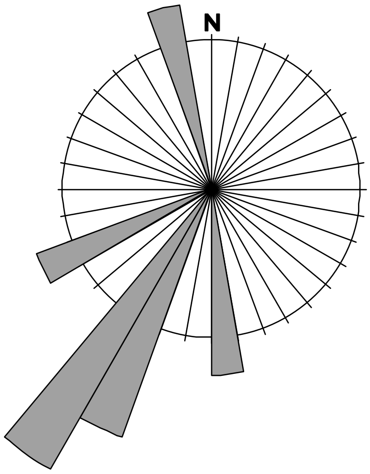


Figure 5.8. Palaeocurrent directions derived from cross-strata of Facies 12. (N=8, outer circle represents 10 %)
5.8. attēls. Slīpo slānīšu krituma azimutu rozes-diagrammas 12. fācijai (N=8, ārējā loka malas vērtība ir 10 %)

Interpretation:

The sigmoidal cross-stratified sandstone is interpreted to have formed as migrating 3-D dunes under tidal influence. According to Kreisa and Moiola (1986), sigmoidal bedding is one of the diagnostic structures for a tidal origin. The sigmoidal tidal bundle pattern reflects deposition from one dominant tidal current that deposits a set of cross-strata, enclosed by two relatively gently dipping, sigmoid-shaped reactivation planes formed by the subordinate current (Kreisa, Moiola, 1986; Shanley et al., 1992). Mud and mica drapes were deposited during the slack water periods (Visser, 1980; Nio, Yang, 1991).

Facies 23: Plane-parallel laminated and structureless dolostone

Plane-parallel laminated and structureless dolostone is light brownish-grey and has cryptocrystalline texture. F23 is mainly structureless and platy; only in places it is plane-parallel laminated. Pyrite concretions occur on bedding surfaces of F23. Plane-parallel laminated and structureless dolostone forms small, tabular units, only 5-25 cm thick. F23 has mainly gradational and sharp contacts with the underlying facies and occurs exceptionally in the topmost part of the studied outrcrops. It overlies F7 and 5.

Interpretation:

The plane-parallel laminated and structureless dolostone is interpreted to have formed in a low energy environment by deposition of carbonate mud (Reineck, Singh, 1980). During diagenetic processes the calcium carbonate mud or already lithified limestone was replaced by dolomite (Tucker, Wright, 1990).

Table 5.1. Summary of facies from the study area: description and interpretation
5.1. tabula. Pētījumu teritorijas fāciju īss apraksts un interpretācija

	FACIES	TEXTURE	STRUCTURE	THICKNESS, GEOMETRY	INTERPRETATION
Siliciclastic Facies					
1	**Mudstone and siltstone conglomerate**	Clay and silt clasts, up to 60 cm in diameter, with fossil flora fragments (*Psilophytites* sp. and *Hostinella* sp.) and fossil fish fragments, matrix-supported, sandstone matrix	Crudely stratified conglomerate	Erosionally based lenticular and irregular units: 5-50 cm	Deposition of rip-up clasts on channel scours and as lags from high energy currents
2	**Trough cross- stratified mudstone conglomerate**	Medium- to coarse-grained sandstone with clay pebbles 0.5-4 cm in diameter, matrix supported	Trough cross-stratification with rounded clay clasts on scour surfaces	Erosionally or sharply based lenticular units, gradational upper contact. Cross sets: 4-40 cm, depositional units: 43-120 cm	Deposition from traction currents by migration of 3-D dunes in channels in a high energy environment
3a	**Trough cross-stratified sandstone, variably-grained**	Fine- to coarse-grained and gravelly sandstone, in places with clay clasts on scour surfaces; In some individual cross strata grain size decreases away from the concave-up scours	Through cross- stratification, in places soft-sediment deformations are present	Erosionally based thick lenticular units, erosional contacts. Depositional units: 1 – 3,5; max – 7-8.2 m	Deposition from traction currents by migration of 3-D dunes in fluvial channels and tidally-influenced fluvial channels. The decrease of grain size in individual cross strata, indicate unsteady current regime
3b	**Trough cross-stratified sandstone, well sorted**	Fine- to coarse-grained sandstone, in places with clay clasts on scour surfaces	Through cross- stratification, in places soft-sediment deformations are present	Erosionally based thick lenticular units, usually gradational upper contact. In outcrops: Cross sets: 4-50 cm, depositional units: 20-330 cm; In cores – 5,3 m, max – 25 m	Deposition from traction currents by migration of 3-D dunes in tidal channels
4	**Cross-stratified sandstone with mud and mica drapes**	Fine-to medium-grained sandstone, usually well-sorted	Trough cross-stratification, where mud and mica laminae systematically drape almost every cross-strata, reactivation surfaces are present; alternating mud and sand forms bundles	Erosionally based lenticular units. Cross sets: 3-35 cm, depositional units: 30-200 cm; In cores – 2.5-4 m, max – 10 m.	Deposition from traction currents by migration of 3-D dunes in tidal channels and bars, where the current strength and directions changed systematically under the tidal influence

5	**Plane-parallel laminated sandstone**	Very fine- to medium-grained sandstone	Plane-parallel stratification, draped in places by non-contiguous mud or mica laminae	Occurs in rare places and as interlayers, sharp or gradational contacts. Sets: 4-37 cm, depositional units: 9-65 cm, in cores max – 5.9- 6 m	Deposition from plain beds in lower flow regime and in upper-flow-regime currents, tidally influenced
6	**Structureless sandstone**	Very fine- to coarse-grained sandstone	Structureless, in places soft-sediment deformed	Small units, in places as interlayers; erosionally based, gradational upper contact. Sets: 9-23 cm, depositional units: 20-200 cm; In cores -1-3 m, max – 6,9-9,5 m.	Deposition as plane beds in lower flow regime and/or rapid sedimentation due to high deposition rates with no preservation of sedimentary structures
7	**Current ripple cross-laminated sandstone**	Very fine- to fine-grained sandstone, mica and mud rich	Ripple cross-lamination; in places opposite ripple migration directions in cross-sets	Interlayers and in toesets of cross-stratified sets; sharp or gradational contact. Laminae: a few mm, units: 2-70 cm; In cores – up to 1.5 m, max – 4 m	Deposition from low energy currents by migration of ripples; reverse directions and mud/mica drapes indicate tidal influence
8	**Climbing ripple-laminated sandstone**	Very fine- to fine-grained sandstone, mica and mud rich	Climbing ripple lamination. Two types are distinguished: ripple laminae in-phase and ripple laminae in-drift	Interlayers, in toesets of cross-stratified sets, erosionally based or gradational contacts. Laminae: a few mm-cm, units: 10-130 cm	Deposition by migration of ripples in a high suspended sediment load with high sedimentation ratio. Vertical repetitive transition of two types of climbing ripple-lamination indicate tidal influence.
9	**Sandstone with deformation structures**	Fine-grained sandstone	Overturned, convoluted and buckled cross-stratification, flame structures	Small units, affects separate layers, usually gradational contact, but sharp in structureless units. Sets: 5-50 cm; Units: 113 cm: In cores – up to 3.3 m	Deposition from traction currents, the secondary soft sediment deformations caused by shear stress and escaping pore water
10	**Large scale cross-stratified sandstone**	Fine- to coarse-grained sandstone with clay pebbles and quartz grains, fossil fish fragments	Inclined master bedding surfaces with superimposed low angle cross strata dipping down the master surfaces	Erosionally based thick lenticular beds, sharp contacts. Sets: 10-30 cm Units: 50-200 cm	Deposition by migration of 3-D dunes in tidal channels and bars
11	**Compound cross-stratified sandstone**	Fine- to coarse-grained sandstone	Inclined main bedding surfaces and cross-stratification with reactivation surfaces with angle of dip from 9 to 28° and draped in places by non-continuous mud	Lenticular bodies with sharp, erosionally based contacts. Sets: 5-40 cm, Units – 20-220 cm	Deposition by migration of 3-D dunes in channels and bars under tidal influence: main bedding surfaces and cross-strata formed from dominant current, reactivation surfaces formed from

			or mica laminae		subordinate currents, mud/mica drapes from slack water periods, subsequently partly eroded by subordinate currents
12	**Sigmoidal cross-stratified sandstone**	Very fine- to fine-grained sandstone	Trough cross-stratification of sigmoidal shape, in places mud laminae drape cross-strata, reactivation surfaces occur in places	Small lenticular units, sharp and gradational contacts. Sets: 8-17 cm, occur as single sets	Deposition by migration of 3-D dunes under tidal influence
13	**Wavy-laminated sandstone**	Very fine- to fine-grained sandstone	Sandstone with irregular wavy lamination, in places stressed by mudstone	Rather small lenticular units: 0,5-2,5 m, sharp and gradational contacts.	Deposition by traction in oscillatory flow conditions
14	**Sandstone with siltstone laminae**	Very fine- to fine-grained sandstone with thin siltstone laminae and lenses	Heterogeneous structure: flaser bedding	Sharp and gradational contacts, units from 0.5 to 2 m thick	Sandstone form by traction currents and mudstone drapes form during slack-water on tidal flats in subtidal environment
15	**Heterogeneous sandstone and siltstone lamination**	Very fine- to fine-grained sandstone and siltstone	Heterogeneous structure: thinly laminated sandstone and siltstone interlayers, plain-parallel or wavy bedding	Sharp and gradational contacts, units from 0.3 to 2.5 m thick	Alternating deposition from suspension and from traction currents, deposited by tidal currents in intertidal environment
16	**Mudstone/Silstone with sand laminae**	Siltstone with very fine- to fine-grained sandstone laminae and lenses	Heterogeneous structure: plain-parallel lamination or lenticular bedding	Sharp and gradational contacts, units up to 3 m thick	Ripples form by traction currents and mudstone drapes form during slack-water on tidal flats in intertidal to supratidal environment
17	**Laminated siltstone**	Siltstone	Small scale plain-parallel lamination	Sharp and gradational contacts, units up to 1 m thick	Deposition in low energy environment by suspension settling
18	**Homogenous siltstone and mudstone**	Siltstone and mudstone	Homogenous structure, bioturbated in places	Gradational contacts, units up to 1 m thick, max 4 – 13 m	Deposition in low energy environment, slack water conditions or rapid deposition from fluid muds
Carbonate-rich Facies					
19	**Variably-grained sandstone with dolomite ooids and peloids**	Sandstone fine- to coarse-grained, dolomite ooids and peloids, in places dolomitic marl matrix present	Lenticular, wavy, cross-stratified, plane-parallel lamination	Cross-laminae: 1-5 mm, finer – 2-8 mm, cross-sets up to 15 cm; Plain-parallel laminas: 0,1-1 cm or 0,1-0,5 cm and 1 mm up to 0,2-1 cm; units: some cm up to 50 - 2m, max 4,8 m	Deposition in tidal currents on tidal flats, dolomitic marl deposited in slack water periods

Table continued
Tabulas turpinājums

20	**Dolomitic marl**	Carbonate rich siltstone, in places gypsum present	Homogenous or plain-parallel laminated, in places desiccation cracks and tepee structures present	Sets: 5-20 cm, units: some tenth cm up to 4 m, max – 20 m	Deposition in low energy carbonate-rich environment from suspension on inter-tidal to supratidal carbonate-rich tidal flats, periodical subaeral exposure
21	**Heterogeneous dolomitic marl with silt and very fine sand lamination**	Carbonate rich siltstone, admixture of sand and silt in laminae and lenses	Heterogeneous structure: lenticular bedding, irregular wavy	Laminae: less than few mm up to some cm, units from 60 cm to 2,5 m, max – 10 m	Alternating deposition from suspension and from traction currents, deposited by tidal currents in intertidal to supratidal environment
22	**Dolomitic marl with sandstone grains**	Carbonate rich siltstone, with irregular distribution of coarse and medium sized quartz and feldspar grains	Irregular wavy and lenticular bedding; tepee structures common	Laminae: some mm up to 3-4 cm; units – up to 1-3 m thick; high of wave ripples - 2,4-2,5 mm, length – 16,5-18,3 mm, max – 40 m.	Deposition on carbonate-rich tidal flats in inter- to supratidal environment; occasional deposition in oscillatory flows under wave action; periodical subaeral exposure
23	**Dolostone**	Carbonate-rich mudstone, admixture of clay, mica, gypsum, halyte pseidomorphs	Structureless , plane-parallel laminated, and wavy structured, in places pyrite concretions occur and desiccation cracks	Small tabular units with mainly gradational, and sharp contact, occurs exceptionally on the very top of sections in outcrops. Sets and units: 5-25 cm. In cores: laminae: 0,1-3 cm; 0,1-0,5 cm and 0,1-0,2 mm; units: 0,25 – 2,0 m, max – 2,5 m	Carbonate precipitation and mud settling in a low energy environment and in oscillatory flow conditions
24	**Nodular facies**	Carbonate-rich mudstone to siltstone, in places with very fine- to coarse-grained sand and claystone pebbles	Carbonate nodules, concretions and networks of calcite veins filled with sandstones and mud clast	Units: 0,3 – 1,0 m, max 3 m	Formed in subaeral conditions due to carbonate dissolution on supratidal flats and top of bars
Bioturbated Facies					
25	**Bioturbated facies**	Sandstone, siltstone, mudstone	Bioturbated, even brecciated, primary structure destroyed	Units: few cm to 1,5 m, max – 4, 8-5,7 m.	Formed in supratidal flats, associated closely with Facies 26

5.1.2 Facies Associations

The sedimentary facies distinguished from deposits of the Pärnu Fm documented in outcrop belt are grouped into four facies associations (FA), according to their lateral and vertical relationships (Figure 5.9.). The facies associations are distinct, what concerns the combinations of sedimentary structures, textures, composition, bed geometry, character of vertical and lateral bed transitions, as well as palaeocurrent directions.

Distribution of facies associations was mapped across the 6 km wide outcrop belt (see Figures 1 and 5.9.). Correlations within continuous outcrops, 25-350 m wide, are based on walking out stratigraphic packages and surfaces, as well as on detailed photomosaics. Correlations between the individual outcrops are based on walking out stratigraphic packages and boundaries and tracing on topographic maps, taking into account the regional tectonic dip.

Fluvial deposits (FA 1) occur throughout the entire outcrop belt, except for the very southern (basinward) part of the area. Stratigraphically they occupy the lower part of the outcrop belt (see Figure 5.9.). The deposits that mark transition from fluvial to tidal dominance (FA 2) occur at three different stratigraphic levels, separated by FA 1 and FA 3. The tide-dominated bar deposits (FA 3) occupy the upper part of the succession and are present across the outcrop belt. They form the bulk volume of the succession in northern part of the outcrop belt. The tidal flat deposits (FA 4) are restricted to the uppermost part of the succession and commonly overlay tide-dominated bar deposits of FA 3. The overall vertical transition is from tide-influenced fluvial to tidal channel deposits at the base (FA 2) to erosionally based fluvial deposits (FA 1), overlain erosionally by fluvial to tidal channel deposits (FA 2), which are erosionally overlain by tide-dominated deposits (FA 3), and they are erosionally overlain by FA 2 further southward. The entire outcrop belt is capped by tidal flat deposits (FA 4). This vertical repetition of facies associations indicates stacking of three separate stratigraphic units (SU) (see Figure 5.9.).

Facies Association 1: Fluvial deposits

Facies Association 1 (FA 1) occurs stratigraphically in the lower part of the Pärnu Fm, and laterally extends throughout the entire outcrop belt across 5 km. FA 1 is documented in 16 measured sections, and is especially well exposed in outcrops 1, 4 and 7 in the northeastern part of the outcrop belt, as well as in outcrop 2 in southwestern part of the study area (see Figure 5.9.). FA 1 is erosionally based, up to 3 m thick (Figure 5.10.), and comprises mainly vertically and laterally stacked, erosionally based, concave-up lenticular units of texturally and compositionally immature coarse sandstones and conglomerates (Figure 5.10B and Figure 5.16D and F). The 1-3 m thick lenticular units are fining upward and have a high width/depth ratio (see Figure 5.10A). In outcrops parallel to paleocurrent directions, the lenticular units are in most places 4-8 m wide, but in some places up to 20 m wide, maximum 150 m wide. In outcrops perpendicular to the palaeocurrent directions the width of individual lenticular units does not exceed 1-6 m. The tops of these depositional units are in most cases eroded by overlying lenticular units. As a result, in most cases only the basal, vertically stacked facies are preserved (see Figure 5.10A). In southern part of the outcrop belt, in outcrop 2, gently inclined lateral accretion sets are characteristic for FA 1, composed of small-scale trough cross-stratified sandstone (F3).

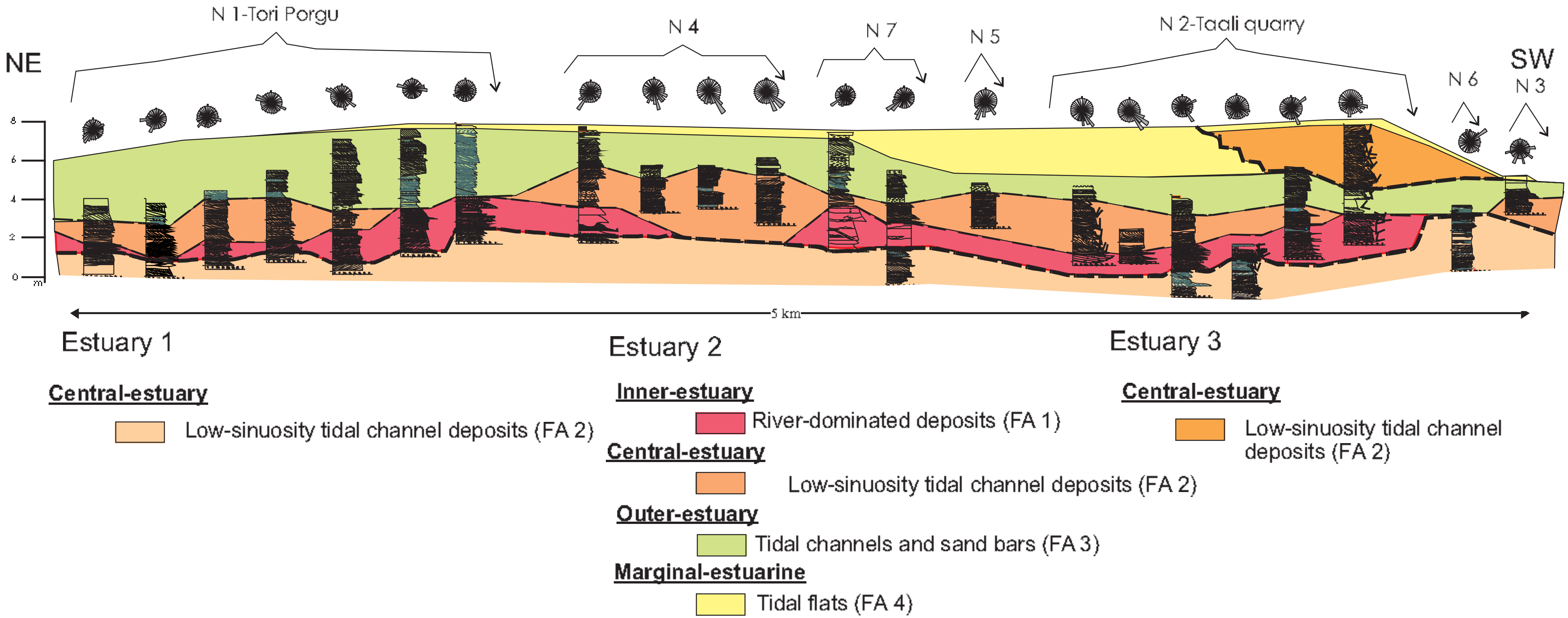


Figure 5.9. A correlation chart of the sedimentary logs derived from the outcrops of the study area. It represents lateral and vertical facies transitions through three Startigraphic Units
5.9. attēls. Atsegumu nogulumu griezumi un to korelācija pētījumu teritorijā. Griezums ilustrē slāņkopas trīsdaļīgo uzbūvi

Sandstones of FA 1 are characteristically poorly sorted and consist of fine to coarse sand and gravel, and have angular grains (Figure 5.10C). The basal erosion surfaces of the individual lenticular units are commonly overlain by mudstone and siltstone conglomerates (F1; Figure 5.10E and F) that grade upward into trough cross-stratified mudstone conglomerates (F2; Figure 5.10B, G, I), which grade into trough cross-stratified sandstones (F3, Figure 5.10C) at the top. In some places, structureless sandstones (F6; Figure 5.10H and I) and deformed sandstones (F9; Figure 5.10D) occur, and in one locality cross-stratified sandstones with irregular mud drapes (F4) are abundant. Volumetrically, FA 1 is dominated by cross-stratified sandstone (F3; 48%), cross-stratified conglomerate (F2; 27%), and structureless sandstone (F6; 19%). Mudstone and siltstone conglomerate (F1), sandstone with deformation stuctures (F9), and cross-stratified sandstones with irregular mud drapes (F4) form together the remaining 6%.

Mudstone rip-up clasts are most abundant at the bases of the erosionally-based depositional units (F1, see Figure 5.10F). Mudstone rip-up clasts also occur on toesets of cross strata (F2; see Figure 5.10G). Both mudstone and siltstone conglomerates are matrix supported and composed of crudely rounded mudstone and siltstone clasts of granule and pebble size (0.5-2 cm, rarely up to 4 cm in diameter). Matrix of the conglomerates consists of medium- to coarse-grained sandstone, similar to the under- and overlaying sandstone beds. Mudstone rip-up clasts that occur at the toesets of cross-strata (F2) are thin lenticular conglomerate units, 0.4-1.2 m thick that drape the basal scours of cross strata. The clasts are rather rounded. Mudstone conglomerate cross-sets are 4-40 cm thick (see Figure 5.10B). The lower bed contacts are usually sharp and erosional, but the upper bed contacts are gradational as F2 grades into trough cross-stratified sandstones of F3 (see Figure 5.10G). Mudstone rip-up clasts contain in places numerous fossil plant remains that are very well preserved, including sporangium (Kalamees, 1988; Figure 5.10E and F, outcrops 1, 2, 4). Due to the small size of plant fragments and the absence of fertile parts, the plants cannot be identified more exactly than that they belong to *Psilophytites* sp. and *Hostinella* sp. (Kalamees, 1988). These plant remains are abundant only in F1. In places fossil fish bones occur: in lens-shaped, more angular varieties of siltstone pebbles (*see also* Kleesment, Mark-Kurik, 1997).

The structureless sandstone (F6; Figure 5.10H, I and Figure 5.16F) is medium to coarse-grained, poorly sorted and occurs as sharp-based lenticular beds 0.2-2 m thick. Sandstone with deformation structures (F9) consists of fine-grained sandstone with overturned cross-strata, convoluted and buckled cross-strata and flame structures. Thickness of the overturned sets is 5 to 50 cm, similar to the undeformed cross set thickness in F3 and F4, where the overturned cross strata commonly occur. The deformation affects only the topmost 10-12 cm of the cross-sets. In contrast, flame structures are closely associated with F6 and form thicker, in places more than 1 m thick units (see Figure 5.10D). The latter have sharp lower contacts, whereas the lower contact of the overturned cross-stratification is gradational. In some places, occasional reactivation surfaces occur in cross strata (F3). Sandstone with mud drapes are very rare, they are found in outcrop 2, which is situated in the southern part of the outcrop belt.

Palaeocurrent measurements, derived from cross strata of F2, 3 and 4 of FA 1 show large variability between 45-230°, with most measurements between 80-140° and mean direction towards 130° (see Figure 5.10.). The erosion surfaces at the bases of depositional units dip towards 148-165°. The direction of cross-strata dip varies within individual sandstone bodies and is generally about 50°. In outcrop 2, which is situated southwards, palaeocurrent directions derived from cross strata reflect reversals in adjacent units, and are towards 60° and 210-230°.

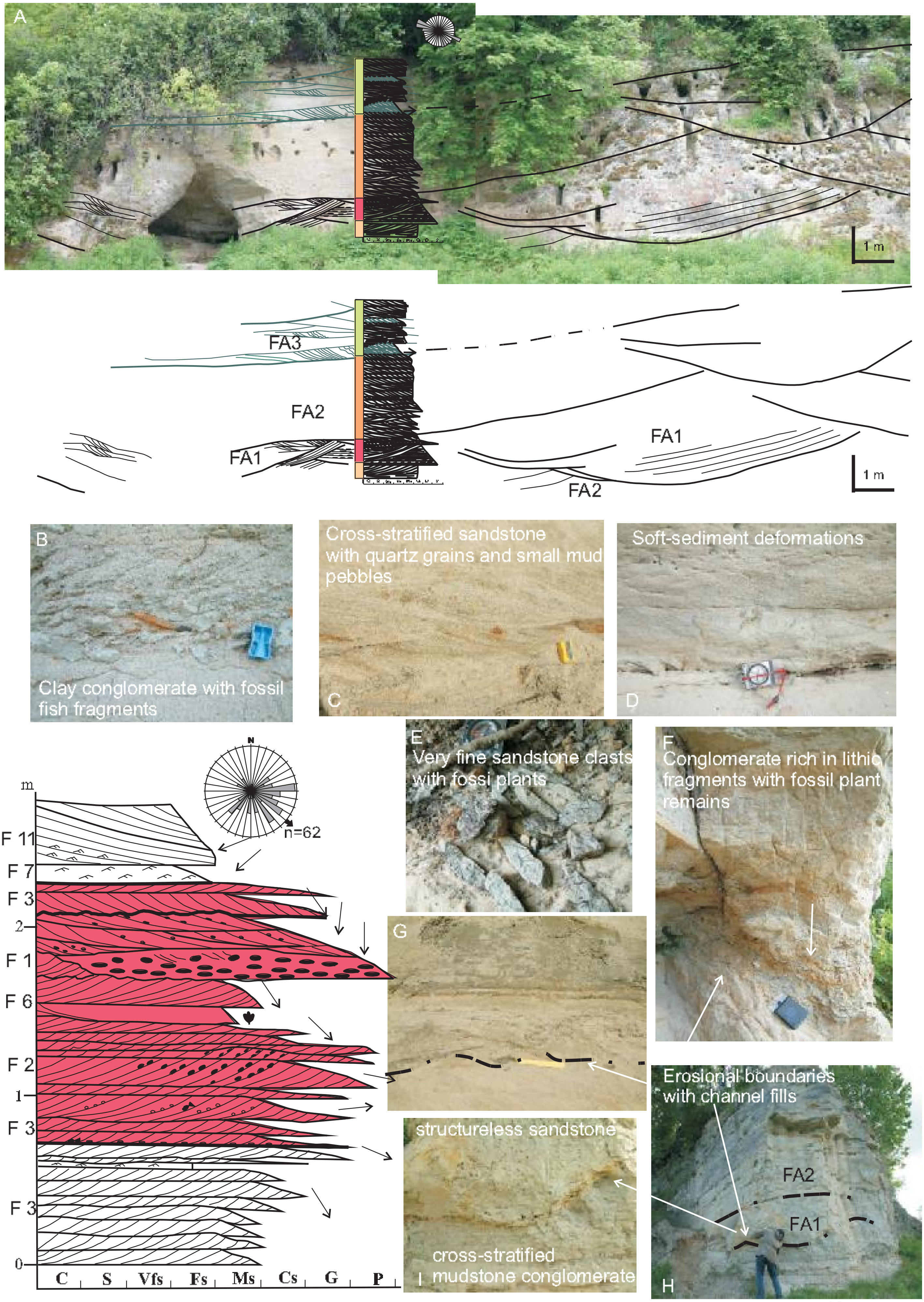


Figure 5.10. See next page
5.10. attēls. Skat. nākamo lappusi

Figure 5.10. A representative measured section (outcrop 1) illustrating the sedimentology of fluvial deposits of Facies Associations 1

Numbers by the graphic log refer to facies (see table 5.1). Palaeocurrent directions derived from the cross-strata are shown in the rose diagram. For key, see Fig. 5.11. A - Photomosaic with measured section from the fluvial channel deposits represented by stacking channel pattern with concave-up scours. All channels are erosively based with fining-upward trend (outcrop 1); B - Trough cross-stratified mudstone and siltstone conglomerate (Facies 2); C - Trough cross-stratified sandstone with quartz grains and mudstone pebbles (Facies 3); D - Sandstone with deformation structures (Facies 9); E - Fossil plant remains; F - Mudstone and siltstone conglomerate with fossil plant remains (Facies 1); G - A distinctive erosional surface associated with fluvial deposits; H - Erosional surfaces separating fluvial deposits of Facies Association 1 from tide-influenced fluvial and tidal channel deposits of Facies Association 2 above; I - A concave-up erosional surface and structureless sandstone of Facies 6. [Scale information: pen sharpener in B and C measures 3 cm; compass in D, E and F measures 7 cm; ruler in G measures 20 cm]

5.10. attēls. Fāciju asociācijas 1. – fluviālo nogulumu - reprezentatīvs griezums (1. atsegums)

Cipari pie griezuma norāda uz fāciju numuriem (skat. 5.1. tabulu). Paleostraumju virzieni pēc slīpo slānīšu krituma azimutu mērījumiem ir norādīti rozes-diagrammā (apzīmējumus skat. 5.11. attēlā. A - Fotomozaīka ar nogulumu griezumiem, kas ataino komplicētu slāņkopas uzbūvi, kuru veido vairāki fluviāli kanāli ar uz augšu vērstām pamatnēm. Visi kanāli saguļ ar izskalojuma virsmām, un nogulumu graudu izmēri tajos samazinās griezumā uz augšu (1. atsegums); B - Slīpslāņots māla un aleirolīta konglomerāts (2. Fācija); C - Slīpslāņots smilšakmens ar kvarca graudiem un māla oļiem (3. Fācija); D - Smilšakmens ar deformācijas tekstūrām (9. Fācija); E - Augu fosīlas atliekas; F - Māla un aleirolīta konglomerāts ar augu fosīlām atliekām (1. Fācija); G - Izteikta erozijas virsma, kas ir raksturīga šai fāciju asociācijai; H - Erozijas virsma, kas atdala 1. fāciju asociācijas fluviālos nogulumus no 2. fāciju asociācijas plūdmaiņu ietekmētiem nogulumiem; I - Uz augšu vērsta erozijas virsma un smilšakmens ar viendabīgu tekstūru (6. Fācija). [Informācija mērogam: zīmuļu asināmais B un C attēlos – 3 cm garš; kompass D, E un F attēlos – 7 cm garš; metramērs G attēlā – 20 cm garš]

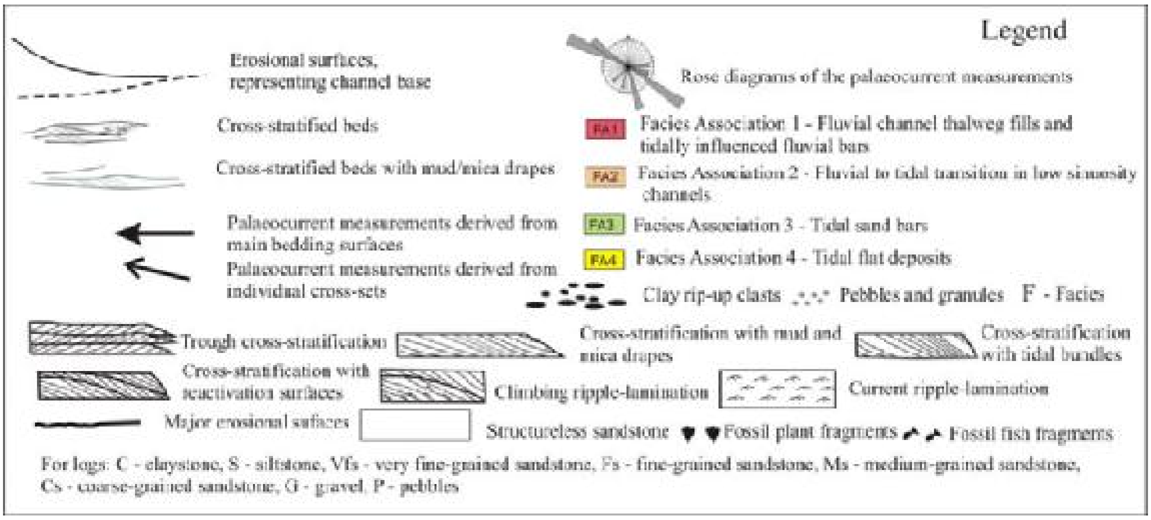


Figure 5.11. Key for the Figures 5.10. and 5.12-5.16

5.11. attēls. Leģenda attēliem 5.10. un 5.12-5.16

Interpetation:

The sharp, concave-up erosional bases, the fining upwards trends and abundance of cross-stratified sandstone with mudstone rip-up clasts (facies 1-3) indicate deposition in channels (Rahmani, 1988). Palaeo-water depth was calculated from cross-strata thickness of F2 (Table 5.2), and indicates channel flow depth of ca. 1-7 m according to Leclair and Bridge (2001) and ca. 1-4 m according to Dalrymple and Rhodes (1995). This is in agreement with the preserved thicknesses of individual channel bodies of 1-3 m, as the channel bodies have erosional boundaries and are thus erosional remnants.

The lateral and vertical amalgamation of individual erosionally-based lenticular units indicates frequent lateral migration or avulsion of channels and repeated episodes of incision and infill (Yang, Nio, 1989). The variable size of the lenticular depositional units, separated by erosion surfaces indicates various channel sizes. The amalgamated channel fills with coarse grain size, the dominance of trough-cross strata, and presence of lateral accretion sets only in one outcrop, suggests deposition in laterally mobile channels (Willis et al., 1999; Bridge et al., 2000). The deviation between the measurements of the channel scour directions and superimposed cross-strata, which is about 50°, indicates migration in relatively low sinuosity channels. The latter is further confirmed by the coarse grain size, mudstone rip-up clasts, and dominant cross stratification that indicate high-energy bedload deposition (Yang, Nio, 1989; Collinson, 1996; Miall, 1996).

Table 5.2

Calculated water depth for Facies 2, after Leclair and Bridge (2001) and Dalrymple and Rhodes (1995)

5.2. tabula

Ūdens dziļuma aprēķini fācijai 2, pēc Leklēras un Bridža (2001), kā arī Dalrimple un Rodesa (1995)

Cross-set thickness (m)	**Dune height (m),** using equation Hs=2.9(+-0.7)X Sm by Leclair and Bridge (2001)		**Water depth (m),** using equation: 3<d/hm<20 by Leclair and Bridge (2001)		**Water depth (m)** using equation: Hd=0.167h by Dalrymple and Rhodes (1995)	
	h(min)	h(max)	d(min)	d(max)	d(min)	d(max)
0.07-0.22	0.20	0.64	0.61-4,06	1.9-12.8	1.22	3.82
0.124 (mean)	0.36		4.13		2.15	
0.09 (mode)	0.26		2.99		1.55	

The mudstone and siltstone conglomerates (F1) were deposited as locally derived rip-up clast lags from high energy traction currents (Collinson, 1996). Trough cross-stratified sandstone (F3), and cross stratified sandstone with mudstone clats (F2) indicate migration in 3-D dunes (Els, Mayer, 1998). Occurrence of structureless sandstones (F6) suggests occasional hight deposition rates (Collinson, 1996), where no traction occurred due to very high sedimentation fallout rates (Miall, 1996). High deposition rates and consequent deformation by pore water escape is also seen by flame structures (Reineck, Singh, 1980). The overturning of the upper parts of the cross sets occurs as deformation due to shear stress caused by overriding traction currents on water-filled sediments (Owen, 1996).

Rare reactivation surfaces in the cross-stratified sandstones indicate that migration of bedforms was occasionally interrupted, and the bedforms were partially eroded due to changes in their migration direction or due to current reversals (Mowbray, Visser, 1984; Yang, Nio, 1989). In case of current reversals, the cross-stratification is formed by the dominant current and the reactivation surfaces are formed by the subordinate current (Visser, 1980; Willis et al., 1999). Irregular mud drapes, found only in one place (outcrop 2), further confirm rare tidal influence and mud deposition during slack water periods (Visser, 1980; Nio, Yang, 1991). Mud drapes may also form in fluvial environments during stagnant periods (Mowbray, Visser, 1984; Nichols, 1999). However the evidence of reversals in palaeocurrent directions in adjacent units derived from cross strata of the outcrop 2 also confirms occasional tidal influence to deposition of FA 1.

High degree of textural immaturity suggests relatively short transport, and lack of reworking by basinal processes. Presence of plant fragments shows a terrestrial sediment source. The coarse-grained texture and immature composition, presence of plant fragments, dominance of lenticular amalgamated small channels, and the net southeasterly palaeocurrent distribution indicate collectively deposition in fluvial channels (see also Collinson, 1996; Miall, 1992; 1994). The channels were up to 7 m deep and filled dominantly by migrating dunes, with larger-scale bedforms present in the southward (basinward) end. The tidal influence was strongest in the south (basinwards), as reactivation surfaces, occasional mud drapes and the reversals of palaceocurrent directions indicate in outcrop 2. It indicates that the fluvial flows in general were much stronger than the opposite currents. In summary, FA 1 is interpreted to be deposited in a bedload-dominated fluvial system that was occasionally influenced by tides during extreme spring cycles, as the landward tidal limit migrated occasionally upstream.

Facies Association 2: Tide-influenced fluvial and tidal channel deposits

Facies Association 2 (FA 2), 1-4 m thick, occurs at multiple stratigraphic levels across the whole 5 km wide outcrop belt and is documented in all sections. In SU 1 (the lower one), FA 2 occurs in outcrops 1, 2, 6 and 7, and is erosively overlain by FA 1, 2 and 3 of SU 2. Thickness of FA 2 increases from 1 to 3 m towards southwest. In SU 2, FA 2, 1-3 m thick, occurs throughout the entire outcrop belt, except for outcrop 6. Here FA 2 erosively overlies FA 1 and is erosively overlain by FA 3. In SU 3 FA 2, 3 m thick, occurs only in the southwestern end of the outcrop belt in outcrop 2, where it erosively overlies FA 3 of SU 2 (see Figure 5.9.).

FA 2 is dominated by trough cross-stratified sandstone (F3; up to 50 %; Figure 5.12F), cross-stratified sandstone with mud and mica drapes (F4; 18%), compound cross-stratified sandstone (F11; 14 %; Figure 5.12B and C), large scale cross-stratified sandstone (F10; 11 %; Figure 5.12H and Figure 5.13E, F, G). Less common facies are current ripple-laminated sandstone (F7; 3 %; Figure 5.12E), sigmoidal cross-stratified sandstone (F12; 2 %; Figure 5.12G), plane-parallel laminated sandstone (F5; 1 %) and trough cross-stratified mudstone conglomerate (F2; 1%). FA 2 consists of relatively poorly sorted, fine- to coarse-grained sandstone, in places with thin mudstone rip-up clast layers on the cross–strata. In the cross-section very often this immature sandstone alternates with a few decimeter thick units of mature fine-grained sandstone with rounded grains.

The deposits of FA 2 occur as erosionally-based, slightly concave-up 1-2 m thick units, or as large-scale inclined (11-17°) accretion sets, 10-50 m long and 1-3 m thick. The large-scale bedding surfaces in the accretion sets are superimposed by 10-30 cm thick cross-sets (Figure 5.12A and Figure 5.13A). The accretion sets have erosive lower and upper boundaries, and can be in places traced for more than 40 m along the outcrops. Trough cross-stratified sandstone (F3; see Figure 5.12F) that is one of the most common facies in FA 2 occurs as up to 3 m thick erosively based, lenticular units with mostly gradational upper contacts. The thickness of cross-sets is 4-50 cm.

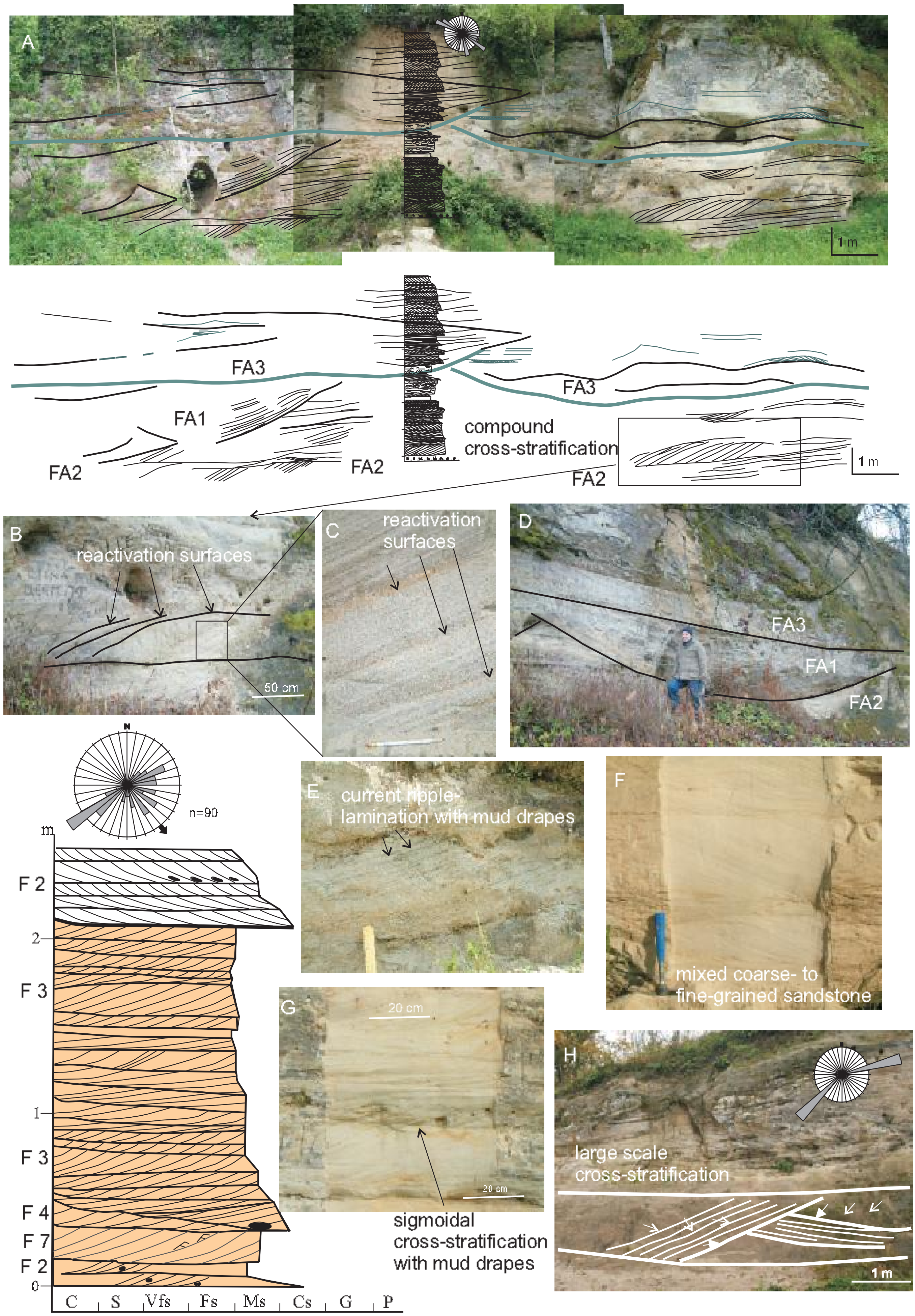


Figure 5.12. See next page
5.12. attēls. Skat. nākamo lappusi

Figure 5.12. A representative measured section (outcrop 2) illustrating the sedimentology of tide-influenced fluvial to tidal channel deposits of Facies Associations 2 from Statigraphic unit I

Numbers by the graphic log refer to facies (see Table 5.1.). Palaeocurrent directions derived from the cross-strata are shown in the rose diagram. For key, see Figure 5.11. A - Photomosaic with measured section showing tide-influenced fluvial to tidal channel deposits of Facies Associations 2, Stratigraphic unit I, repesented by simple channel thalweg fills and bars with compound cross-stratification (outcrop 1); B - Compound cross-stratified sandstone with reactivation surfaces (Facies 11); C - Enlarged fragment from photo B, illustrating reactivation surfaces (Facies 11); D - Fluvial deposits of Facies Association 1 representing channels fill, cut into tide-influenced fluvial to tidal channel deposits of Facies Associations 2; E - Trough cross-stratified sandstone with current ripple lamination, stressed by mud drapes (Facies 4 and 7); F - Fine to coarse-grained trough cross-stratified sandstone (Facies 3); G - Sigmoidal cross-stratification stressed by mud drapes (Facies 12); H - Large scale cross-stratified sandstone (Facies 10; outcrop 1). The rose diagram shows the orientation of the channel scour dip directions. Dip direction of the concave-up erosional scours at the base of two depositional units differ by 140°. The arrows correspond to dip directions of concave-up erosional scours and the cross-strata. The dip direction of cross-strata in these units mainly coincides with the dip directions of the master bedding surfaces within ca 41°. Both the master bedding surfaces, as well as cross-strata dip at rather high angles of 26-32°. It indicates bar migration, which coincides with channel floor direction, suggesting downstream-accretion. [Scale information: pencil in C measures 15 cm; part of a ruler in E measures 10 cm; hammer in F measures 30 cm]

5.12. attēls. Fāciju asociācijas 2. – plūdmaiņu ietekmēto fluviālo un plūdmaiņu nogulumu - reprezentatīvs griezums no 1. slāņkopas (1. atsegums)

Cipari pie griezuma norāda uz fāciju numuriem (skat. 5.1. tabulu). Slīpo slānīšu krituma azimutu mērījumi ir norādīti rozes-diagrammā. Citus apzīmējumus skat. 5.11. attēlā. A - Fotomozaīka ar 2. fāciju asociacijas 1. slāņkopas nogulumu griezumiem, kuri sastāv no vienkāršiem aizpildījuma kanāliem un sērēm, kam raksturīgs saliktais slīpslāņojums (1. atsegums); B - Smilšakmens ar salikto slīpslāņojumu un reaktivācijas virsmām (11. fācija); C - Palielinājums no B. attēla, ar izceltām reaktivācijas virsmām (11. fācija); D - 1. Fāciju asociācijas fluviālie nogulumi, kas sastāv no kanālu aizpildījumiem un kuri ar izteiktu diskordanci pārsedz 2. Fāciju asociācijas plūdmaiņu ietekmētos fluviālos un plūdmaiņu nogulumus; E - Slīpslāņots smilšakmens ar straumju ripsnojumu uz slīpajiem slānīšiem, kurus pasvītro māla kārtiņas (4. un 7. fācijas); F - Smalk- līdz rupjgraidains slīpslāņots smilšakmens (3. fācija); G - S-veida slīpslāņojums, ko pasvītro māla kārtiņas (12. fācija); H - Smilšakmens ar lielām slīpslāņotām sērijām (10. fācija, 1. atsegums). Rozes diagramma norāda uz slāņkopas pamatņu (kanāli) krituma azimutiem. Tā, slāņkopas pamatnēm krituma azimuti atšķiras par 140°. Bultas norāda uz slāņkopas pamatņu, kā arī slīpslāņojuma kritumu aizmutiem. Slīpslāņojuma krituma azimuti galvenokārt sakrīt ar slāņkopu pamatņu krituma azimutiem un to atšķirība ir 41° robežās. Gan galveno slīpslāņojuma virsmu, gan slīpo slānīšu krituma leņķi ir salīdzinoši lieli - 26-32°. Mērījumi norāda uz to, ka sēru augšana ir notikusi vienādā virzienā ar paša kanāla migrāciju. [Informācija mērogam: zīmulis C attēlā – 15 cm garš; metramēra daļa E attēlā – 10 cm; āmurs F attēlā – 30 cm garš]

In some places overturned and convolute cross-strata occur. These deformation structures affect only parts of cross-sets, rarely the whole sets. In most cases the upper parts of the individual cross-sets are deformed. The 1-3 m, most commonly 2-3 m thick accretion sets with superimposed low angle cross-strata, dipping down the master surfaces, form large scale and compound cross stratification of F10 and F11 (see Figure 5.12D and Figure 5.13E, F, G). The compound cross-stratified sandstones (F11) form 20 to 220 cm thick lenticular units, in which cross-sets are 5-40 cm thick. F11 consists of fine to coarse-grained sandstone draped occasionally by non-continuous mud or mica laminae. Boundaries with other facies are in most places sharp, and the lower boundaries are erosional. In compound cross strata of F11, the decimetre-scale superimposed cross sets are bounded by more gently inclined reactivation surfaces (Figures 5.12B and C). Reactivation surfaces with variable inclination (from gentle to steep) are very common in FA 2 (see Figure 5.13A). Foresets of the reactivation surfaces vary from low to high angle and dip at 9 to 28°. They are also associated with sigmoidal cross strata (F12) in very fine to fine-grained sandstone, in places rich in mica and mud drapes. A variation in the lateral spacing of reactivation surfaces is typical (see Figure 5.12G). Mud and mica drapes rarely extend to the top of the cross strata, they are truncated by reactivation surfaces in mid-foreset. Mud and mica drapes occur along the entire cross-strata, or only on the toesets (Figure 5.16E), forming an alternation of mica/mudstones and sandstones (see Figure 5.13A and Figure 5.16H). In general, mica drapes prevail to mud drapes in FA 2 in comparison to FA 3, where mud drapes are dominant. In most places the reactivation surfaces separate cross-sets into a series of sigmoidal sets, several meters wide (see Figure 5.12G). In coarser grained varieties of FA 2 cross sets show grain size decrease upward in each individual cross strata (from gravelly and coarse sand to fine or very fine sand and silt). These cross strata are bounded by more gently inclined reactivation surfaces (Figures 5.13B, C, D). Although rare, current ripple lamination is superimposed on cross-strata (F7; Figure 5.12E). Palaeocurrent directions derived from cross strata of the compound cross-stratified sandstone of F11 show unimodal distribution 195-285°, and the large-scale inclined accretion surfaces are oriented towards 255-285°. In most cases the inclined surfaces and the cross-strata dip in a similar direction. In outcrop 4, palaeocurrent directions derived from cross strata of F10 show bimodal distribution and vary between 101°-219° while the master bedding surfaces dip towards 200-240° (Figure 5.13G). Overall, the palaeocurrent directions in FA 2 group into two sectors, towards 120-160° and 200-240° and the difference in dip directions between the major channels scours and associated cross strata varies in each SU (see Figure 5.12H and Figure 5.13G).

In SU 1, FA 2 deposits change gradually from coarse-grained and mixed coarse to fine-grained sandstones in the northeast to fine-grained sandstones towards southwest (see Figure 5.12A). They occur as large-scale inclined accretion sets, 10-50 m long and up to 3 m thick, as well as concave-up 1-2 m thick units (Figure 5.12.). FA 2 of SU 1 consists mainly of F11, as well as F10 and F3. Reactivation surfaces, as well as gradual grain size decrease upward at cross-strata level is common in it, mud drapes, although quite indistinct, are also abundant. Tracing inclined accretion sets across the outcrop walls shows that they dip uniformly towards 79° and 221°, essentially parallel to the dip of superimposed cross strata oriented towards 101-111° and 222-228°. In the southern part of the outcrop belt (outcrop 6; Figure 5.16C) the cross-strata dip is directed to 33-349°. Overall palaeocurrent directions derived from cross-strata in the SU 1 show polymodal distribution and vary between 60-250°. Most of the palaeocurrents are towards 220-240° and 60-90°, and the mean direction is towards 143°.

In SU 2, the accretion sets are dominantly slightly inclined. There is no significant trend in grain size across the outcrop belt, as FA 2 here is composed of mixed coarse- to fine-grained sandstone, where sets of coarser grained and poorly sorted sandstone with fossil fish fragments alternate with finer-grained and well-sorted sandstone. Reactivation surfaces, irregular and indistinct mud drapes and especially cross strata with gradual grain size decrease upward are more common compared to SU 1. Erosionally based, lenticular units here are up to 2 m thick, and the width of the channels is commonly up to 8 m, reaching maximum of 48 m. In contrast to the underlying deposits of FA 1, the FA 2 consists of vertically stacked accretion sets, rather than stacked simple erosionally-based cross-stratified lenticular units (see Figures 5.13A and G). Palaeocurrent measurements derived from the large-scale accretion surfaces, as well as superimposed cross-sets show variable directions at an individual scale. Such individual variations, in contrast to the underlying facies, indicate the shift in dune-scale cross-strata dip relative to that of the inclined master bedding surfaces. Palaeocurrent directions derived from cross-strata in SU 2 group into two directions 120-160° and 210-240° with the mean direction towards 187° (see Figure 5.13). Most of accretion sets dip towards 207-273° while the dip of superimposed cross strata is towards 109-215°. The accretion set and superimposed cross-set dip directions vary in individual sandstone bodies, but usually exceed 50°.

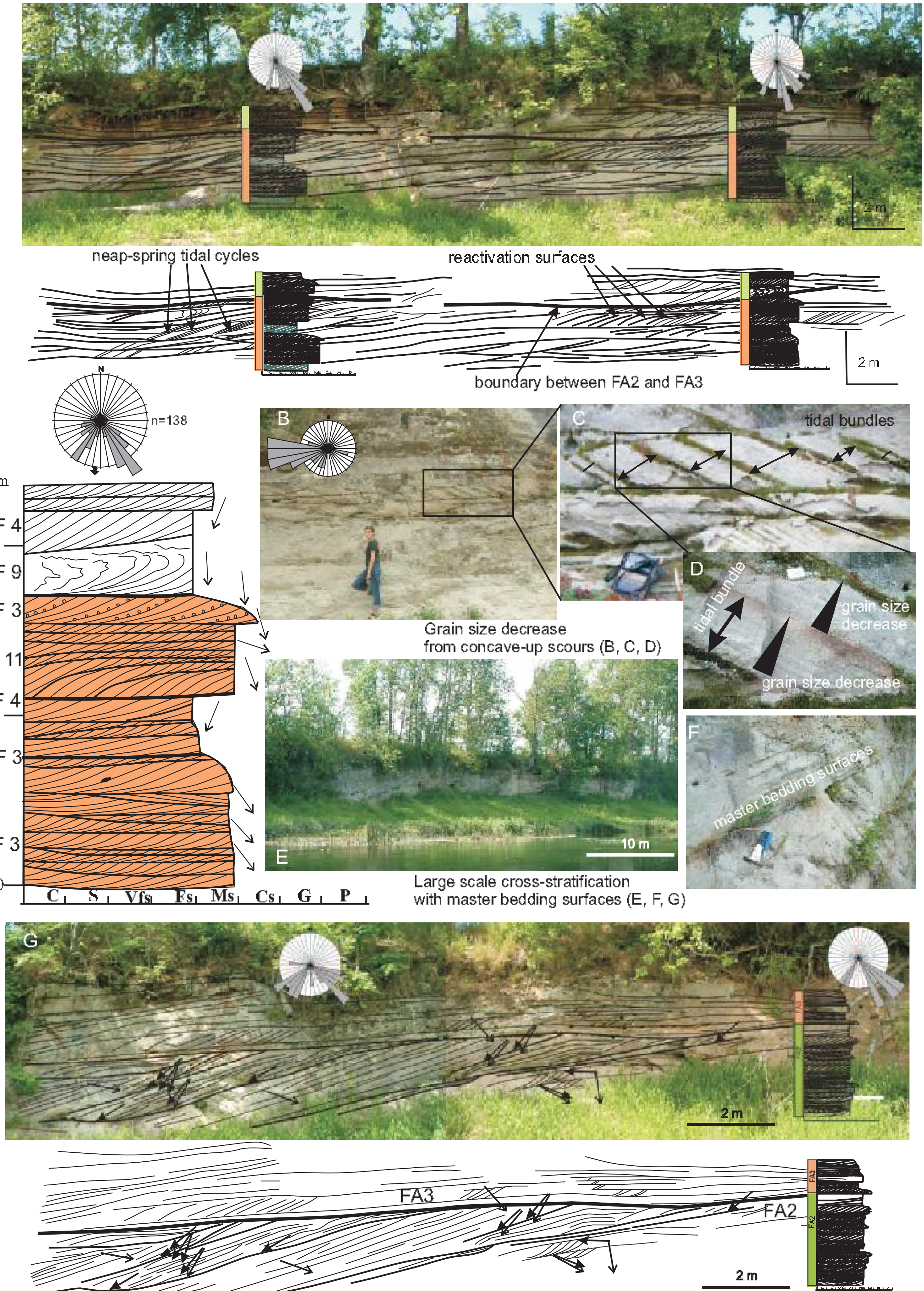


Figure 5.13. See next page
5.13. attēls. Skat. nākamo lappusi

Figure 5.13. A representative measured section (outcrop 4) illustrating the sedimentology of tide-influenced fluvial to tidal channel deposits of Facies Associations 2 from Stratigraphic unit II

Numbers by the graphic log refer to facies (see Table 5.1). Palaeocurrent directions derived from the cross-strata are shown in the rose diagram. For key, see Figure 5.11. A: Photomosaic with measured sections showing tide-influenced fluvial to tidal channel deposits of Facies Associations 2, Stratigraphic unit II repesented by gently inclined master bedding surfaces with superimposed cross-sets, containing reactivation surfaces, mud and mica drapes and tidal bundles (outcrop 4); B, C, D - Compound cross-stratified sandstone with alternation of thicker and thinner sets - tidal bundles, in which grain size systematically decreases from concave-up scours on individual strata: from coarse-grained and even gravely sand to mud, the latter is stressed by moss. The rose diagram illustrates palaeocurrent directions derived from cross-strata (outcrop 1). E, F, G - Photomosaics, illustrating Facies Association 2 represented by slightly inclined (11-17°) master bedding surfaces with superimposed cross-sets of large scale cross-stratified sandstone (Facies 10) and overall view of SW part of the outcrop 4 (E) with closer view of the inclined bedding surfaces (F). The deviation between palaeocurrent directions derived from master bedding surfaces and cross-strata is illustrated by the arrows at individual scale (see Fig. 5.11), as well as by the rose-diagram for all measurements taken. [Scale information: height of the bag in C measures 20 cm; eraser in D measures 2 cm; hammer in F measures 30 cm]

5.13. attēls. Fāciju asociācijas 2. – plūdmaiņu ietekmēto fluviālo un plūdmaiņu nogulumu - reprezentatīvs griezums no 2. slāņkopas (4. atsegums)

Cipari pie griezuma norāda uz fāciju numuriem (skat. 5.1. tabulu). Slīpo slānīšu krituma azimutu mērījumi ir norādīti rozes-diagrammā. Citus apzīmējumus skat. 5.11. attēlā. A - Fotomozaīka ar 2. Fāciju asociacijas 2. slāņkopas nogulumu griezumiem (4. atsegums). Šī slāņkopa sastāv no smilšakmens ar lēzeni krītošām noslāņojuma virsmām, un kuram raksturīgs slīpslāņojums, tajā sastop reaktivācijas virsmas, māla un vizlu kārtiņas, kā arī plūdmaiņu kārtas; B, C, D - Smilšakmens ar salikto slīpslāņojumu, kurā redzamas biezākas un plānakas slīsplāņotas kārtas – plūdmaiņu kārtas – uz kurām graudu izmēri sistemātiski samazinās virzienā uz augšu: no rupjgraudaina un pat granšaina smilšakmens līdz aleirītiskam mālam, kuru labi pasvītro sūnas atsegumu sienā. Rozes diagramma norāda uz slīpo slānīšu krituma azimutiem (1. atsegums); E, F, G - Fotomozaīka ar 2. Fāciju asociācijas atsegumu uzbūvi un nogulumu griezumu, kas sastāv no smilšakmens ar liela izmēra slīpslāņojumu un lēzeni krītošām (11-17°) noslāņojuma virsmām (10. fācija) ar kopēju 4. atseguma DR gala attēlu (E) un noslāņojumu virsmām tuvplānā (F). Bultas (skat. 5.11. att.) norāda uz noslāņojuma virsmu un slīpslaņojuma krituma azimutiem individuālos slāņos, kā arī rozes diagramma ar kopējiem mērījumiem. [Informācija mērogam: somas augstums C attēlā – 20 cm; dzēšgumija D attēlā – 2 cm; āmurs F attēlā – 30 cm]

In SU 3, FA2 occurs as laterally and vertically stacked erosionally based lenticular units present only in outcrop 2 (see Figures 5.15A and C). FA 2 here is composed of coarse-grained sandstone that grades into mixed sandstone, where coarser-grained and poorly-sorted sandstone with fossil fish fragments are present together with finer-grained and well-sorted sandstone with mud and mica drapes. Reactivation surfaces, as well as irregular and indistinct mud drapes are frequent in these deposits. Palaeocurrent measurements in SU 3 show polymodal distribution towards 105 and 235°, while the dip of the large-scale accretion surfaces is towards 280°.

Interpetation:

The erosionally based concave-up units with large scale accretion sets, cross-stratified sandstones with alternating coarse-grained poorly-sorted and fine-grained well-sorted sandstones with mud and mica drapes, abundant reactivation surfaces, compound cross-strata, sigmoidal cross-strata and decreasing grain size at individual cross strata scale indicate deposition by fluvial as well as tidal currents. The poorly-sorted coarse sandstones are similar to those in FA 1, although somewhat finer. The well-sorted sandstones indicate reworking by basinal processes.

The common large-scale accretion sets and compund cross strata indicate migration of bars or large compound dunes (see Dalrymple, Choi, 2007). Such accretion sets are also called inclined cross-bedding (Dalrymple, 1984) or double-cross stratification (Kurshs, 1975; 1992). The master surfaces represent bar accretion surfaces and the dunes migrated superimposed on the barforms. Individual accretion surfaces reflect past positions of the barforms, and internal variations within the beds reflect discharge variations of the channel system (Miall, 1992). The compound cross-stratified sandstone is interpreted to be formed by migration of 3-D dunes on tidal or tide-influenced bars or compound dunes (Dalrymple, 1984; Dalrymple, Choi, 2007). Most large-scale accretion sets indicate downcurrent migration, although some are lateral accretion sets (where difference between accretion set dip and superimposed cross strata is 60-90°). The dominant trough cross-stratified sandstone (F 3) indicated main deposition from traction currents by migration of 3-D dunes (Els, Mayer, 1998). The associated overturned cross lamina formed as a result of post-depositional deformation due to drag under overpassing currents (Collinson, 1996). Systematic grain size decrease upwards from the cross-strata scours on individual cross strata is interpreted as tidal bundles and reflects deceleration of river currents due to tides (Pontén, Plink-Björklund, 2007). The sigmoidal cross-stratified sandstone is interpreted to have formed as migrating 3-D dunes modified by tidal current reversals, and is one of the diagnostic structures for tidal origin (Kreisa, Moiola, 1986). The sigmoidal tidal bundle pattern reflects deposition from one dominant tidal current that deposits a set of cross-strata, enclosed by two relatively gently dipping, sigmoid-shaped reactivation planes formed by the subordinate current (Kreisa, Moiola, 1986; Shanley et al., 1992), and thus indicate bidirectional flows (Nio, Yang, 1991). Variations in the lateral spacing of reactivation surfaces within the cross sets indicate significant changes in the rate of bedform migration during successive tidal periods. Similarily, the variable inclination of the reactivation surfaces indicates the amount of erosion by the subordinate current from neap to spring tides (Dalrymple, Rhodes, 1995). Reactivation surfaces are absent or only weakly developed at neap tides, but become prominent, low-angle surfaces at spring tides. The mud/mica drapes indicate deposition during slack water periods (Dalrymple, 1992). The relatively scarce superimposed ripple cross-lamination and weakly abundant and indistinct mud drapes suggest subsequent erosion, weakly developed slack-water periods and/or low suspended sediment concentrations, which is typical for the fluvial-tidal transition zones (Dalrymple, Choi, 2007). However the broken-up mud drapes on cross-strata indicate that occasionally mud drapes were produced and preserved.

The significant shift in dune-scale cross-set dip relative to that of the inclined master bedding surfaces in different SUs, indicates that dunes migrated over the bar surfaces at the same direction in Unit 1, at a slightly oblique direction in Unit 3 or in cases at significantly oblique direction to the bar migration in Units 2 and 3. This suggests that deposition occurred mainly by downstream accretion in mid-stream bars and less by lateral accretion in point bars (Miall, 1994; Willis et al., 1999). The occurrence of both mid-stream and point-bars, as well as the degree of dispersion of the palaeocurrent pattern reflect somewhat the sinuosity of the channels (Miall, 1994). It indicates that deposition in FA 2 occurred in channels with higher sinuosity than those of FA 1. Only in SU 1 the dimensions of the sandstone bodies and their internal bedding architecture, as well as palaeocurrent measurements suggest mainly downstream accreting units (Miall, 1994). This indicates lower sinuosity pattern of the channels, similar to the fluvial channel deposits of FA 1.

Calculated palaeo-water depths (Table 5.3.) using cross-set thicknesses of F3 in FA 2 are ca. 0.7-5 m (Leclair, Bridge, 2001), and ca. 0.9-4 m (Dalrymple, Rhodes, 1995). These calculations are in agreement with the measured accretion set and individual channel unit thicknesses of up to 3 m.

Table 5.3

Calculated water depth for Facies 3, after Leclair and Bridge (2001) and Dalrymple and Rhodes (1995)

5.3. tabula

Ūdens dziļuma aprēķini fācijai 3, pēc Leklēras un Bridža (2001), kā arī Dalrimple un Rodesa (1995)

Cross-set thickness (m)	**Dune height (m)** Hs=2.9(+-0.7)X Sm (Leclair, Bridge, 2001)		**Water depth (m)** 3<d/hm<20 (Leclair, Bridge, 2001)		**Water depth (m)** Hd=0.167h (Dalrymple, Rhodes, 1995)	
	h (min)	h (max)	d (min)	d (max)	d (min)	d (max)
0.05-0.21	0.14	0.61	0.4-8,7	1.83-12.2	0.87	3.65
0.085 (mean)	0.25		2.8		1.47	
0.07 (mode)	0.2		2.3		1.2	

The overall bimodal palaeocurrent distribution indicates a net southeast and southwest sediment transport. The major bedforms in SUs 1 and 2 also migrated towards southeast to southwest, whereas in Unit 3 they migrated towards northwest. The direction of dune migration in SU 1 was towards northeast in outcrops 1, 2, 4 and 6, and indicates deposition in tidal currents.

In summary, the erosionally based cross-stratified units of FA 2, in which tidal signatures are abundant but variably present, grain size and sorting varies and, with mainly bar migration both landward and basinward, indicate deposition in tide-influenced fluvial to tidal channels and bars. The frequent alternation of coarser-grained and more poorly-sorted sandstones with fine-grained, well-sorted sandstones with well-rounded grains, as well as reverse palaeocurrent directions in adjacent outcrops indicate deposition from two different sediment sources – tidal (basinal) and fluvial. The presence of both fluvial and basinal flows is reflected by a difference in structure and texture or lithology: the fluvial sediment being coarser, poorly sorted, than the sediment deposited by the tidal currents. The latter together with the basinward (southward) increase in tidal activity in SU 2 indicate deposition in the transition zone from fluvial- to tidal-dominance (see Dalrymple, Choi, 2007). Such environments occur e.g. along the transition from tide-influenced fluvial to tidal channel in central estuarine zone of a tide-dominated estuary (see Dalrymple et al., 1992). Alternatively, a similar transition may occur in a deltaic distributary channel (Dalrymple, Choi, 2007).

Facies Association 3: Tidal bar deposits

Facies Association 3 (FA 3), 1-5 m thick, occurs across the whole 5 km wide outcrop belt (see Figure 5.9.). FA 3 occurs in SU 2 and forms the upper part of the outcrop in northeast and the middle part in southwest. It is documented in 20 measured sections. FA 3 is based by significant erosion surfaces, in places lined by large mudstone clasts, and erosively overlies FA 1 and FA 2 of the SU 2 (Figures 5.14A and B). FA 3 is erosively overlain by FA 2 and FA 4 of the SU 3 in southwest. FA 3 thickens from 1 to 4.5 m towards northeast.

FA 3 mainly consists of trough cross-stratified sandstone with mud/mica drapes (F4; 35 %; Figures 6.1.2.6.B, D and G), compound cross-stratified sandstone (F11; 2 3%; Fig.8), trough cross-stratified sandstone (F3; 16 %; Figure 5.14C), as well as current ripple-laminated sandstone (F7; 9 %; Figure 5.14F and Figure 5.16D) and climbing ripple-laminated sandstone (F8; 8 %; Figure 5.14E). In rare places small units of structureless sandstone (F6; 3%), plane parallel-stratified sandstone (F5; 2%), and sandstone with deformation structures (F9; 2 %; Figure 5.14G) occur. Trough cross-stratified conglomerate (F2) and sigmoidal cross-stratified sandstone (F12; Figure 5.14H) form together the remaining 2 %. FA 3 is mainly composed of mature, well-sorted fine-grained sandstone with rounded grains.

The basal erosion surface (Figure 5.14I) is lined by mudstone clasts up to 15 cm in diameter (Figures 5.14B and C) in northern part of the outcrop belt, where FA3 erosively overlies fluvial channel deposits of FA 1. In other places FA 3 erosively overlies mainly tide-influenced fluvial to tidal channel deposits of FA 2. In such places, the lower part of FA 3 consists of fine-grained sandstone with current ripples (F7), in some places lined by a thin siltstone layer (Figures 5.15B and E). In the south, the base of FA 3 is marked by a thin mudstone clast layer, coarse quartz grains and fossil fish conglomerate.

FA 3 is composed of 2-5 m thick and up to 350 m long sandbodies with low angle accretion sets that form master bedding surfaces, and are superimposed by smaller scale bedforms (Figure 5.14A). The best preserved sandbodies occur in the northernmost part of the outcrop belt (outcrop 1). Outcrop 1 provides a 3-dimensional view of these sandbodies across several hundreds of meters. The distribution of mud varies within these sandbodies of outcrop 1, as the upper parts of accretional sets are sandier, whereas more mud and mica drapes occur in the lower parts or toes of these sandbodies, thus reflecting a cleaning-upwards pattern (see Figure 5.15A). In F4, which is most typical facies of this FA, 2-10 mm thick mud or mica laminae systematically drape individual cross strata (see Figure 5.14H), and are especially significant and thick towards the toesets of cross strata, in places forming continuous, thick layers (Figure 5.14D). There are however intervals, where mud drapes are not present. The individual thickness of cross-strata changes systematically laterally along the cross sets from 1 to 10 cm. Muddy, thinner intervals composed of closely-spaced or even amalgamated mud drapes, typically pass into thicker and sandier strata within the cross-sets (Figure 5.16G). These deposits are associated with reactivation surfaces (see Figure 5.14G).

The mud drapes are typically single ones with thickness of several mm. In rare cases double mud drapes occur. In places the thickness of cross-sets varies laterally and is associated with erosion surfaces at the toesets (see Figure 5.14D). Thickness of cross sets of F4 is 3-35 cm, and the depositional units are in most places 0.3-2.5 m thick.

Current ripple-lamination (F7; see Figure 5.14F and Figure 5.15E) is common in FA 3. It commonly superimposes cross strata and shows opposite dip directions compared to the cross-strata dip directions (see Figure 5.14F). Climbing ripple laminated sandstone (F8), which is also very typical for FA 3, consists of very fine to fine-grained sandstone that is mica- or mud-rich. The thickness of cross-laminae of climbing ripple-laminated sandstone is from a few mm to 1 cm thick, the thickness of the units varies from 10 to 130 cm. Cross lamina are in most places lined by mica drapes, and in some places by mud drapes (see Figure 5.14E). In places, climbing ripple-laminated sandstone forms thicker (up to 1-1.5 m) and more extensive (about 10 m) units, where the angle of climb changes repetitively upwards from vertical to low angle. The bounding surfaces in F8 dip in the opposite directions to the ripple lamina.

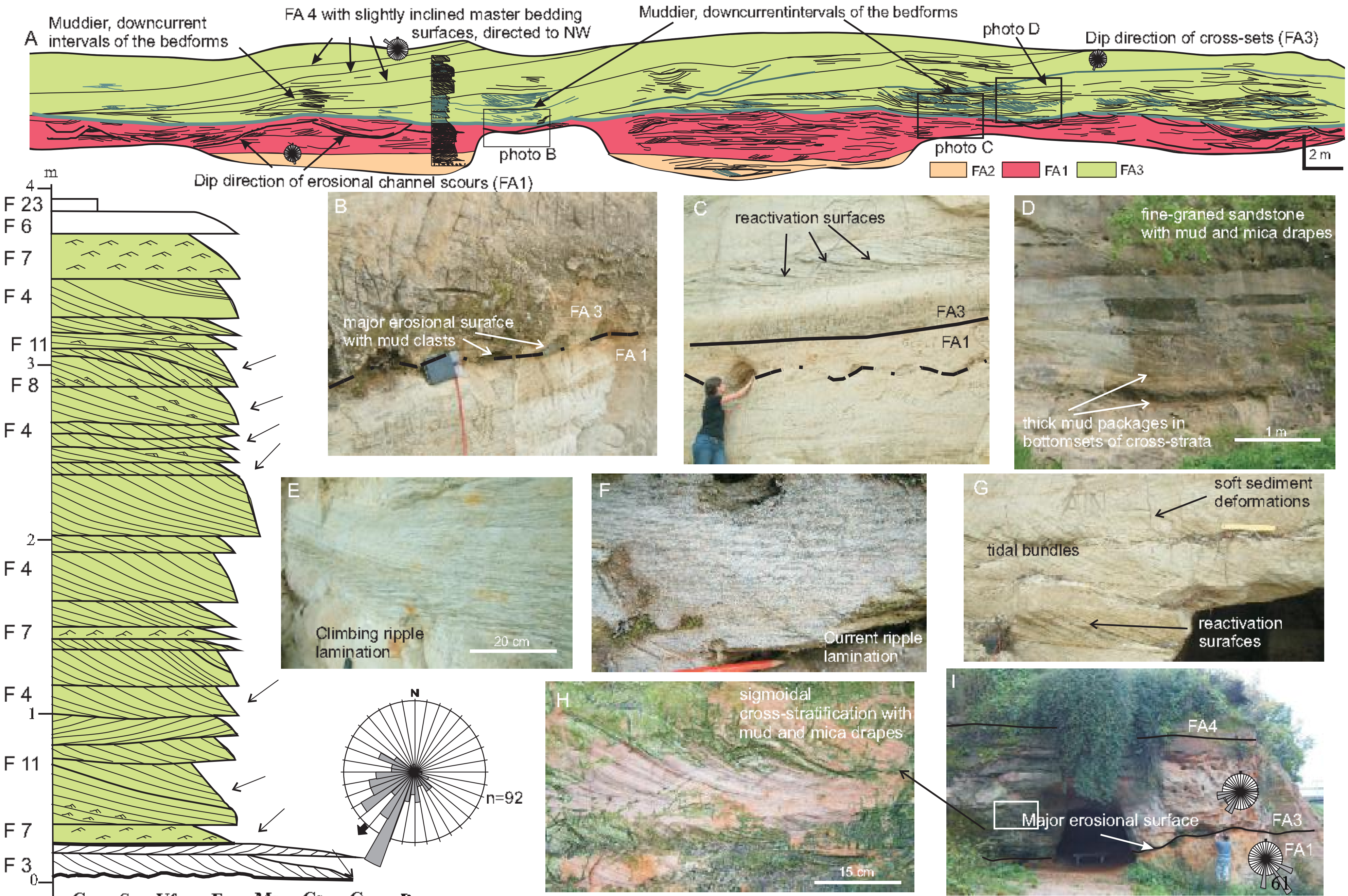


Figure 5.14. See next page. 5.14. attēls. Skat. nākamo lappusi

Figure 5.14. A representative measured section (outcrop 1) illustrating the sedimentology of tidal bar deposits of Facies Associations 3

Numbers by the graphic log refer to facies (see table 5.1.). Palaeocurrent directions derived from the cross-strata are shown in the rose diagram. For key, see Figure 5.11. A - Photomosaic with measured sections showing tidal bar deposits, illustrating master bedding surfaces of Facies Associations 3 (in green) that overly erosively channelized units of Facies Association 1 (in purple). The channel scours of Facies Association 1 are oriented towards SSE, whereas master bedding surfaces of Facies Association 3 to NW. Note, that the upcurrent parts of the larges-scale microform are sandier, whereas more mud and mica drapes appear in the downcurrent parts of this bar (outcrop 1); B - A boundary between Facies Association 1 and Facies Association 3 stressed by a distinct erosional surface with clay clasts up to 15 cm in diameter at the tidal bars basal surfaces; C - Variably-grained trough cross-stratified sandstones (Facies 3) of Facies Associaion 1 overlain erosively by fine-grained deposits (Facies 4) of Facies Association 3, rich in reactivation surfaces, mud and mica drapes and tidal bundles; D - Trough cross-stratified sandstones (Facies 4) with abundant mud and mica drapes; E - Climbing ripple-laminated sandstone (Facies 8); F) Current ripple-laminated sandstone (Facies 7); G - Fine-grained cross-stratified sandstone (Facies 4) with soft sediment deformations and reactivation surfaces; H - Sigmoidal cross-stratified sandstone with distinct mud and mica drapes (Facies 12); I - The southernmost end of the outcrop 1: a major erosional surface between Facies Association 3 and Facies Association 1 marks a bottom of a large macroform. This erosional surface is traced at a distance of more than 350 m along the outcrop. [Scale information: compass in B measures 7 cm; part of the pen in F measures 5 cm; ruler in G measures 20 cm]

5.14. attēls. Fāciju asociācijas 3. – plūdmaiņu sēru nogulumu - reprezentatīvs griezums (1. atsegums)

Cipari pie griezuma norāda uz fāciju numuriem (skat. 5.1. tabulu). Slīpo slānīšu krituma azimutu mērījumi ir norādīti rozes-diagrammā. Citus apzīmējumus skat. 5.11. attēlā. A - Fotomozaīka ar 3. fāciju asociācijas nogulumu griezumiem, kuras slāņkopa sastāv no smilšakmens ar lēzeni krītošām noslāņojuma virsmām (zaļā krāsā), kas ar erozijas virsmu pārsedz 1. fāciju asociācijas fluviālo kanālu nogulumus (sarkanā krāsā; 1. atsegums). 1. fāciju asociācijas fluviālo kanālu pamatnes ir orientētas uz DDA, bet 3. fāciju asociācijas noslāņojuma virsmas uz ZR. Jāatzīmē, ka slāņkopas augšējā daļa ir smilšaināka, bet vairāk māla un vizlu kārtiņas koncentrējas zemākās slāņkopas daļās (1. atsegums); B - Robeža starp 1. fāciju asociāciju un 3. fāciiju asociāciju, kuru pasvītro izteikta erozijas virsma un līdz 15 cm diametrā lieli māla oļi; C - 1. fāciju asociācijas dažādgraudainais smilšakmens (3. fācija), kuru ar erozijas virsmu pārsedz 3. fāciju asociācijas smalkgraudainais smilšakmens (4. fācija), kurā sastop daudz reaktivācijas virsmu, māla un vizlu kārtiņu un plūdmaiņu kārtas; D - Slīpslāņots smilšakmens ar māla un vizlu kārtiņām (4. fācija); E - Smilšakmens ar kāpjoša ripsnojuma tekstūru (8. fācija); F - Smilšakmens ar straumju ripsnojuma tekstūru (7. fācija); G - Smalkgaudains slīpslāņots smilšakmens ar deformāciju tekstūrām un reaktivācijas virsmām (4. fācija); H - Smilšakmens ar S-veida slīpslāņojumu un izteiktām māla un vizlu kārtiņām (12. fācija); I - 1. atseguma pats dienvidu gals: izteikta erozijas virsma, kas atdala 1. fāciju asociāciju no 3. fāciju asociācijas un iezīmē lielas makroformas pamatni, kuru var izsekot 350 m garumā visā atseguma joslā. [Informācija mērogam: kompass B attēlā – 7 cm; zīmuļa daļa F attēlā – 5 cm; metramērs G attēlā – 20 cm; āmurs F attēlā – 30 cm garš]

The overall trend of palaeocurrent directions for this FA derived from cross-strata and current ripple lamination is in range of 140-310°, with the mean direction to 210° and most of currents in range of 200 to 290° (Figure 5.14.). The dip directions of large-scale accretion sets or master bedding surfaces differ from paleocurrent directions derived from superimposed cross-strata and cross-lamina by 5° to 180°. In outcrop 1, the palaeocurrent directions derived from master bedding surfaces are to 303° and the dip of superimposed bedforms is directed towards 195-225°. The general orientation of the large-scale accretion sets varies across the outcrop belt. In outcrop 2, the the accretion sets are oriented towards 183°, 42° and 65°; in outcrops 4 and 7 towards 135° and 145°, in outcrop 6 towards 170°, and in outcrop 1 and 5 towards 303° and 349°.

Interpetation:

The common and systematic occurrence of mud and mica drapes, frequent reactivation surfaces, occurrence of tidal bundles in form of sand and mud couplets in cross-strata and cross-laminae, as well as the repeated changes in the angle of climb in sandstone with climbing ripple-laminated structure indicate deposition by tidal currents. The erosive base, the large-scale accretion sets, landward and basinward oriented dip directions of master bedding surfaces between adjacent outcrops, as well as the large size of the accretion sets and the large angle between the master bedding surfaces and the superimposed bedforms suggests deposition in large tidal sand bars in a subtidal environment (see Dalrymple et al., 1992). The mature nature of the sandstone, compared to FAs 1 and 2, indicates a sediment source different from FA1 or considerable reworking of sediments. The tidal bars are thus interpreted to be accumulated from marine, rather than fluvial derived sediment. The superimposed trough cross-strata indicate that 3-D dunes migrated across the tidal bar master surfaces.

The dip of the master bedding surfaces (accretion sets) indicates major bedform migration towards north that is landward in outcrops 1 and 5, and towards south that is basinward in outcrops 2, 4, 6 and 7. Such reversals of bedform directions in adjacent outcrops reflect the role of both flood and ebb currents in sediment transport and formation of these major bedforms. The northerly and southerly migrating bars may also indicate that flood and ebb currents used slightly different paths, as is commonly the case in tidal environments (Dalrymple, 1992). The overall palaeocurrent distribution of FA3 derived from cross-strata and ripple cross-lamination indicates southeast and southwest small-scale bedform migration. The variation of palaeocurrent directions between the master bedding surfaces and the superimposed bedforms within individual sandbodies indicates both downstream and lateral accretion of the tidal bars. The persistent southward dip of the superimposed smaller scale bedforms (3-D dunes and ripples) indicates that they were mainly formed by ebb-dominated flows. Landward oriented cross-strata and ripples with mud drapes, as well as abundant reactivation surfaces, record the action of flood tides.

The dimension of the preserved accretion sets indicate that the bars were at least 3-9 m high and more than 300 m wide. As tidal bars tend to accrete to sea level in their highest parts, they indicate a water depth of up to 9 m. Calculated water depth from cross strata of palaeochannels of facies 4 (Table 5.4.) indicates similar water depths, ca. 1.5-9 m (Leclair, Bridge, 2001) and ca. 1.5-7 m (Dalrymple, Rhodes, 1995).

Table 5.4

Calculated water depth for Facies 4, after Leclair and Bridge (2001) and Dalrymple and Rhodes (1995)

5.4. tabula

Ūdens dziļuma aprēķini fācijai 3, pēc Leklēras un Bridža (2001), kā arī Dalrimple un Rodesa (1995)

Cross-set thickness (m)	**Dune height (m)** Hs=2.9(+-0.7)X Sm (Leclair, Bridge, 2001)		**Water depth (m)** 3<d/hm<20 (Leclair, Bridge, 2001)		**Water depth (m)** Hd=0,167h (Dalrymple, Rhodes, 1995)	
	h(min)	h(max)	d(min)	d(max)	d(min)	d(max)
0.08-0.39	0.23	1.13	0.7-4.64	3.39-22.6	1.39	6.77
0.16 (mean)	0.46		5.34		2.78	
0.15 (mode)	0.43		4.9		2.57	

Couplets of alternating sand and mud are interpreted as tidal bundles formed during one single tidal cycle, with muds deposited during the slack water periods (Nio, Yang, 1991). A systematic changes in bundle thickness along the cross-sets, as well as changes in angle of climb in climbing ripple successions indicate neap-to-spring cyclicity (see Nio, Yang, 1991; Dalrymple, 1992; Dalrymple, Rhodes, 1995).

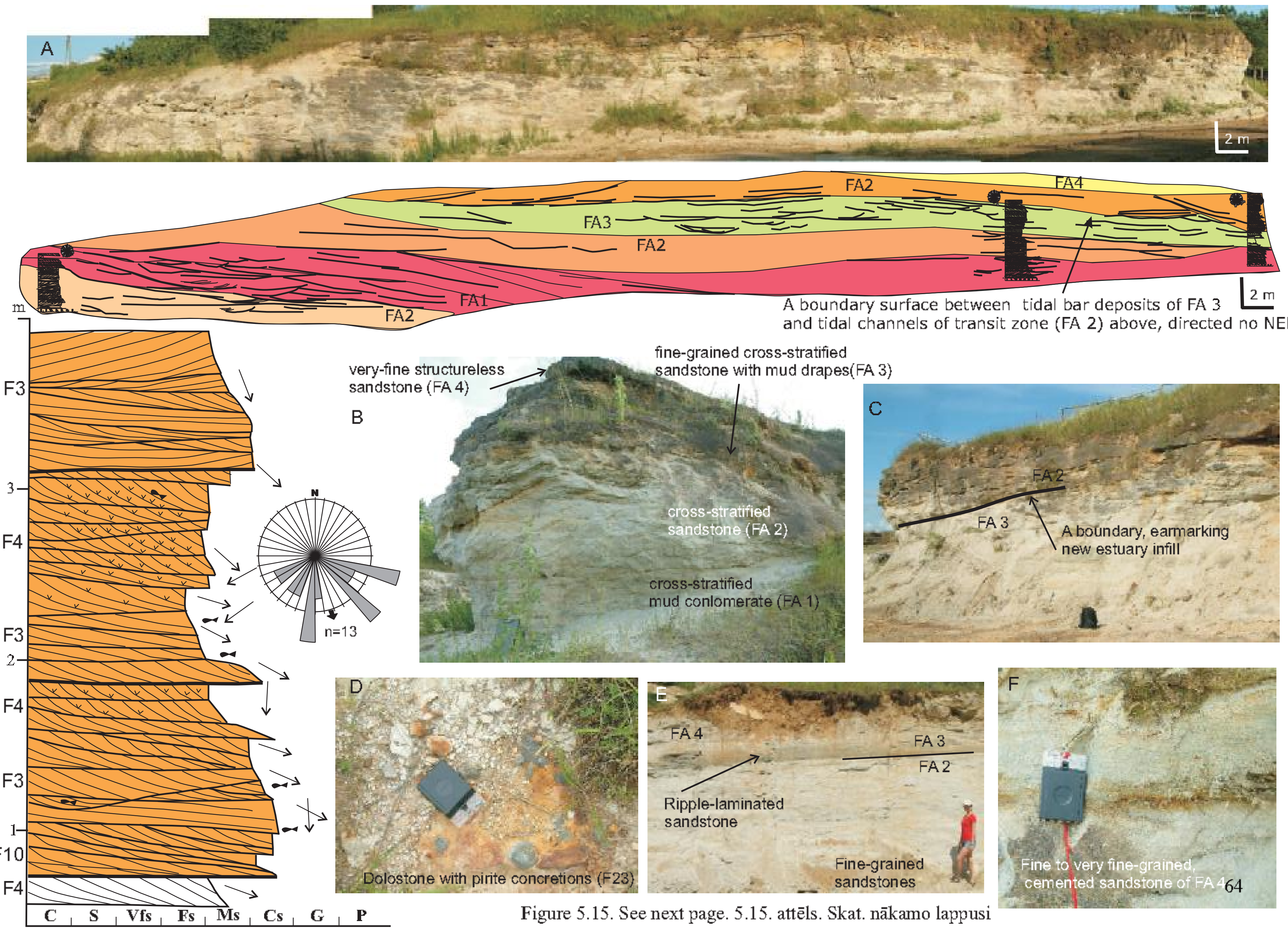


Figure 5.15. See next page. 5.15. attēls. Skat. nākamo lappusi

Figure 5.15. A representative measured section (outcrop 2) illustrating the sedimentology of tide-influenced fluvial to tidal channel deposits of Facies Associations 2 from Stratigraphic unit III

Numbers by the graphic log refer to facies (see Table 5.1.). Palaeocurrent directions derived from the cross-strata are shown in the rose diagram. For key, see Figure 5.11. A - Photomosaic with measured sections showing the fluvial deposits of Facies Association 1, tide-influenced fluvial to tidal channel deposits of Facies Associations 2, tidal bar deposits of Facies Association 3 and tidal flat deposits of Facies Association 4 (outcrop 2); B - Structureless sandstone (Facies 6) on the upper part of the outcrop representing tidal flat deposits of Facies Association 4; C - A surface earmarking boundary of fluvial to tidal channel deposits of Facies Associations 2 of the Stratigraphic unit III; D - Structureless dolostone (Facies 13) with pirite concretion from the top of the outcrop; E - Fine to very fine-grained sandstone with current ripple-lamination earmarking the boundary of Facies Association 3, very common for the studied outcrops; F - Current ripple-laminated sandstone (Facies 7) and structureless sandstone (Facies 6) representing tidal flat deposits of Facies Association 4. [Scale information: height of the bag in C measures 50 cm; compass in D and F measures 7 cm]

5.15. attēls. Fāciju asociācijas 2. – plūdmaiņu ietekmēto fluviālo un plūdmaiņu nogulumu - reprezentatīvs griezums no 3. slāņkopas (2. atsegums)

Cipari pie griezuma norāda uz fāciju numuriem (skat. 5.1. tabulu). Slīpo slānīšu krituma azimutu mērījumi ir norādīti rozes-diagrammā. Citus apzīmējumus skat. 5.11. attēlā. A - Fotomozaīka ar nogulumu griezumiem, kurus veido 1. Fāciju asociācijas fluviālie nogulumi, 2. Fāciju asociācijas plūdmaiņu ietekmētie fluviālie un plūdmaiņu nogulumi, 3. Fāciju asociācijas plūdmaiņu sēru nogulumi un 4. Fāciju asociācijas plūdmaiņu līdzenumu nogulumi (2. atsegums); B - Smilšakmens ar viendabīgu tekstūru (6. fācija) atseguma augšdaļā, kas veido plūdmaiņu līdzenumu fāciju asociāciju; C - Erozijas virsma, no kuras uz augšu sastop 3. slāņkopas plūdmaiņu ietekmētos fluviālos un plūdmaiņu nogulumus; D - Dolomīts ar viendabīgu tekstūru (13. fācija) ar pirīta konkrēcijām, no atsegumu augšdaļas; E - Smalk- un ļoti smalkgraudains smilšakmens ar straumju ripsnojuma tekstūru 3. fāciju asociācijas pamatnē, kas ir plaši izplatīts atsegumu teritorijā; F - Smilšakmens ar straumju ripsnojumu (7. fācija) un smilšakmens ar viendabīgu tekstūru (6. fācija), no kuriem sastāv 4. fāciju asociācijas plūdmaiņu līdzenumu nogulumi. [Informācija mērogam: somas augstums C attēlā – 50 cm; kompass D un F attēlos – 7 cm]

The presence of mainly single mud drapes, as well as the lower inclinations of associated reactivation surfaces in comparison to underlying and overlying FA 2, indicate that subordinate currents were stronger and possibly eroded the second mud drape as well as significantly modified the bedforms (Visser, 1980; Dalrymple, Rhodes, 1995). The lack of the mud drapes in some places and their preservation only in toesets of cross-strata, as association with erosion surfaces at the toesets is also interpreted to be due to erosion by increased current activity e.g. during spring tides (Rahmani, 1988). Climbing ripple laminated sandstone formed by migration of ripples in conditions of high suspended sediment load, causing high sedimentation rates (Reineck, Singh, 1980). The repetitive vertical changes of the angle of climb indicate cyclic changes in the ratio between sedimentation rates and downstream ripple migration rates, caused by periodically waning and accelerating tidal currents (see Lannier, Tessier, 1998). Tidal bars are interpreted to be formed in the outer-estuarine zone of tide-dominated estuaries, where sediment is derived by tidal currents (Dalrymple et al., 1992; Dalrymple, Choi, 2007). The landward, seaward as well as lateral migration of the tidal bars and lack of thick muddy deposits, as well as the lack of inclined heterolithic strata are considered characteristic for the outer estuarine zone (Dalrymple, Choi, 2007). The fine-grained texture of sandstone, the presence of mud drapes and relatively high proportion of mud in succession of FA 3 in comparison to other parts of the section, suggests that the deposition took place in the inner (proximal) reaches of the outer estuary. Tidal bars also occur in the delta front of the tide-dominated and influenced deltas (Willis et al., 1999; Willis, 2005; Dalrymple, Choi, 2007; Pontén, Plink-

Björklund 2007; Tanavsuu-Milkieviciene, Plink-Björklund, 2009). However, the delta-front tidal bars consist of river-derived rather than marine-derived material, and display systematic basinward migration (Dalrymple, Choi, 2007).

Facies Association 4: Marginal tidal flat deposits

Facies Association 4, documented in 12 measured sections, is up to 1 m thick and occurs at the top of the outcrops, almost across the whole outcrop belt (see Figure 5.9.). In places it is rather poorly exposed and eroded. FA 4 belongs to the SU 3 in the southwestern part of the outcrop belt, where it overlies FA 2 of SU 2 (Outcrop 2, see Fig. 5.9). Further northeast FA 4 belongs to the SU 2 and overlies FA 3 of the SU 2 with a gradational contact.

FA 4 consists of structureless sandstone (F6; 48 %; Figure 6.1.2.7.B) current ripple-laminated sandstone (F7; 28 %; Figure 5.15F), and plane-parallel laminated and structureless dolostone (F23; 22 %; Figure 5.15D). Climbing ripple laminated sandstone (Facies 8), plane-parallel laminated sandstone (F5), and cross-stratified sandstone with mud/mica drapes (Facies 4) form the remaining 2 % (see Figure 5.15). Sandstone is very fine- to fine-grained, well-sorted, and has rounded grains.

FA 4 occurs as a fining- and thinning-upward succession, usually few dm thick with upward increasing interbedding of mudstone and structureless sandstone. The structureless sandstone (F6) is very fine- to fine-grained, well-sorted and cemented, 9-25 cm thick. The current ripple laminated sandstone (F7) consists of very fine- to fine-grained sand material, which is mud- and mica-rich. The thickness of cross-lamina is several mm, the thickness of depositional units varies from 2 to 70 cm. The individual cross-laminae are in most places lined by mud or mica drapes. The contacts with other facies are sharp and gradational, rarely erosive (see Figure 5.15B).

F7 does not form separate depositional units, it occurs normally as interlayers and in toesets of cross-stratified sandstones (F4) with ripple migration directions opposite to the direction of cross-sets. Cross-stratification of F4 and cross-lamination of F7 is lined by mud, but mainly by mica drapes. Mud laminae are commonly non-continuous. The very topmost parts of FA 4 display gradual vertical transition from fine sandstone to very fine sandstone with dolomitic cement, which progressively amalgamates and grades upward into dolostone (see Figure 5.15D). Plane-parallel laminated and structureless dolostone (F23) has cryptocrystalline texture. It is mainly structureless and platy, only in places it is plane-parallel laminated. Pyrite concretions occur on bedding surfaces of F23. Plane-parallel laminated and structureless dolostone forms small, tabular units, 5-25 cm thick. F23 has mainly gradational and sharp contacts with the underlying facies. The dolomitic beds occur exceptionally on the topmost part of the studied outcrops and are in many places eroded.

FA 4 is closely related to FA 3 in SU 2 and to FA 2 in SU 3. Both facies associations form an overall upward-fining succession, where cross-bedded sands with tidal bundles and mud drapes grade into interbedded sands and muds, capped by dolomitic deposits (see Figure 5.15E).

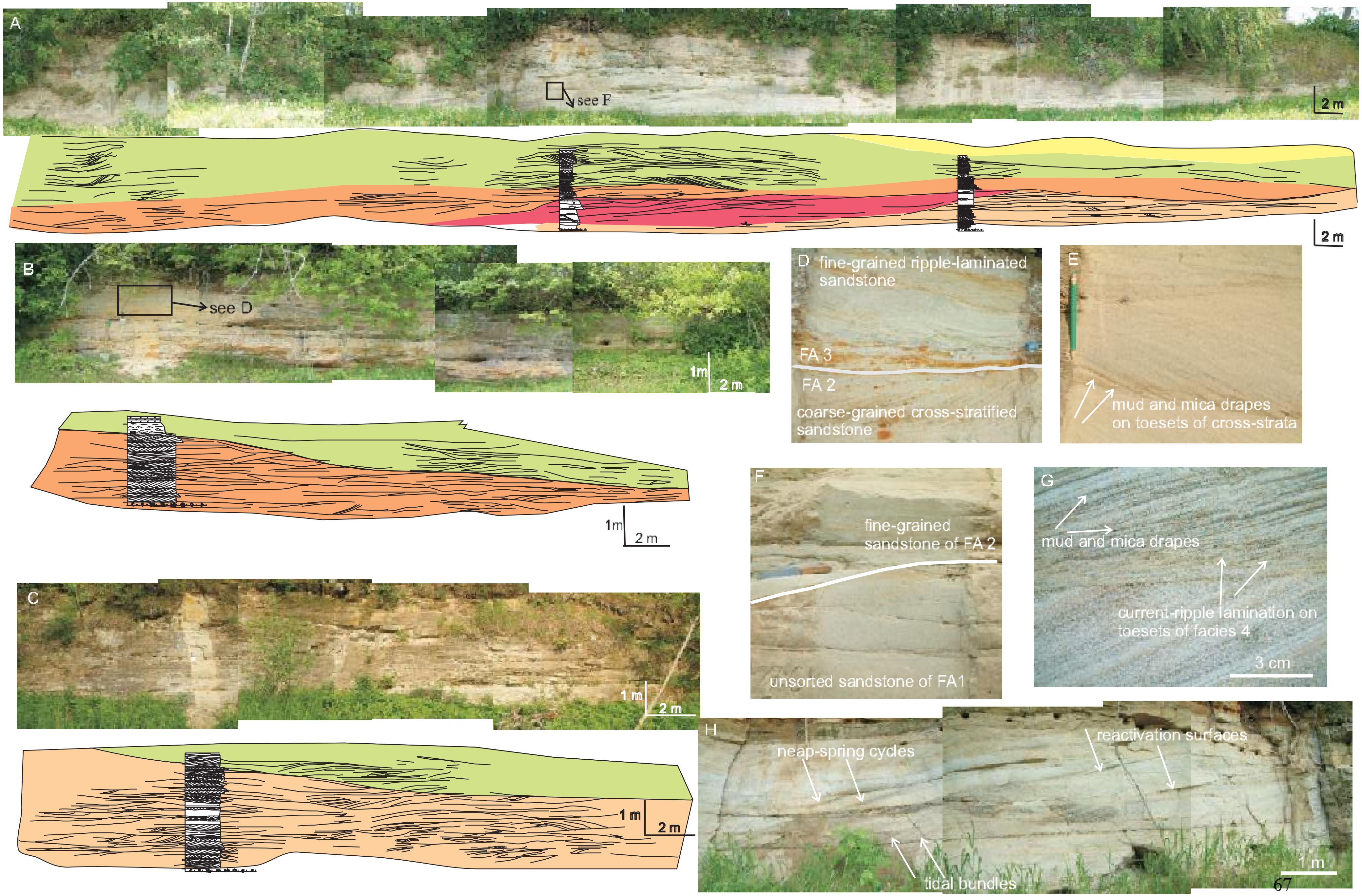


Figure 5.16. See next page. 5.16. attēls. Skat. nākamo lappusi

Figure 5.16

A - Photomosaic with measured sections showing fluvial deposits of Facies Association 1, tide-influenced fluvial to tidal channel deposits of Facies Associations 2 (Stratigraphic units I and II), tidal bar deposits of Facies Association 3 and tidal flat deposits of Facies Association 4 (outcrop 7); B - Photomosaic with a measured section from the tide-influenced fluvial to tidal channel deposits of Facies Associations 2 (Startigraphic unit II) and tidal bar deposits of Facies Association 3 (outcrop 5); C - Photomosaic with a measured section from the tide-influenced fluvial to tidal channel deposits of Facies Associations 2 (Stratigraphic unit I) and tidal bar deposits of Facies Association 3 (outcrop 6); D - Enlarged fragment from Figure B, illustrating the boundary between coarse-grained sandstones of Facies Association 1 and of fine-grained ripple-laminated sandstones of Facies Association 3; E - Trough cross-stratified sandstone with mud drapes on toesets of cross-strata (Facies 4); F - Enlarged fragment from Figure A, illustrating the boundary between unsorted fluvial sandstone of Facies Association 1 and finer-grained sandstone of tide-influenced fluvial to tidal channel deposits of Facies Association 2; G - Cross-stratified sandstone with mud and mica drapes. Current ripple-lamination occurs on toesets of cross-strata; H - Photomosaic of tide-influenced fluvial to tidal channel deposits of Facies Associations 2 and tidal bar deposits of Facies Association 3, illustrating several signatures of tidal influence to the deposition: tidal bundles, neap-spring cycles, mud and mica drapes and reactivation surfaces [Scale information: pen sharpener in D measures 3 cm; pencil in E measures 12 cm; knife in F measures 30 cm]. For key, see Figure 5.11

5.16. attēls

A - Fotomozaīka ar nogulumu griezumiem, kuru veido 1. fāciju asociācijas fluviālie nogulumi, 2. fāciju asociācijas plūdmaiņu ietekmētie fluviālie un plūdmaiņu nogulumi (1. un 2. slāņkopas), 3. fāciju asociācijas plūdmaiņu sēru nogulumi un 4. Fāciju asociācijas plūdmaiņu līdzenumu nogulumi (7. atsegums); B - Fotomozaīka ar nogulumu griezumiem, kuru veido 2. fāciju asociācijas plūdmaiņu ietekmētie fluviālie un plūdmaiņu nogulumi (2. slāņkopa) un 3. fāciju asociācijas plūdmaiņu sēru nogulumi (5. atsegums); C - Fotomozaīka ar nogulumu griezumiem, kuru veido 2. fāciju asociācijas plūdmaiņu ietekmētie fluviālie un plūdmaiņu nogulumi (1. slāņkopa) un 3. fāciju asociācijas plūdmaiņu sēru nogulumi (6. atsegums); D - Palielinājums no B attēla, kas norāda uz robežu starp 1. fāciju asociācijas rupjgraudaniem smilšakmeņiem ar 3. fāciju asociācijas smalkgraudainiem smilšakmeņiem ar ripsnojuma tekstūru; E - Slīpslāņots smilšakmens ar māla kārtiņām uz slīpo slānīšu apakšas (4. fācija); F - Palielinājums no A attēla, kas norāda uz robežu starp 1. fāciju asociācijas vāji šķirotiem fluviāliem smilšakmeņiem un 2. fāciju asociācijas plūdmaiņu ietekmētiem fluviāliem un plūdmaiņu smalkgraudainākiem smilšakmeņiem; G - Slīpslāņots smilšakmens ar māla un vizlu kārtiņām. Straumju ripsnojuma pazīmes redzamas uz slīpo slānīšu pamatnēm; H - Fotomozaīka ar nogulumu griezumiem, kuru veido 2. fāciju asociācijas plūdmaiņu ietekmētie fluviālie un plūdmaiņu nogulumi (2. slāņkopa) un 3. fāciju asociācijas plūdmaiņu sēru nogulumi, kuros redzamas vairākas plūdmaiņu pazīmes: plūdmaiņu kārtas, plūdmaiņu cikli, māla un vizlu kārtiņas un reaktivācijas virsmas [Informācija mērogam: zīmuļu asināmais D attēlā - 3 cm garš; zīmulis E attēlā - 12 cm garš; nazis F attēlā - 30 cm garš]. Apzīmējumus skat. 5.11. attēlā

Interpretation:

The fine sand grain size, the ripple-cross lamination and the mudstone interbeds in plane-parallel laminated sandstones, together with the common mud and mica drapes indicate deposition in lower energy tidal currents, compared to the other facies associations. Plane-parallel laminated sandstone can be fromed by unidirectional (Collinson, 1996), bidirectional or even oscillatory currents by movement of plain beds. The close association with F7, as well as the presence of mud and mica drapes on the parallel laminae suggest tidal origin of this facies. The sandstone of F6 may only appear structureless due to the similar-size (small) sand grains, as it locally displays faint lamination. Because of its fine-grained texture it is interpreted to be deposited as plane beds in lower flow regime. Current ripples that occur on the toes of the dune lee-sides can form as back-flow ripple sets deposited on the toesets of the dunes from unidirectional currents, or may be deposited from the subordinate current of assymetrical tidal currents.

The plane-parallel laminated and structureless dolostone is interpreted to have formed in a low energy environment by carbonate precipitation (Reineck, Singh, 1980). The precipitation of carbonate deposits if compared with siliciclastics usually takes place in normal to high-salinity conditions. The replacement of carbonate mud or consolidated limestones with dolomite (dolomitization) usually takes place in arid climate settings, in lagoonal, tidal flat or sabkha environment, as well as in restricted to open epicontinental sea basins (Purser et al., 1994).

The fine-grained composition of the deposits and abundance of dolostones, the close association with the uppermost part of the tidal-fluvial transition (FA2) and inner part of outer-estuarine deposits (FA3), as well as the overall occurrence at the top of the succession, suggest deposition on marginal tidal flats (Dalrymple, Rhodes, 1995).

5.1.3 Depositional model and estuary evolution

Vertical and lateral facies transitions derived from the outcrop data, the comparison with the adjacent deposits of Pärnu Fm, derived from the drill-core data, as well as the overall stacking pattern of the succession indicate that the Pärnu Fm in the study area is formed in tide-dominated estuary in a shallow, epicontinental sea. The estuarine succession consists of three erosionally based vertically stacked packages (see Figure 5.9. and Figure 5.17.). The lower part of the studied section shows progradational character, as the tide-influenced fluvial and tidal channel deposits (FA2 of SU1) at the bottom pass to fluvial deposits (FA1 of SU2). Upwards they pass retrogradationally to tide-influenced fluvial and tidal channel deposits (FA2 of SU 2) and further to tidal bars (FA3 of SU2). This part of succession is overlain by aggradational tidal-flat deposits (FA4 of SU2), which in turn are overlain by tide-influenced fluvial and tidal channel deposits (FA2 of SU3) at the very southern end of the outcrop belt. In such manner the studied succession of the Pärnu Fm indicates a few steps in change in balance between sedimentation and relative sea-level rise.

The lowermost of the three vertically stacked tide-dominated estuarine units (SU 1, see Figure 6.1.2.1. and Figure 6.1.3.1.) is an erosional remnant and reaches thickness of only few m, whereas the overlying unit (SU 2) is 6-7 m thick and represents estuary fill, while the topmost unit in southern part of the outcrop belt (SU 3) is also an erosional remnant, up to 3.5 m thick and is spatially limited.

Estuary in this study is defined according to Dalrymple et al. (1992), as "*a seaward portion of the drowned incised valley system, which receives sediments from both fluvial and marine sources and which contains facies influenced by tide, wave and fluvial processes*". According to Dalrymple et al. (1992), tide-dominated estuaries are divided into three main zones on the basis of energy distribution: 1) inner zone characterized by a single relatively straight to meandering channel, becoming straight again towards the opening of the estuary; 2) middle zone, developed by a braided system and dominated by multiple channels and shoals; 3) outer zone with tidal sand bars. The inner part of the fluvial-tidal transition zone is river dominated with a net seaward transport of sediments, which inner end coincides with the "tidal limit". The main volume of estuarine deposits, such as tide-influenced fluvial and tidal channel deposits (FA2; SU1, 2 and 3), and tidal bars (FA3, SU2), was accumulated in tidal-fluvial transition zone and inner reaches of the outer estuary (see Figure 5.17.). The muddy and mixed tidal flat deposits (FA4) occur at the top of the FA2 and FA3 of SU2 and SU3. Fluvial deposits (FA1) are found at the bottom of the SU2 and form the base of the retrogaradtional succession of SU2.

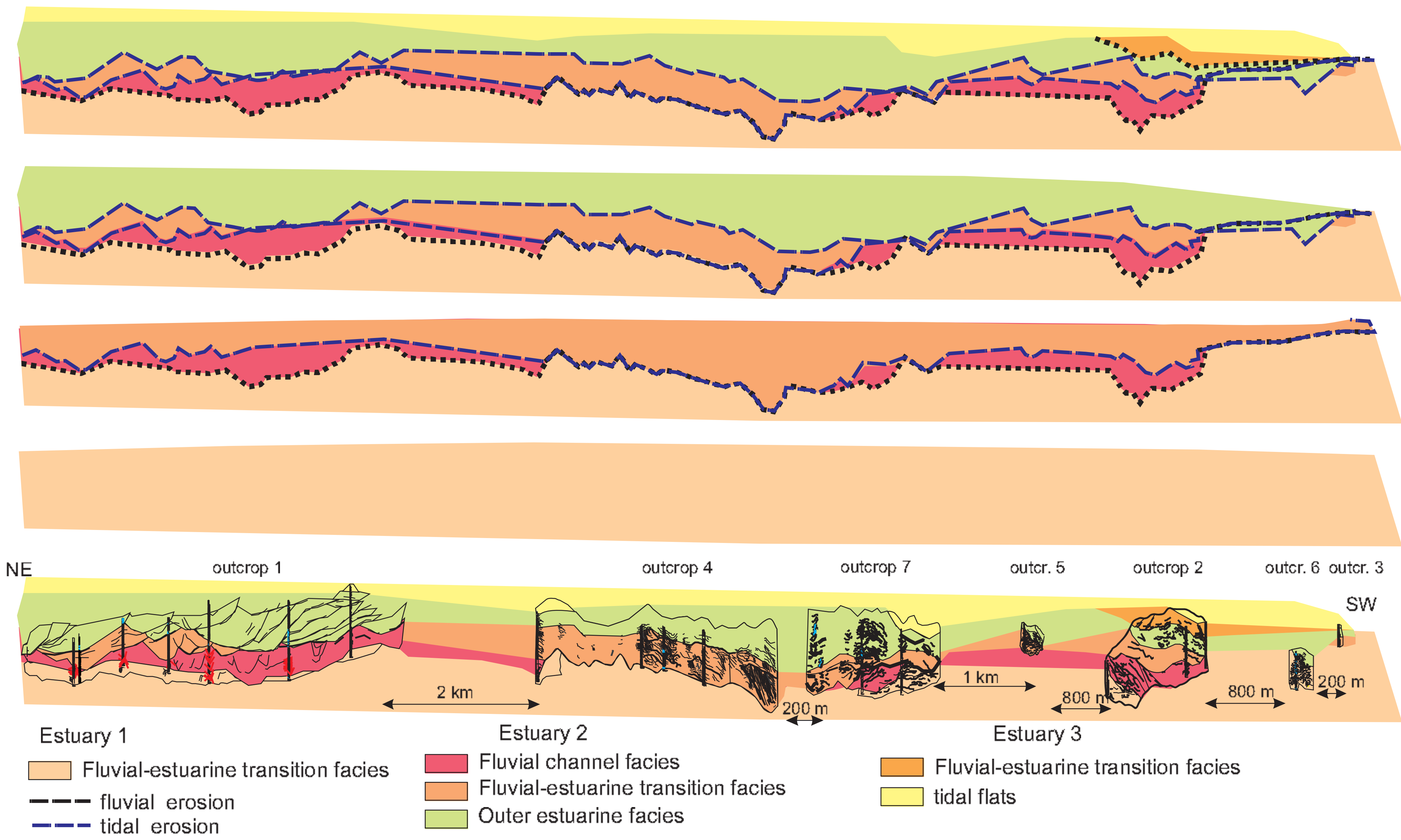


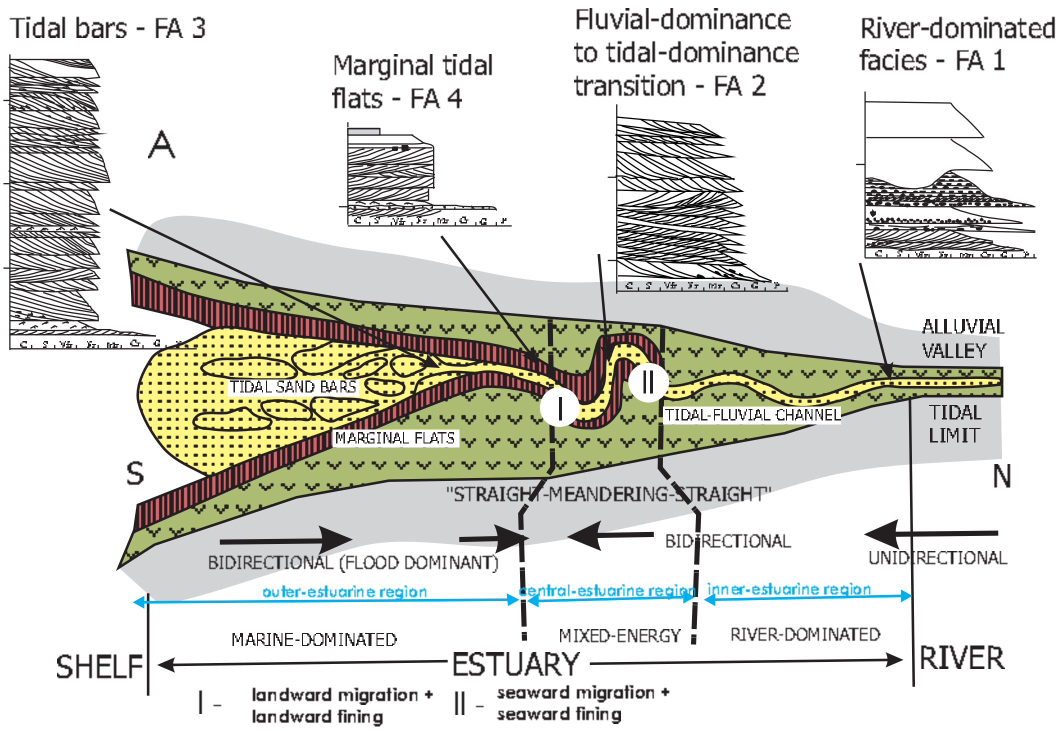
Figure 5.17. A correlation chart of the outcrops derived from the study area representing lateral and vertical facies assocation transitions through three Startigraphic Units

5.17. attēls. Atsegumu nogulumu griezumu korelāciju panelis pētījumu teritorijā, kas norāda uz fāciju asociāciju laterālām un vertikālām izmaiņām griezumā un trīsdaļīgo slāņkopas uzbūvi

During estuary evolution, the depositional architecture and estuary morphology changed throughout SU 1 to SU 3 (Figure 5.18.). The occurrence of tide-influenced fluvial and tidal channel deposits (FA2) at the bottom of the Pärnu succession (SU1) in the study area indicates deposition in central part of an estuary, in fluvial-tidal transition zone. The lack of fluvial deposits at the base of SU 1, the small thickness of the unit, and the erosive character of the overlying upper members of SU2 suggest fluvial erosion (see Figure 5.12.). The landward orientation of smaller bedforms in number of outcrops suggests deposition in the outer straight reach of the central estuary which is tidally dominated due to strong flood-tidal currents (Dalrymple et al., 1992). The channel contained alternate, bank-attached bars and some mid-channel bars. The bars were oriented both landwards and basinwards, which indicates that flood and ebb-dominant channels coexisted. As the water and transported sediment follow a circuitous route in and out of the estuary, it leads to spatially varied systems, where some parts of the estuary are flood-dominated and other parts ebb-dominated (Ashley, 1990; see Figure 5.17.). Moreover, main flows of flood and ebb tides in the estuaries and deltaic distributaries are steered toward different directions because of the Coriolis effect, resulting in different current strength and depositional patterns along the two different banks (Daidu, 2012). In addition, the morphological irregularities that exist because of the presence of channel meanders and elongate tidal bars, which are slightly oblique to the flow, create localized areas of ebb- and flood-directed residual movement of sediment (Dalrymple et al., 2012).

The meandering part of the "straight-meandering-straight" zone of the central estuary marking the lowest energy portion of the system and being in position of net bedload convergence, where grain sizes in the channel become finer towards this area from both directions according to Dalrymple and Choi (2007), is missing in SU1 (see Figure 5.18). The location of the turbidity maximum is not stable, as it depends on the magnitude of river discharge and the tidal phase (Van den Berg et al., 2007). At low river discharge the maximum mud concentration and deposition of mud does not take place in the area of flood and ebb channels, but is related to the upstream meandering reach. Another explanation could be that the lack of the bedload convergence zone of an estuary, where the turbidity maximum zone occurs, can be explained by the erosion by stronger tidal currents removing evidence of fine-grained deposits, and re-depositing large mud clasts, which are observed on the boundary with underlying deposits of SU2. The hypersyncronous (tidal range and tidal currents increase landward) character of tidal currents is responsible for sediment removal from the central estuary zone. This is similar to distributary mouth bar area of tide-dominated and tide-influenced deltas of BDB, where sediment fines seaward from the distributary channels (Plink-Björklund, 2012).

The complete axial facies succession of SU 2 reflects the tripartite depositional realm: from fluvial at the base through mixed fluvial-tidal to tidal at the top, indicating a transgression during its formation. Erosively based SU 2 cuts into deposits of SU 1 and is composed of fluvial deposits (FA 1) at the base of the unit. It suggests fluvial erosion followed by significant sediment input, which is explained by the major erosional surface, channel scours at the base and coarse grain-size and immature material of the channel fill deposits at the base of this unit. The compositional nature and channelized pattern of the lower part of the succession supports interpretation that fluvially-dominated deposits of inner estuary form the base of SU 2 in the study area (see Figure 5.10.). Lateral migration of channels and increased fluvial discharge caused frequent erosion of the marginal tidal flats, as the floor of these channels consist of a patchy lag of mud pebbles derived from erosion of the banks.



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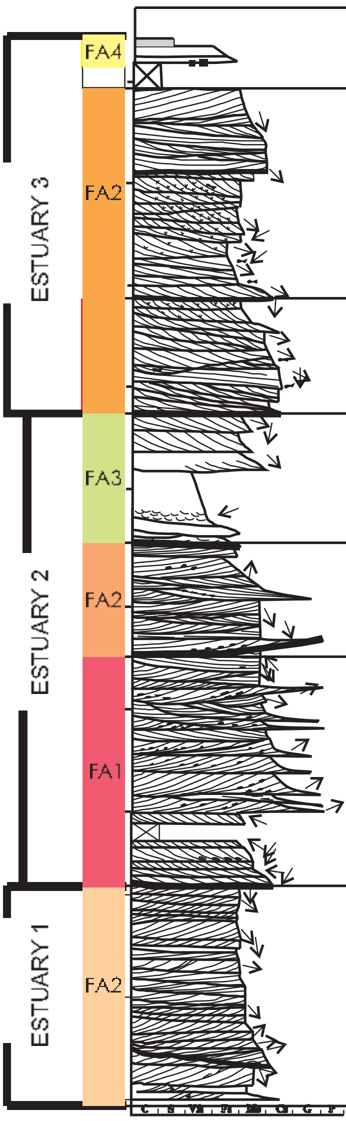

Unit	DESCRIPTION	DEPOSITIONAL ENVIRONMENT		
Estuary 3, FA4	Fine- to very fine sandstone and dolomite, structureless or plane parallel-stratified.	Mixed energy, tidal influence increase	Tidal flats	Marginal estuary: tide-dominated
Estuary 3, FA2	Medium- to fine sandstone cross-stratified, mud and mica drapes abundant, other tidal signatures typical		Inner estuary channels and bars in mixed fluvial-tidal zone, increased tidal influence	Central estuary: Transition to tidal dominance
Estuary 3, FA2	Trough-cross stratified sandstone, poorly sorted, fish remains abundant	River dominated ?	Tide-influenced fluvial channels and bars	Inner-estuary: Fluvial dominance, some tidal influence
Estuary 2, FA3	Fine-grained sandstone, trough cross-stratified with ripple lamination, many typical mud and mica drapes	Tide dominated	Outer estuary channels and bars, tidally dominated	Outer estuary: tide-dominated
Estuary 2, FA2	Trough cross-stratified sandstone, fine to medium grained	Mixed energy	Inner estuary channels and bars in mixed fluvial-tidal zone	Central estuary: Transition to tidal dominance
Estuary 2, FA1	Erosively based, clay conglomerate and trough-cross stratified sandstone, poorly sorted, plant remains abundant Tidal signatures very scarce	River-dominated	Tide-influenced fluvial channels and bars, occasionally tidal influenced	Inner-estuary: Fluvial dominance
Estuary 1, FA2	Trough cross-stratified sandstone, fine to medium grained, superimposed on dipping master bedding surfaces. Tidal signatures abundant	Mixed energy	Inner estuary channels and bars in mixed fluvial-tidal zone, increased tidal influence	Central estuary

Figure 5.18. See next page
5.18. attēls. Skat. nākamo lappusi

Figure 5.18. Pārnu FM in the outcrop area is interpreted to be deposited in fluvial channels of inner estuary, fluvial-estuarine transition zone of the central estuary, tidal bars of outer estuary and marginal tidal flats
A - estuary model redrawn from Dalrymple, Choi (2007); B – a whole representative measured section of the deposits in the outcrop belt (for keys, see Figure 5.11.)

5.18. Attēls. Pērnavas svīta atsegumu teritorijā tiek interpretēta kā estuāra iekšējās daļas fluviālu kanālu, estuāra centrālās daļās fluviālo-plūdmaiņu apstākļu pārejas zonas, kā arī estuāra ārējās daļas plūdmaiņu sēru un piegulošo plūdmaiņu līdzenumu veidojums
A - estuāra modelis pārzīmēts no Dalrimple, Choi (2007); B – nogulumu slāņkopas kopgriezums no atsegumu teritorijas (apzīmējumus skat. 5.11. att.)

The overall southerly orientation of palaeocurrent directions with abundant current reversals at the bed scale in southern end of the study area is interpreted as fluvial and hence seaward-directed currents are stronger than landward-directed flows (Dalrymple, Choi, 2007). The fluvial processes governed channel dynamics and only during some episodes the flood current penetrated in the seaward part of this zone. These currents were however weak and of short duration and unable to scour channels.

Overlying compositionally immature, coarse to fine-grained sandstones alternating with relatively mature cross-stratified sandstones (FA 2) suggest that they are deposited in fluvial-tidal transition zone (Dalrymple et al., 1992; Richards, 1994; see Figure 5.13.). In the most downstream part of the fluvial-tidal zone the presence of sedimentary intervals of coarser sand may be an important diagnostic feature for this zone (Van den Berg et al., 2007). The alternation of fluvial and tide dominated deposits in the section typical for FA 2, suggests the fluctuation in time of fluvial and tidal influence related to river discharge or tides, and is typical for fluvial-tidal transition zone (Van den Berg et al., 2007).

Since more bedforms appear to be deposited as lateral accretion units, in comparison to FA 2 deposits of the SU1, the succession is interpreted to be formed in more sinuous channels close to the outer straight reach of the "straight-meandering-straight" zone of the central estuary (Dalrymple et al., 1992; Billeaud et al., 2007; Burningham, 2008). This is reconfirmed by the stronger presence of tidal activity to deposition in more southerly situated outcrops of the study area, since the tidal modulations here are more significant. The channels show a repetitive pattern of channel bends and tidal bars (Figure 5.19). The flood and ebb channels are separated by tidal bars consisting of or a complex series of bars separated from each other by one or more by oblique channels, called swatchways (Robinson 1960). The lack of tidal flat deposits that would have flanked the main channels, indicate erosion by the currents.

The gradual increase of tidal activity upwards is characteristic for SU 2, reflecting deposition in strongly tide-dominated channels and bars (FA3) of outer estuary flanked by tidal flats (FA4). In the outer part of tide-dominated estuaries, the ebb- and flood-dominant channels form a mutually evasive system of channels that are separated by elongate tidal bars (Dalrymple et al., 2012; see Figure 5.14.). These large bars, therefore, form a linear or very gently curved "bar chain" (Dalrymple et al., 1990), that are separated by swatchways, that dissect the bar chain and connect the ebb and flood channels (Dalrymple et al., 2012). Ebb-dominated channels form the bulk of the succession. Flood-dominated channels are invariably discontinuous, terminating headward into sand bars. They are separated from the main ebb channel by tidal bars that attaches to other bar (Dalrymple et al., 2012; see Figure 5.19.). The tidal flat deposits covering the tidal bars indicate aggradation and gradual infill of estuary.

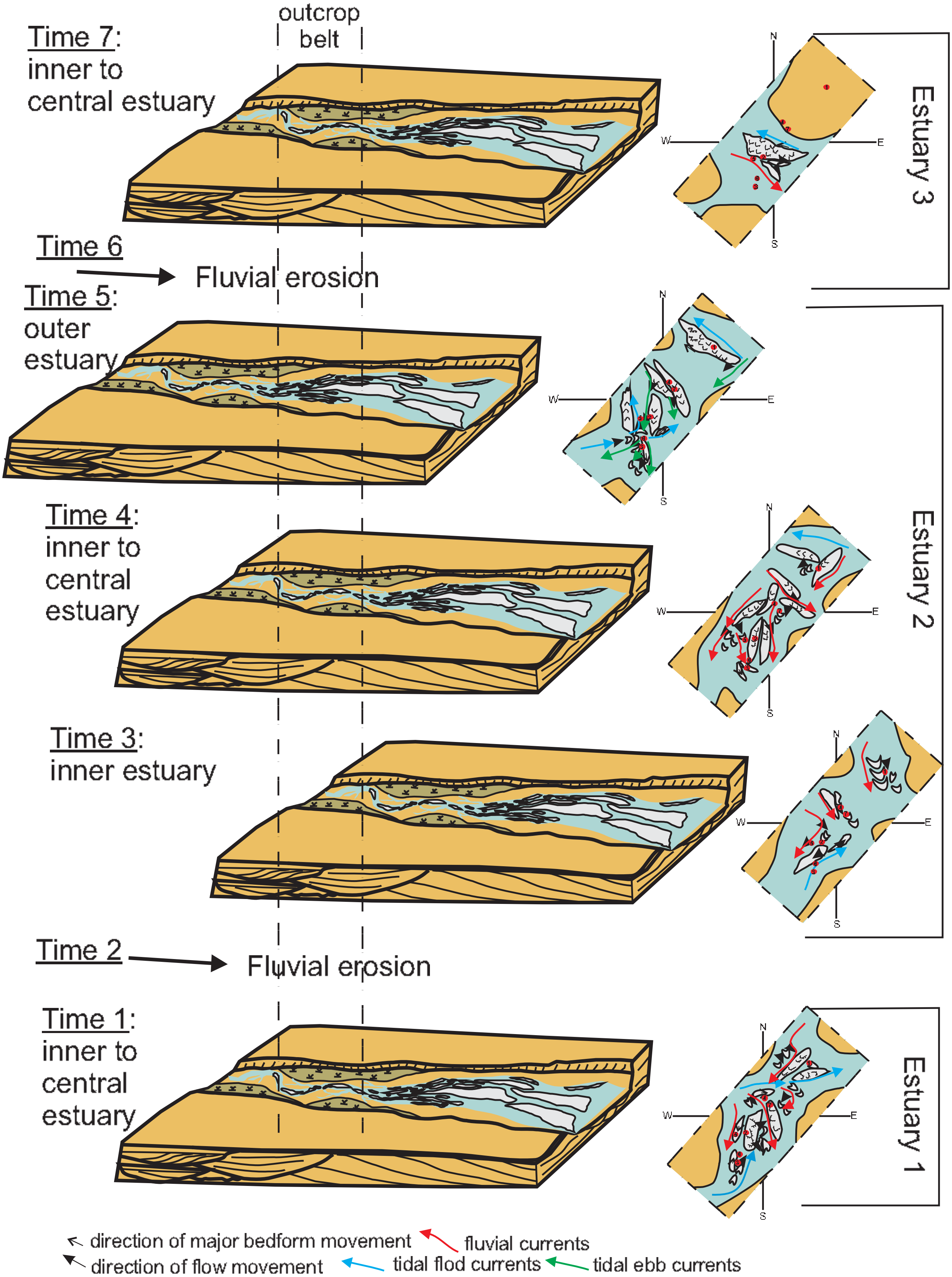

Figure 5.19. See next page
5.19. attēls. Skat. nākamo lappusi

Figure 5.19. Block diagram with hypothetical reconstruction of major morphological elements and associated bedforms, illustrated in different time maps reflecting various stages in the estuary evolution
On the right hand side time slices illustrate the major bedform (bars) and smaller scale bedform (dunes) orientation reconstructed from palaeocurrent data and bedform architectural elements in all outcrop belt

5.19. attēls. Bloka diagramma ar galveno morfoloģisko elementu un formu hipotētisku rekonstrukciju, kas attēlota dažādās "laika plāksnēs", un norāda uz estuāra attīstību vairākās fāzēs
Labajā pusē ir norādītas atsevišķas „laika plāksnes", kas ilustrē lielo un mazo paleoformu orientāciju baseinā, kas ir rekonstruētas pēc slāņu virsmu un slīpslāņojumu krituma azimutu datiem un slāņkopas uzbūves īpatnībām visā atsegumu joslā

According to Dalrymple et al. (2012), the width of the mudflats that flank the channels of estuary become broader in a seaward direction, which probably explains the presence of the tidal flat deposits associated with FA 3 and lacking in association with FA 2 in SU 2, where it is eroded and present as mud clasts.

The occurrence of which exactly part of outer estuary (distal to proximal) deposits of FA 3 of Pärnu succession belong to, can be debatable, as according to Dalrymple et al., 1992, the outer estuary zone is charaterized by coarse-grained elongate sand ridges, where mud drapes are not abundant because of erosion by subsequent currents. The FA 3 of SU 2 is interpreted to be formed in the proximal reaches of the outer estuary, where large bars are abundant, but elongate sand ridges are absent. According to Dalrymple et al. (2012) this area that lies landward of the elongate sand ridges, consists of fine to very fine sand and occupies the zone of strongest tidal currents. Mud drapes are most common in relatively sheltered areas, and especially in the troughs of the tidal bars (Dalrymple et al., 2012), which is typical for FA 3, since this is the muddiest part of all section. Dune cross-bedding is also most common in the transition to the elongate tidal ridges, because this is the area where grains are coarse enough to support dunes (Dalrymple et al., 2012).

The sediment in tide-dominated estuaries is typically coarsest at its mouth and head, and finest in the vicinity of the bedload convergence (Dalrymple et al., 1990). Thus, sediment in the outer estuary, and in the flood-dominant areas in particular, tends to be composed of medium to coarse, or even very coarse, sand, whereas the middle and inner estuary are characterized by fine and very fine sand. If dunes are present on the channel floor, the muds are preferentially preserved in their troughs generating muddy bottomset and toeset deposits, characteristic for FA 3. The sands in these channel deposits will fine upward, whereas the amount of mud and mud-layer thickness will decrease upward, producing an upward-cleaning, but upward-fining succession (Dalrymple, 2010), similar to those in outcrop 1, FA 3. At any one location, the cross bedding is likely to have a unidirectional paleocurrent direction because of the local dominance of the flood or ebb current (Dalrymple et al., 1990) also characteristic for this part of the section.

Thus, in summary, according to the definition of Dalrymple et al. (1992), the succession of SU2 in the study area formed in fluvial and slightly tidally influenced fluvial environment of inner estuary with gradual increase of the sinuosity of channels basinwards, towards central estuarine transitional zone with tidal channel and bars of high sinuosity, and to rather straight channels and bars of outer estuary, reflecting the seaward increasing dominance of tidal processes (Figure 5.20.). The upward-fining from the base of SU 2 to its top and the development of the estuarine succession indicate gradual relative sea-level rise rate through the estuary evolution. The erosively based SU 3 in the southern part of the study area composed of tide-influenced fluvial and tidal channel deposits (FA 2) indicates a new stage of incision and infill (see Figure 5.15.).

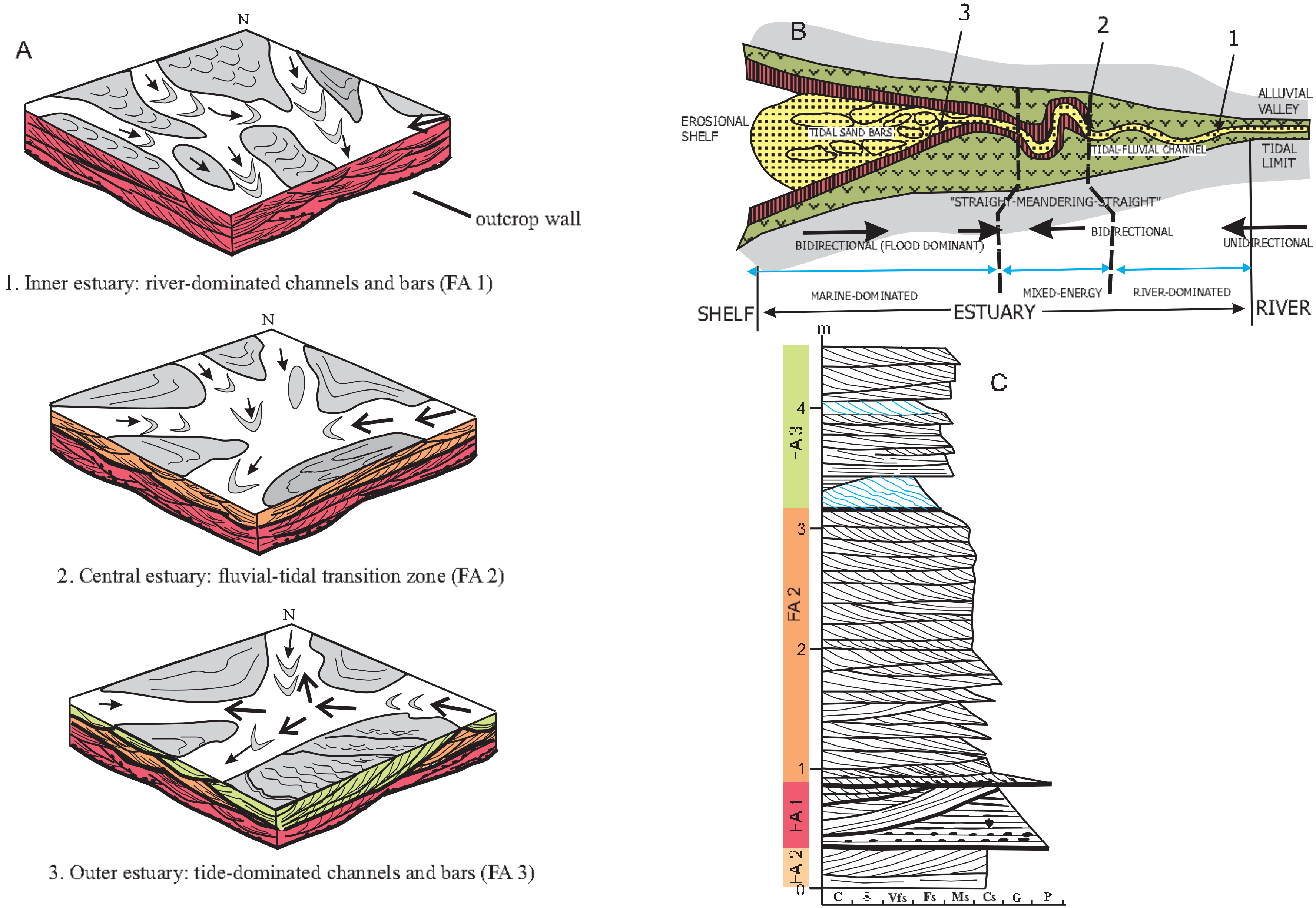

Figure 5.20. See next page
5.20. attēls. Skat. nākamo lappusi

Figure 5.20. Bedform reconstruction from Stratigraphic unit 2, Tori outcrop (Nr 1)
A - A three-dimensional depositional model representing the development of the estuarine bedforms in Stratigraphic unit 2 from outcrop 1, based on cross-strata and major bedform palaeocurrent directions, as well as on facies architecture. The thinner arrows mark directions of fluvial flows, while thicker arrows mark the directions of tidal currents; B - Estuary model, which illustrates the position of various stages of the Tori estuary (after Dalrymple, Choi, 2007); C - A representative measured section from Tori outcrop

5.20. attēls. 2. slāņkopas paleoformu rekonstrukcija Tori (Nr 1.) atsegumā
A - 3-D modelis, kas ilustrē estuāra paleoformu attīstību pēc slāņu virsmu krituma azimutu mērījumiem un slāņkopas uzvbūves īpatnībām. Tievākās bultas norāda uz fluviālām plūsmām, bet biezākās uz plūdmaiņu plūsmām; B - Tori estuāra attīstības stadiju modelis (pēc Dalrymple, Choi, 2007); C - Tori atseguma nogulumu griezums

The overall fining-upward pattern of the succession is interpreted to be a result of vertical channel aggradation and lateral channel migration by point bars (Miall, 1992). Major bedding surfaces with flood oriented strata dips are observed only locally in southern end of the outcrop belt. However most facies dip basinwards, suggesting ebb-dominated currents. These cross-sets presumably formed on similar scale bedforms that migrated across broad shallow-water shoals where the dominant current was locally flood directed and free to cross the shoal morphology. As a result the ebb channels dictated the large morphological changes (see Figure 5.19.).

Overall the Pärnu Fm in the study area is interpreted to represent an estuary filling an incised valley (Emery, Myers, 1996; Boyd et al., 2006; Dalrymple, 2006). Poorly sorted fluvial channel deposits with significant erosional surfaces and channel-fill architecture in the lowermost part of the succession derived from adjacent drill core data, suggests that deposits have been formed due to a valley cut into underlying Silurian carbonate rocks.

The retrogradational, fining upwards stacking pattern of each SU indicates that the rate of sediment input was less than the rate at which depositional (accommodation) space generated by sea-level rise. Such a stacking pattern is particularly common during transgression (Reading, Collinson, 1996), resulting in flooding of this valley (Allen, Posamentier, 1993). A successive change from tidal bedforms into tidal flat, capped by overlying shallow marine deposits of Narva Fm (Tänavsuu-Milkeviciene, Plink-Björklund, 2009), records continued deepening of the basin. The river valley was incised during sea-level falling stage and lowstand, and was transformed into an estuary during rising stage of relative sea level (Reisnos, 1992), that underwent 3 stages of incision and infill (see SU 1, SU 2, SU 3). According to V. Kuršs (Kurshs, 1992) and Kleesment (1997) the beginning of Pärnu time throughout the BDB is marked with a major transgression. The lowstand deposits of SU 1 and the relatively thin lowstand fluvial deposits of SU 2 must have been largely eroded, as the valley acted as a bypass zone. The lowermost boundary of these beds, cut into underlying Silurian carbonate rocks, is interpreted to mark the sequence boundary of the succession that can be traced farther basinwards and marked by a significant uncomformity (see Figure 5.17.). This is possibly why no fluvial facies are preserved at the base of SU 1 and the valley fill begins with tide-influenced fluvial-tidal channel and bar deposits, because fluvial incision has been modified by the initial stages of marine transgression (Reinson, 1992). The overlying tidal bars and tidal flat of the Pärnu succession formed in response to relative sea level rise and belong to a fining upward unit of transgressive system tracts, developed during the major period of a relative sea level rise (Dalrymple et al., 1994). It formed due to the progradation and lateral migration of tidal channels and tidal flat deposits across the estuarine system.

5.1.4. Tidal signatures in coarse-grained sediments and tidal influence to the deposition

Tide-influenced environments produce special types of cross-strata that leave tidal signatures (Nio, Yang, 1991; Dalrymple, 1992). Although a lot has been learned in recent years about the stratigraphy of the deposits of tide-dominated estuaries, much less is known about the detailed nature of the facies within them (Dalrymple et al., 2012).

General criteria for recognizing tidal signals in recent and ancient strata are relatively well known (De Raaf, Boersma 1971; Nio, Yang 1991; Dalrymple, 1992; 2010; Fenies et al., 1999; Dalrymple, Choi, 2007), and have also been well used for the BDB (Pontén, Plink-Björklund, 2007; Tänavsuu-Milkeviciene, Plink-Björklund, 2009).

The aim of this chapter is to acquaint with various tidal signatures documented in the Pärnu Fm in outcrop area, such as tidal bundles and neap-spring cycles, mud drapes, reactivation surfaces together with associated sigmoidal cross-stratification, current reversal indicators, rhythmic current velocity changes and rhythmic sediment fallout rate changes at cross-strata and cross-laminae scale. These tidal signatures are important in reconstructing the ancient environmental settings of deposition and to distinguish in which environment they occur: fluvial-tidal transition zone and proximal reaches of outer estuary zone.

The hydraulic and morphological characteristics makes the transitional zone different from the "pure" fluvial and estuarine environment and make as such it distinguishable on the basis of sedimentary structures and textures as a separate environment. Not much information exists on this zone (Dalrymple, Choi, 2007; Van den Berg et al., 2007). In this dissertation it is aimed to pinpoint common characteristics that allow recognition of the fluvial-tidal transition signatures from signatures in outer estuary zone (Table 5.5).

Many structures of the fluvial-tidal zone and proximal reaches of outer estuary suggest conditions of rapid deposition. This may reflect the high mobility of channels in both environments. The diagnostic features of the fluvial-tidal transition zone and inner outer estuary, as manifested in the studied outcrops, are described and illustrated in detail here. They primarily refer to large-scale dune cross bedding and decimetre-scale successions. However none of the distinctive structures mentioned hereby of the fluvial-tidal zone and inner reaches of outer estuary can be considered as exclusively diagnostic for the recognition of these zones. Therefore, the criteria listed here should be used as a complex of diagnostic features and have a comparative value.

Tidal bundles and neap-spring cyclicity

Tidal bundle is defined by Boersma (1969) as the deposit of a single, dominant tide, which is bounded by mud drape or erosional surfaces produced during the slack-water periods or subordinate current. Tidal bundle consists of the material that is deposited on lee side of the bedform during the dominant current stage. The horizontal extent of each bundle shows the net migration during a single tide. The succession of systematic changes in bundle thickness commonly reflects deposition over a series of neap-spring tides (Nio, Yang, 1991; Dalrymple, Rhodes, 1995). During spring tides, which are characterised by higher tidal ranges and current velocities, thicker tidal bundles are deposited, whereas neap periods result in thinner tidal bundles.

Tidal bundles are one of the most typical tidal signatures in the study area. Tidal bundles are observed in FA 2 and 3 that occur in SU 1, 2 and 3. They are especially prominent in FA 3 of SU 2, which represents outer estuary. There in many places muddy, neap-tide intervals composed of closely-spaced or very often amalgamated mud drapes typically pass laterally within cross-sets into thicker and sandier, spring-tide bundles (Figures 5.21A and B). Apart from the variations in bundle thicknesses, variations in shape and thicknesses of cross-stratified units are also observed for FA 3. The thick bottomsets are related to the thinner bundles, indicating that well developed bottomsets have been formed during neap tides. The variations in current speed, causing erosion during spring tides and infilling during neap tides, produced also changes in dune height, which are reflected in the depth of erosion (Figure 5.21B). Larger dunes were formed during the stronger currents of spring tides and smaller dunes - during the neap tide (Nio, Yang, 1991).

For FA 2, which represents tidal-fluvial transition zone, tidal bundles are highlighted usually by reactivation surfaces, and less by mud drapes, which are rarer and indistinct. Besides the lateral sequence of tidal bundles in FA 2 is not arranged in such a consistent way representing clear pattern of neap-spring cycles of mud-draped tidal bundles as for FA 3. This coincides with the observations of Van den Berg et al. (2007), for the tidal bundles of the tidal-fluvial transition zone deposits.

Attempts to count the tidal bundles based on time-series analysis in the study area were unsuccessful. The number of neap-spring tidal bundles in ideal pattern are 7-14, but it is rarely seen in ancient rock record (Nio, Yang, 1991), including the study area, mainly due to the amalgamation and omission of the bundles, which makes impossible to make calculations (Figure 5.21C). The amalgamation of the bundles is caused by the interruption in sedimentation or ceased bedform migration during neap-tides, which are characterized by very low current velocities below the threshold of sediment movement, thus current velocity is too low for sand movement (Tessier, Gigot, 1989; Stupples, 2002). Amalgamation can be caused also by spring tides, when several sand layers are deposited without intervening muds, which are removed by the stronger spring tidal currents (Stupples, 2002). Similarly, at neap tides low energy conditions prevail and deposition of sand might not occur for several tides, allowing the amalgamation of mud layers (Stupples, 2002). Omissions of bundles can be caused by possible reworking by storm event or wave processes (Yang, Nio, 1985; Kvale et al., 1995), which is unlikely for the study area. Most probably, it is caused by stronger spring tidal current.

Summarizing, the inconsistent way of arrangement of tidal bundles in deposits of FA 2 reflects deposition under tidal influence in fluvial-tidal transition zone, with strong, but irregular tidal flow presence. In the channels and bars of the proximal part of outer estuary (FA 3) tidal bundles are more abundant and better organized. They are distinct, with clear pattern of neap-spring cyclicity, thus reflecting stronger and constant tidal influence to the deposition.

Mud and mica drapes

Mud drapes form on the lee side of the bedforms during slack water periods, a time of water-stand-still (Nio, Yang, 1991). Such slack-water mud drapes may form in different depositional environemts, in particular in rivers, which have only seasonal flow, but they are most common in tidal settings, where abundant, regular mud drapes together with tidal bundles are good indicators of tidally dominated environments (Nio, Yang, 1991).

Mud and mica drapes are very common tidal signature in the study area and are present in FA1, FA 2 (SU 1, 2, 3), and FA 3. In fluvial deposits of FA 1 few mud drapes occur only in southern part of the study area. For FA 2 in all stratigraphic units mud drapes are thin and relatively indistinct, they appear to be also more silty and mica rich.

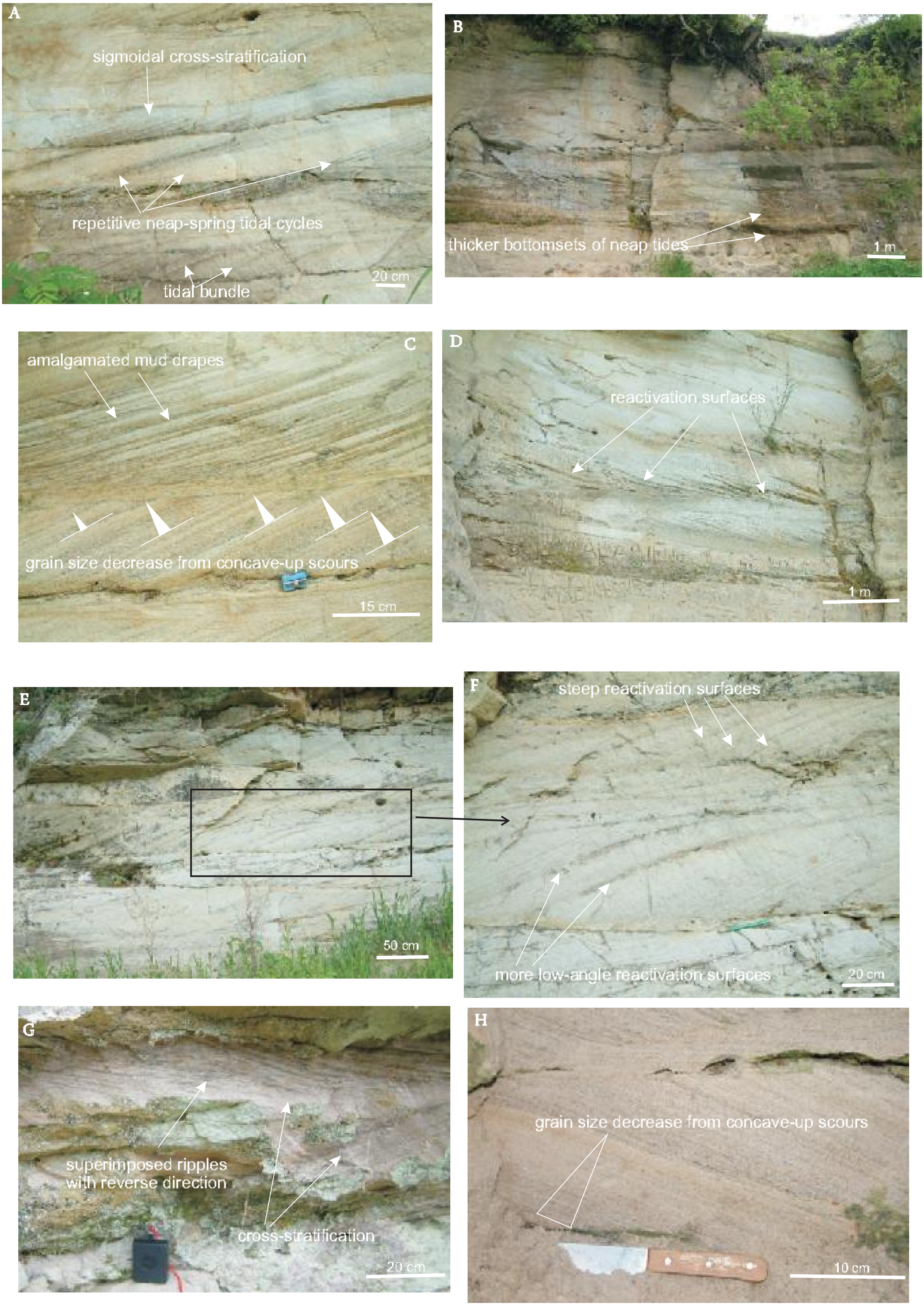

Figure 5.21. See next page
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Figure 5.21. Illustrations, showing variety of tidal signatures in the deposits of Pärnu Formation in the outcrop area

A – Tidal bundles, neap-spring tidal cycles and mud/mica drapes in deposits of Facies Association 2, Stratigraphical Unit 2, outcrop 4; B – Fine-grained cross-stratified sandstone with thick mud drapes in the middle part of the outcrop wall, representing Facies Association 3, Stratigraphical Unit 2 from outcrop 1. Note erosional scours at the bottom of the package with thick mud layers; C – Tidal bundles in cross-stratified sandstone, representing cyclic grain size decrease from concave-up scours in the lower part of the photo, amalgamated mud and mica drapes in the upper part of the photo, Facies Association 2, Stratigraphical Unit 2, outcrop 4; D – prominent and low angle reactivation surfaces and thick mud drapes in the deposits of Facies Association 3, Stratigraphical Unit 2, outcrop 1; E – in comparison to D, relatively steep reactivation surfaces in coarser-grained sediments of Facies Association 2, Stratigraphical Unit 2, outcrop 4; F – enlarged photo of E; G – fine-grained cross-stratified sandstone with superimposed reversely-directed current-ripple lamination, stressed by mud drapes; H – an illustration of grain size gradual change within cross-strata in variably-grained sandstones of Facies Association 2, Stratigraphical Unit 1, outcrop 7

5.21. attēls. Plūdmaiņu pazīmes Pērnavas svītas nogulumos atsegumu teritorijā

A – Plūdmaiņu kārtas, plūdmaiņu cikli un māla/vizlu kāriņas 2. fāciju asociācijas nogulumos, 2. slāņkopa, 4. atsegums; B – Smalkgraudainas, slīpslāņots smilšakmens ar biezām māla kārtām atseguma vidusdaļā, 3. fāciju asociācija, 2. slāņkopa, 1. atsegums. Slāņkopas lejasdaļā erozīvs iegrauzums pildīts ar biezākām māla kārtām; C – Plūdmaiņu kārtas slīpslāņotajā smilšakmenī, kas redzamas cikliskās graudu izmēru izmaiņās slīpslāņotās kārtās attēlā apakšdaļā; saplūdušas kopā māla/vizlu kārtas attēla augšdaļā, fāciju asociācija, 2. slāņkopa, 4. atsegums; D – izteiktas un lēzeni krītošas reaktivācijas virsmas, kā arī biezas māla kārtas 3. fāciju asociācijasnogulumos, 2. slāņkopa, 1. atsegums; E – salīdzinājumā ar D attēlu, relatīvi stāvi krītošas reaktivācijas virsmas rupjgraudainākos 2. fāciju asociācijas nogulumos, 2. slāņkopa, 4. atsegums; F – palieināts fragments no E attēla; G – smalkgraudains slīpslāņots smilšakmens ar pretējā virzienā vērstu straumju ripsnojumu, ko pasvītro māla kārtiņas; H – graudu izmēru ritmiskas izmaiņas slīpslāņotās kārtās dažādgraudainajā 2. fāciju asociācijas smilšakmenī, 1. slāņkopa, 7. atsegums

On the contrary, regular and distinct mud drapes are most abundant in bedforms of FA 3, indicating increased suspended sediment load and better-developed slack-water periods (see Figure 5.21B).

In deposits of FA 2, in contrast to FA 3, mica drapes are more characteristic. According to Van den Berg et al. (2007), mud drapes in the fluvial-tidal zone seem to be of a more silty nature as compared with estuarine slack water deposits. The mica drapes are similar to mud drapes and organic material drapes already described (Visser, 1980; Nio, Yang, 1991; Fenies et al., 1999). Nevertheless, mica and mud drapes are less abundant than in FA 3, more indistinct and amalgamated, thus reflecting that the waning currents were largely absent. Also, the low-tide slack-water drape is restricted to the bottomset of cross-sets and drapes the emergence run-off current ripples, generated in very shallow water by run-off currents that flowed along the dune troughs just before or during the emergence of the bar at low tide (Fenies, 1999). The absence of regular mica and mud drapes may be because currents were generally more continuous during deposition of this facies or that stronger currents removed evidence of depositional pauses (Fenies, 1999; Willis et al., 1999).

For the deposits of FA3 mud drapes are typical, which are often more than 1 mm thick, reflecting high-suspended mud concentrations generated by regular water-stand-still. The differences in thicknesses of mud drapes reflect their deposition in neap and spring tidal cycles. Thicker and amalgamated mud drapes, deposited during neap-tide, are well preserved and occur in the bottomsets of cross-strata (see Figures 5.21B and D), forming thick packages. This indicates that the thick bottomset intervals correspond to neap-tide conditions, when dune height decreases and the dune trough is partly filled in (Van den Berg, 1982; Rahmani, 1988). During the spring tide, when current activity increased, the slack-water mud drapes were eroded at the dune crest by the tidal current and were preserved mainly bottomsets (Dalrymple, Rhodes, 1995). According to Van den Berg et al.

(2007), such thick bottomsets in tidal deposits can be distinguished from their counterparts in fluvial-tidal transitional zone by a much better organization. Mud drapes are also common in cross-laminated sandstones draping regularly each cross-laminae. At the larger bedform scale, in FA 3, where slightly inclined master bedding surface are present with superimposed cross-strata, mud drapes are concentrated in the lower parts of these bedforms and decrease upward thus reflecting a cleaning-upward pattern (see Figure 5.14.). This probably reflects that mud was washed away from the upper parts of the bars in stronger flood currents. In general, single mud drapes are typical for the entire succession, whereas double mud drapes are nearly absent, which together with occurrence of abundant reactivation surfaces indicate possible erosion of first mud drape by relatively strong subordinate currents (Dalrymple, Rhodes, 1995; Willis, 1999).

The presence of the mud couplets and tidal bundles is diagnostic feature of subtidal sedimentation (Rahmani, 1988). Few mud drapes in fluvial deposits of FA1 at the very basinward part of the outcrop belt indicate occasional tidal influence. Indistinct and irregular mud drapes of tidal-fluvial transition zone (FA 2) reflect deposition most probably due to relatively low suspended sediment concentration and too high water mobility (Dalrymple, Choi, 2007) in fluvial-tidal transition zone. This is in contrast to Van den Berg et al. (2007), who describes thick bottomsets of mud in cross-strata in deposits formed in fluvial-tidal transition zone. According to Fenier et al. (1999), who provides indications from intertidal deposits of Gironde estuary, in intertidal sediments during neap tides, mud drape couplets are not visible because the bundle of the subordinate current is too thin. Mica drapes typical for FA 2, imply stronger fluvial currents, or weaker subordinate tidal current than mud drapes (Pontén, Plink-Björklund, 2007), therefore they are common for tidal-fluvial transition zone of the estuary. Most abundant, regular and thick mud drapes of FA 3 suggest deposition in tidal currents in proximal parts of outer estuary.

Reactivation surfaces

Reactivation surfaces are minor erosional surfaces within the cross-stratified beds and are widely documented in the deposits of the Pärnu Fm in the study area. Reactivation surfaces can form in different depositional environements (Nichols, 1999). In fluvial depositional environment they indicate interruption in migration of the bedforms during low river stage, exposing the bedform and causing the erosion, with the following rise in fluvial stage allowing the bedform to resume migration and build over (Collinson, 1996). However the wide presence of reactivation surfaces together with mud and/or mica drapes, tidal bundles and current reversals, indicate formation due to tidal currents.

Reactivation surfaces, interpreted to be formed due to the interruption of bedform migration, are very typical tidal signatures for the deposits of the study area. They occur in FA 1, 2, 3 throughout the entire section, except for the topmost part formed by tidal flat deposits (FA 4).

In the fluvial deposits of FA 1, where reactivation surfaces are irregular, it is suggested that the bedform migration was occasionally interrupted due to episodes of either stonger fluvial currents or by tidal currents (Collinson, 1996; de Mowbray, Visser, 1984). In the dominant part of the succession (FA 2 of SU 1, 2, 3 and FA 3 of SU 2) the abundant occurence of reactivation surfaces together with the evidence of mud drapes, tidal bundling and sigmoidal cross-stratification (FA 3; Figure 5.21D) is interpreted to be formed due to the erosion caused by subordinate tidal currents (de Mowbray, Visser, 1984; Nio, Yang, 1991; Shanely et al., 1992).

The character and shape of reactivation surfaces in the study area varies in different parts of the section. The bedforms, eroded by subordinate currents, illustrate different angles in removed crests of the subaqueous dunes. For cross-stratified beds of FA 2 relatively steep inclination in dip of reactivation surfaces records short term changes in tidal bedload transport (Dalrymple, Rhodes, 1995) and weaker subordinate current (see Figures 5.21E and F). Whereas for cross-strata of FA 3 the reactivation surfaces are low-angled (see Figures 5.21B and D), illustrating increase in strength of the subordinate current. Thus, erosional reactivation surfaces for the deposits of FA 3 become prominent, low-angle surfaces at spring tides. Moreover, the slope of the reactivation surfaces may change laterally because of the changes in tidal velocities in spring and neap tides (Fenies et al., 1999). Variations in the spacing of reactivation surfaces and the thickness of the bundles reflect changes in bedload transport over neap-spring tidal cycles (Richards, 1994), and the reactivation surfaces in deposits of FA 3 are closely spaced, reflecting stronger tidal currents (Nio, Yang, 1991; Dalrymple, Rhodes 1995).

Reactivation surfaces are quite prominent in many ancient stratigraphic sequences and are represented in subtidal and tide-influenced environments (Nio, Yang, 1991). In the study area occasional reactivation surfaces occur in the fluvial channel bedforms (FA 1) of the Pärnu time estuary. Increased number of reactivation surfaces is most common in the central (FA 2) and outer parts (FA 3) of the Pärnu time estuary, indicating gradual increase in subordinate currents and stronger tidal influence in comparison to inner estuary.

Tidal current reversals

Typical palaeocurrent reversals such as herringbone cross-stratification is not an abundant feature in the studied succession. According to Nio and Yang (1991), tidally dominated system is ususally characterised by an asymetrical tidal pattern. Whereas, bi-directional or herringbone cross-stratification is usually formed within a symmetrical tidal pattern, which is very uncommon. One of the tidal currents is usually stronger than the other, moreover, the ebb and flood currents may follow different paths. The bidirectional cross-stratification requires thus ebb and flood currents to occur at different times in a place where the rate of sedimentation is high enough to preserve the cross-stratification (Nichols, 1999). According to Van den Berg et al. (2007), herringbone cross-stratification is one of the diagnostic features of tidal-fluvial transition zone.

Current reversals at different scales are typical for the deposits of the study area, such as 1) bipolar direction in large scale bedform migration in the section, 2) bipolar/bimodal directions of bedforms (mainly at dune and bar level) from adjacent sites/ outcrops and 3) opposite directions of ripple-cross lamination and the cross-strata.

One of the indicator of tidal current reversals is bipolar direction of barform migration. For example, in the largest outcrop of the study area the overall palaeocurrent distribution has a bidirectional pattern. Here 207 measurements were derived from master bedding surfaces, cross-beds and current ripple lamination. Such an overall bidirectional palaeocurrent pattern was obtained due to the large number of measurements taken, as well as due to the 3-dimensional character of the outcrop that unlike other outcrops of the study area, reflects almost equal distribution of FA 1, FA 2 and FA 3 in the section.

For the deposits of FA 2, especially in SU2, more ebb-dominated bedforms are typical, and herringbone cross-stratification is absent. Unlike in the outer estuary, the flood and ebb flow in the tidal-fluvial transitional zone is forced into one and the same channel, in which the flood current is relatively weak (Van den Berg et al., 2007; see Figure 5.19.). This results in the production of bedsets displaying the dominant ebb-direction. On the contrary, for FA 3 bipolar palaeocurrent orientation of bars in adjacent sites is common, which is interpreted to be deposited in mutually evasive channels of outer estuary. Here the pattern results in deposits entirely composed of either ebb or flood dominated major bedforms (see Figure 5.19.). This is in accordance with Van den Berg et al. (2007), with the exeption that herringbone cross-stratification is absent in the deposits of Pärnu succession.

One of the most significant evidences of current reveals in the study area are in general oppositely, i.e. landward directed current ripple-lamination, typical for FA 3, and rarely for FA 2. They occur as small scale bedforms, climbing up along the entire lee faces of dunes, thus excluding the formation by the flow-separation vortex, where the ripples are restricted to the lower parts of the cross-strata. Therefore such ripple-lamination is interpreted to be generated by relatively weak flood currents (Figure 5.21G). Such oppositely-directed ripple-lamination is most common for mud rich facies of the FA 3. In contrast, in the deposits of FA 2, ripple-lamination is also typical, however, the ripples superimposed on cross-strata appear to be oriented in various directions, and occur mainly at the bottomsets of cross-beds. This is probably caused by the deterioration of the flow vortex in the dune's trough by a temporal weakening or reversal of the flow during rising tide (Van den Berg et al., 2007).

Grain size decrease on cross-strata scale

One of the evidences for the tidal current activity in the deposits of Pärnu succession is the regular decrease of the grain size from the concave-up scours within individual cross-strata in the cross sets. These features are more distinct and abundant in FA 2 of SU 2, but are also common in FA 2 of SU 1. The grain-size changes from coarse to gravelly sand gradually into medium- and fine-grained sand passing to very fine-grained sand and in places to silt. The regular variations of grain-size within cross-strata reflect the regular variations in tidal current speed. Such a regular alternation of grain size is interpreted to be caused by tidal velocity variations, suggesting a regular and orderly fluctuation in energy conditions common to tidal processes (Dalrymple, Choi, 2007). Dalrymple and Choi (2007) suggest that streamwise alternations in fining and coarsening of cross-strata reflect tide-induced variations in flow strength. Coarse to medium sand was deposited during the higher flow velocities, while the finer sand and silt during lower flow velocities (Middleton, 1991; Nio, Yang, 1991; Pontén, Plink-Björklund, 2007). These features, observed in FA 2 in association with reactivation surfaces (see Figures 5.21C and H) form similar to tidal bundles structures and are absent in deposits of FA 3. The difference in such tidal bundling with FA 3, is that in FA 3 mud drapes are usually present and underline each bundle sequence, while in FA 2 they are less common and each bundle is identified by gradual and systematic decrease of the grain size at individual cross-strata.

The abundance of the regular upward-decrease of the grain size from the concave-up scours within individual cross-strata in the cross sets is prominent in SU 2, and it reflects deposition in somewhat more tide-dominated environment in comparsion to the same FA 2 of the SU 1. This in addition to the interpretation based on architectural elements and palaeocurrent directions of FA 2 of both these stratigraphic units, supports suggestion that the FA 2 of SU 2 represents more distal reaches of the tidal-fluvial transition zone of the estuary in comparsion to SU 1. The question arises, what type of sedimentary signatures indicating tidal influence might be expected in deposits generated in the upstream part of the fluvial-tidal zone where the tide is reduced to temporal discontinuities because of the strength of the river flow? According to Davis (2012) not all tidally-influenced environments produce tidalites and one of the most common environments where this situation prevails is the fluvial-tidal transition zone in estuaries.

From the few tidal signatures present in fluvial depoists of FA 1, such as reactivation surfaces and few and indistinct mud drapes, followed by irregular foresetting (Van den Berg et al., 2007) in deposits of FA 2, are the first signs in fluvial deposits that announce tides causing modulation of fluvial currents in the most upstream part of the fluvial-tidal zone.

Rhythmical variations of the shape of climbing ripple lamination

Climbing ripple-lamination is formed in a high suspended sediment load, where deposition from suspension exceeds the rate of traction deposition. The development of climbing-ripple lamination from ripples requires that abundant sediment is continuously available in a current so that the ripples are built upward in an overlapping series rather than merely migrating in a downcurrent direction (Reineck, Singh, 1980).

Climbing ripple-laminated facies are typical for deposits of FA 3 in the middle and upper part of the succession. The ripples occur as superimposed bedforms on cross-stratified units, representing larger bedforms, such as dunes and bars. In outcrop 7 such facies are observed in slighly inclined tidal bar deposits (Figure 5.22A) of FA 3. They form a series of climbing ripples on the main bedding surfaces and are about 1.3 m thick and 5 m long. The height of the climbing ripples is from 0.5 to 1.5 cm. Depending on the character of the shape of laminae, in the outcrop 7 of the study area the vertical progressive transition of ripples from in-drift type to in-phase type occurs according to Jopling and Walker (1968) (Figure 5.22B and C).

For ripple lamination in-phase, each ripple crest is situated directly above the other, in places with slight shift of crests in one direction. On the contrary, ripple laminae in-drift have nearly paralel bounding planes which dip in the upstream direction. The planes represent the surface of nondeposition or even slight erosion of the stoss side of ripples (see Figure 5.22C).

Such a progressive transition from ripple laminae in-drift to ripple laminae in-phase indicates a slight, but regular, progressive decrease in current velocity with an increase in depth (a progressively waning stream). Increasing angles of climb indicate that the ratio of the vertical aggradation rate to the downstream ripple migration rate increased (Ashley et al., 1982), thus increases the ratio of deposition between suspended material and traction bedload (Jopling, Walker, 1968). It must be noted, that climbing ripple lamination is typically present in various depositional environments such as fluvial and those with tidal influence (Reineck, Singh 1980; Davis, 2012). According to Davis (2012), unless a cyclic reversal of the directional orientation of the cross-strata occurs, there is no reason to interpret climbing ripple-lamination as tidalites. In this study it is argued, that regular variations in the pattern of shape of climbing ripple-lamination together with occurrence of mud and mica drapes on laminae, indicating rhythmically waning current flows and slack water periods, reflect the strong influence of tidal processes to the deposition and is considered as a tidal signature.

Similar structures of climbing ripple-lamination from tide-influenced deposits have been described by Lanier et al. (1993) in ancient sediments, Lanier and Tessier (1998), both in ancient and modern tidal environments, as well as by Yokokawa et al. (1995) in tidalites of the middle Pleistocene age in Japan, and in recent and Tertiary deposits of the Rhine and Meuse Rivers in the Netherlands described by Van den Berg et al. (2007). For the study area, these signatures are typical in fine-grained deposits of FA 3, interpreted to be deposited in proximal reaches of the outer estuarine zone.

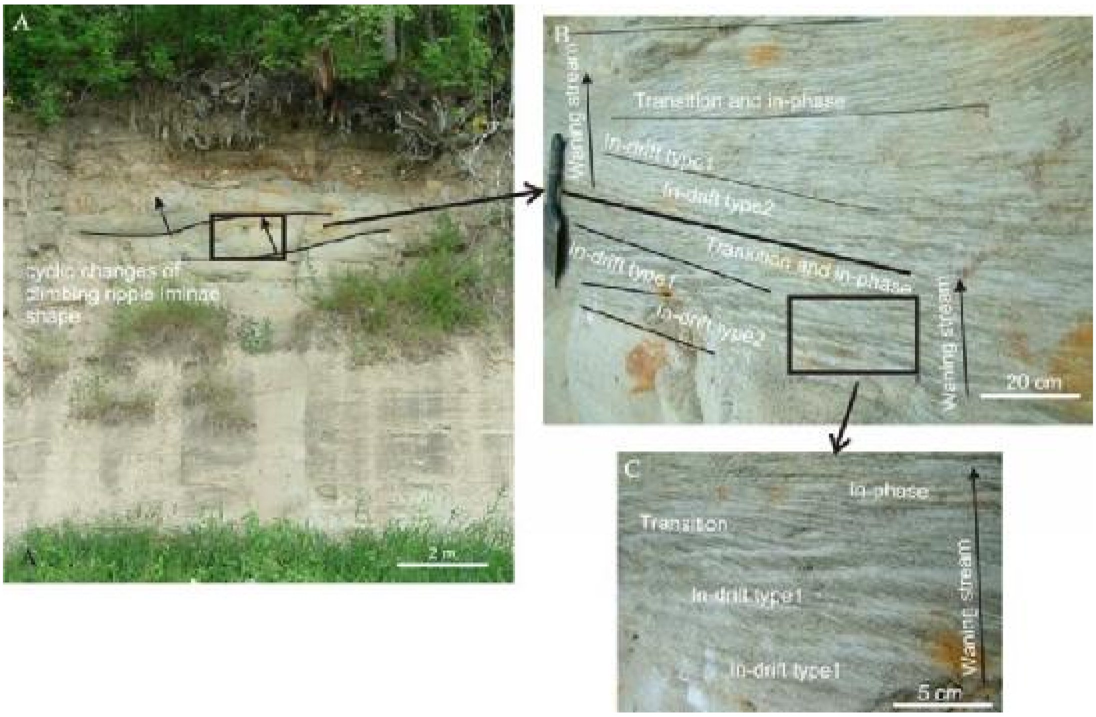

Figure 5.22. Illustrations, showing vertical progressive transition of the shape of climbing ripple lamination, observed within tidal bar deposits of Facies Association 3

Such a transition from ripple laminae in-drift to ripple laminae in-phase indicates rythmical, progressive decrease in current velocity, interpreted to be formed due to tidal current activity (outcrop 7)

5.22. attēls. Progresīvas izmaiņas kāpjošā ripsnojuma morfoloģijā vertikālā virzienā, kas novērojamas 3. fāciju asociācijas plūdmaiņu sēru nogulumos

Šādas izmaiņas ripsnojuma formā griezumā norāda uz ritmisku un progresīvu straumes ātruma samazināšanos, ko nosaka plūdmaiņu straumju darbība (7. atsegums)

Table 5.5. Tidal signatures and their comparison in fluvial-tidal transition zone and outer zone of coarse-grained estuaries
5.5. tabula. Plūdmaiņu pazīmes un to salīdzinājums fluviālu-plūdmaiņu apstākļu pārejas zonā un rupjgraudaino nogulumu estuāros

Tidal signatures	Fluvial-tidal transition zone (FA 2)	Outer estuary (FA 3)
Tidal bundles	Description: • Inconsistent way in arrangement of tidal bundles. Interpretation: • Strong, but irregular tidal flow presence.	Description: • Tidal bundles are more abundant and better organized, more distinct with clear pattern of neap-spring cyclicit. Interpretation: • Stronger and constant tidal influence to the deposition.
Mud and mica drapes	Description: • Mud drapes are thin, more indistinct and amalgamated, they appear to be also more silty and mica rich. Interpretation: • Mica suggests stronger fluvial currents or weaker subordinate tidal current. • Indistinct and irregular mud drapes reflect deposition most probably due to relatively low suspended sediment concentration and too high water mobility.	Description: • Reular and distinct mud drapes are most abundant, mud drapes dominate over mica drapes and are often more than 1 mm thick. Interpretation: • Regular and distinct mud drapes indicate increased suspended sediment load and better-developed slack-water periods, reflecting high-suspended mud concentrations generated by regular water-stand-still.
Reactivation surfaces	Description: • For more inner reaches of estuaries reactivation surfaces occur ocasionally and irregularly. • For transition zone reactivation surfaces are prominent and abundant, with steep dipping of surfaces. Interpretation: • Occasional and irregular reactivation surfaces in inner reaches of estuaries suggest that the bedform migration was ocasionally interrupted due to epizodes of stronger tidal currents. • Relatively steep inclination of reactivation surfaces of FA 2 records short term changes in tidal bedload transport and weaker subordinate current.	Description: • Reactivation surfaces in deposits of FA 3 are closely spaced. • Reactivation surfaces are prominent and low-angled. Interpretation: • Variations in the spacing of reactivation surfaces and the thickness of the bundles reflect changes in bedload transport over neap-spring tidal cycles, and closely spaced reactivation surfaces reflect stronger tidal currents. • Prominent reactivation surfaces with low-angle inclination illustrate increase in strength of the subordinate current. More low angled surfaces are thus developed at spring tides.
1) Bipolar/bimodal directions of bedforms (mainly	1) Description: • More ebb-dominated bedforms are typical, and herringbone cross-stratification is absent.	1) Description: • FA 3 illustrates bipolar palaeocurrent orientation of bars in adjacent sites.

Table continues. Tabulas turpinājums

dunes and bars) in adjacent outcrops	Interpretation: • Unlike in the outer estuary, the flood and ebb flow in the tidal-fluvial transitional zone is forced into one and the same channel, in which the flood current is relatively weak.	Interpretation: • Bipolar palaeocurrent orientation of bars in adjacent sites is interpreted to represent deposition in mutually evasive channels, which is typical in outer estuaries. Here the pattern results in deposits entirely composed of either ebb or flood dominated major bedforms.
2) Ripple-cross lamination opposite directions to the cross-strata directions.	2) Description: • Ripple-lamination is typical, however, the ripples superimposed on cross-strata appear to be oriented in various directions, and occur mainly at the bottomsets of cross-beds. Interpretation: • Ripple orientation in various directions is probably caused by the deterioration of the flow vortex in the dune's trough by a temporal weakening or reversal of the flow during rising tide.	2) Description: • One of the most significant evidences of current reversals in the study area is oppositely to cross-strata directed superimposed current ripple-lamination. Besides the lamination occurs throughout the cross-beds, and is not limted to the bottomsets only. Interpretation: • Such ripple-lamination is interpreted to be generated by relatively weak flood currents.
Grain size decrease on cross-strata scale	Description: • Grain size decrease on cross-strata scale is more distinct and abundant. The grain size changes upwards from coarse-grained to gravelly sand gradually into medium and fine-grained sands passing to very fine-grained sand and sometimes to silt, forming bundles. Interpretation: • Such a regular alternation of grain size is interpreted to be caused by tidal velocity variations, suggesting a regular and orderly fluctuation in energy conditions common to tidal processes.	Description: • Grain size decrease on cross-strata scale is in general absent. The difference with FA 2 is that tidal bundling mud drapes are usually present and underline each bundle sequence, unlike in FA 2, where mud drapes are nearly absent. Interpretation: • Absence of grain size decrease on cross-strata scale, and presence of tidal bundles with a distinct mud drapes, suggest stronger and constant tidal influence to the deposition with better-developed slack-water periods.
Morphological changes in climbing ripple-lamination	Not present in FA 2 (tidal-fluvial) transition zone in the study area.	Description: • Climbing ripple-laminated facies are typical for deposits of FA 3 in the middle and upper part of the succession with gradual changes in ripple morpholgy in the cross-section (from ripple laminae in-drift to ripple laminae in-phase). Interpretation: • Such a progressive transition in ripple laminae morphology indicates a slight, but regular, progressive decrease in current velocity with an increase in depth (a progressively waning stream) caused by tides.

5.2. Depositional environment in the tidally-controlled transgressive succession: Baltic Devonian Basin during the Rēzekne and Pärnu times

This chapter focuses on facies associations (FA) of the Rēzekne and Pärnu RS in all the study area, which is the part of the Baltic Devonian Basin corresponding to the present area of the Baltic States (see Figure 1). The facies association descriptions and interpretations include also the facies and their associations studied in detail in the outcrop belt (see Chapter 5.1.).

5.2.1. Facies associations and sedimentary environments

Nine FAs have been identified for the whole study area, from drill-cores and outcrops, and have been defined by the assemblage of characteristic facies (see Table 5.1 for details). Sedimentary facies have been classified into three groups by carbonate content: dolomite (with more than 66 % of carbonate); dolomitic marl (with 33 % to 66 % of carbonate); siliciclastic rocks (claystone, calcareous or dolomitic clay, siltstone, sandstone with <33 % of calcite or dolomite). Defined FAs are organized into two genetically related groups: 1) carbonate-rich FAs, where carbonate deposits dominate; and 2) siliciclastic-rich FAs, which consist mainly of siliciclastic material. Carbonate-rich FAs are further subdivided into: 1.1) intertidal to supratidal carbonate mudflats, 1.2) intertidal to supratidal carbonate shoals, 1.3) palaeosols. Siliciclastic-rich FAs are subdivided into the following FAs: 2.1) fluvial and tidally-influenced fluvial deposits of inner estuary, 2.2) central estuary deposits, 2.3) tidal channels and bars, 2.4) tidal ridges of outer estuary, 2.5) intertidal to subtidal channels and flats, 2.6) intertidal to supratidal mudflat. The description of FAs is based on the core and outcrop data; core logs from the literature are used additionally in order to clarify mainly the transitional areas and boundaries between FAs.

Carbonate-rich Facies associations

<u>**FA 1.1: Intertidal to supratidal carbonate mudflats**</u>

Facies Association 1.1 consists of dolomitic marl (F20), heterogenous dolomitic marl with silt and very fine sand lamination (F21), dolomitic marl with sandstone grains (F22) and structureless plane-parallel laminated, and wavy laminated dolostone (F23). FA 1.1 occurs mainly in the eastern part of the study area and is up to 20 m thick, reaching 40 m in Liozno drill-core. Thinner units of FA 1.1, up to 2 m thick, occur also in the northern, central and western parts of the study area (Figure 5.23.).

Dolomitic marl (F20) is one of the main facies in this association. The thickness of beds is changing from some tenth of cm to 4 m, maximum 20 m. Sometimes dolomitic marls are distributed as thinner interbeds in other facies. It is dominated in the lower part of the Rēzekne Fm in drill-cores Šķaune-103, Višķi-25 and Ludza-15. It occurs in Alūksne-99 drill-core, where it forms about 5 m thick unit at the top of the Rēzekne Fm and about 4 m thick unit in the middle part of Pärnu Fm. It occurs at the top of Rēzekne Fm in Butkjaliai-250 drill-core forming about 5 m thick beds. In Drissa-1ST drill-core the thickness of dolomitic marls becomes considerable, and this facies forms the bulk of Pärnu Fm, which is about 15-20 m thick. In Mehikoorma-421 drill-core it occurs as bioturbated beds at the top of Rēzekne Fm and is intercalated with F21 and F24. In Tsiistre-327 it appears at the top of Rēzekne Fm and is about 3 m thick. Below it is associated with wave-laminated dolostone (F23) and nodular facies (F24) of the same thickness. About 13 m thick beds of heterogenous dolomitic marl with silt and very fine sand lamination (F21), occur in Rēzekne Fm in Värska-6 drill-core, forming the bulk of the succession.

Dolomitic marls have greenish-grey colour and usually homogenous structure. In some places tepees structures occur as small folds with amplitude not more than 1 cm, that sometimes are steep or even overturned (Stinkulis, 1998). Tepees as defined by Adams and Frenzel (1950) are structures having an inverted V-shaped profile similar to the American Indians tents. However, this form of tepee is only occasionally present and it normally appears as irregular and low ridges (Pratt, 2002). Desiccation cracks are also typical features for F20 and occur throughout the section in FA 1.1. Desiccation cracks may show a variety of sizes depending on exposure time, layer thickness, and the presence or absence of microbial mats (Shinn, 1986). According to Stinkulis (1998) two types of desiccation cracks can be distinguished in deposits of the eastern part of the study area: 1) Very well marked cracks filled with silty sand and coarser sand grains. They are typically up to 1,5 cm wide and at least 5 cm deep. The cracks are V-shaped and typically display a very slight bending near the base. The abundance of these cracks is associated with the presence of the clastic material. This type of cracks is the most typical for the F21 and is distributed in the major part of the dolomitic marl sequence. 2) Cracks filled with dark greenish-grey clay – similar by its composition to the dolomitic marl, but there is more silty sand grains, as well as mica, and less carbonates. The width of these cracks is less than few mm and the depth is up to few mm. In Šķaune-103 drill-core in the depth of 399-400 m the cracks in dolomitic marl are filled with sandy and silty material. It is covered by 2 m thick sandstone bed, which consists of identical material as these cracks are filled with. Moreover, this sandstone contains dolomitic marl pebbles from the lower bed. Gypsum lenses sometimes occur in F20 (Stinkulis, 1998), the thickness of which is from mm up to 3 cm, orange and brown colour.

Dolomitic marl with irregularly distributed coarse and medium sized quartz and feldspar grains (F22) are typical mainly for deposits of the eastern part of the study area. It is associated with homogenous dolomitic marl and forms 11 m of Šķaune-103 drill-core. It also comprises almost 10 m of Rēzekne Fm in the Ludza-15 drill-core. In Aknīste-5 drill-core it reaches 7 m in thickness and forms the top of Rēzekne Fm. In Liozno drill-core it rhythmically alternates with F20 and F19 and forms the bulk of Pärnu Fm, reaching its maximum - 40 m. The quartz and feldspar grains are very well rounded (Stinkulis, 1998) and concentrate in basal parts of dolomitic marls. They are also often associated with desiccation cracks, where the amount of siliciclastic grains increases up to 10-30% (Stinkulis, 1998). Rarely carbonate ooids and peloids occur mixed with siliciclastic material. Irregularly wave-ripple laminated dolomitic marls are present as lenses in those intervals, where sandy material is increasing. Silty sand consists of quartz, feldspar as well as micritic dolomite peloids (Stinkulis, 1998). Wave-laminated dolomitic marl with admixture of clastic material contains wave ripples, where the height of waves is 2.4-2.5 mm and the length is 16.5-18.3 mm.

Dolomitic marl with silty and sandy interlaminae (F21) is also common for this FA, however its distribution is irregular. It occurs in Viļāni-11 drill-core as 2.5 m thick beds, as well as forms about 10 m thick units at the top of Rēzekne Fm. However, mostly it is present as small lenses, with wavy laminae stressed by silty material. In places the amount of silty and sandy material gradually decreases upwards, and few such rhythmic variations are observed.

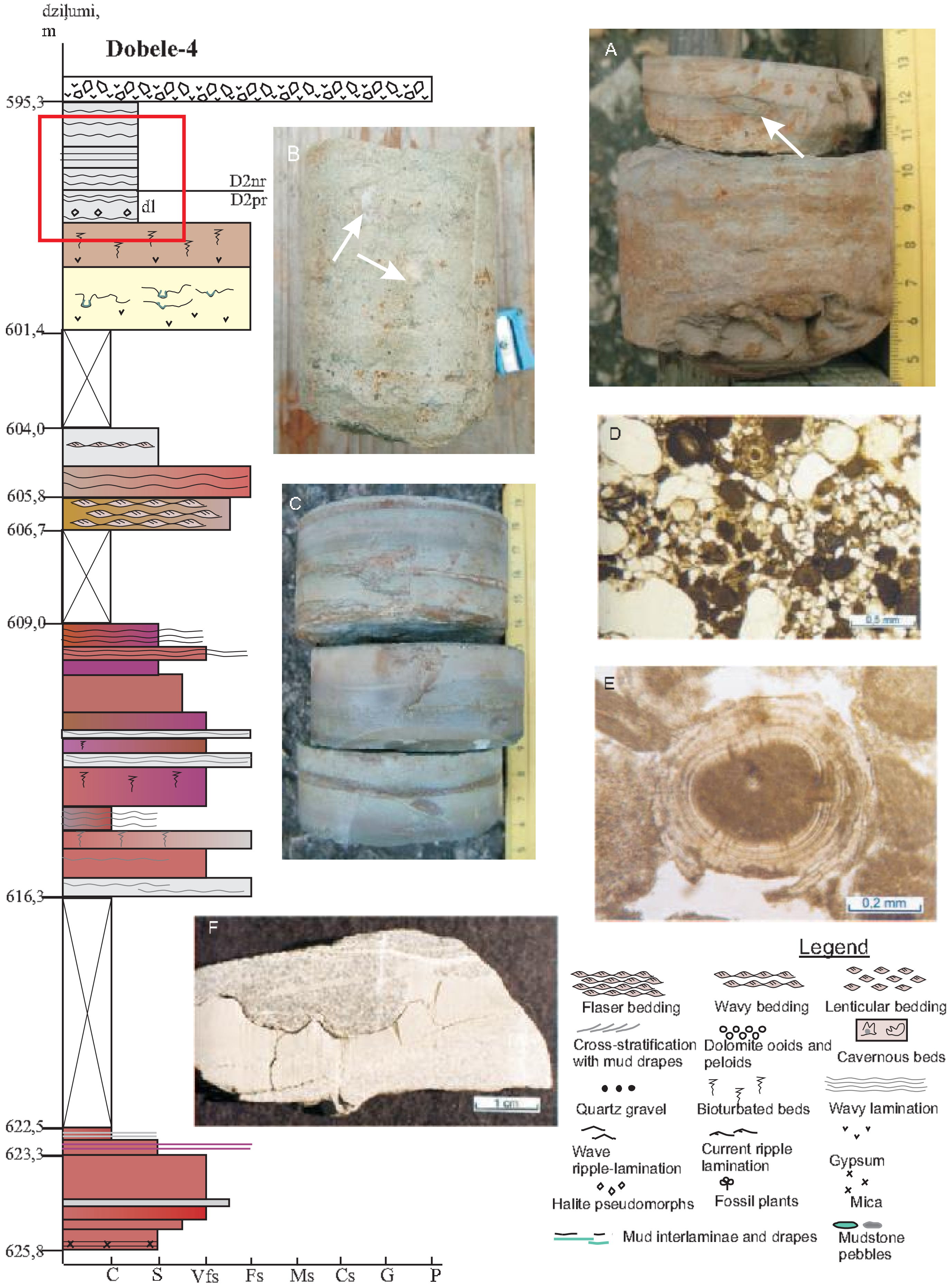

Figure 5.23. See next page

5.23. attēls. Skat. nākamo lappusi

Figure 5.23. Representative photographs and a measured section of the carbonate-rich facies associations: Facies Association 1.1 – intertidal to supratidal carbonate mudflats and Facies Association 1.2. – intertidal to supratidal carbonate shoals

A - Sandy dolomite with birdseye structure (marked with an arrow) - a diagnostic feature to tidal flat environments (Talsi-263 drill- core at 201 m depth); B - Dolomite with pseudomorphs of halite crystals (marked with arrows) in the carbonate mudflat deposits (Dobele-4 drill-core at 598 m depth); C - Dolomite with silty irregular interlaminae (Talsi-263 drill-core at 197-198 m depth); D - Thin-section micro-photo of the sandstone with quartz and feldspar grains (light-colored), dolomite ooids and peloids (dark-colored). Both siliciclastic grains, as well as ooids and peloids are well sorted. Šķaune-103 drill core, 380.2 m, Rēzekne Formation (Stinkulis, 1998); E - Thin-section micro-photo of the ooid, Liozno drill-core (Belarus), around 383 m, Middle Devonian Lower Morsowo beds (analogue of the Pärnu Formation) (Photo by V. Kurshs; Stinkulis, 1998); F - Dolomitic marl with thin silty sandstone laminae (below), overlain by variably-grained sandstone with dolomite ooids and peloids (above). Darker color is stressed by dolomite ooids and peloids, rich in sulphides. Distinct boundary between two facies indicates erosion of the dolomitic marl, Ludza-15 drill-core, 425.0 m, Pärnu Formation (Stinkulis, 1998). The color of the rocks in the logs is close to the original. Near the vertical scale of the log: *dm* – dolomitic marl, *dl* – dolostone, keys for the horizontal scale of the log see in Figure 5.11. [Scale information: pen sharpener in B measures 2 cm]

5.23. attēls. Karbonātisko nogulumu fāciju asociācijas - Fāciju asociācija 1.1. - Vidējais-augšējais plūdmaiņu līdzenums ar karbonātu sedimentāciju, Fāciju asociācija 1.2. - Vidējais-augšējais plūdmaiņu sēklis ar karbonātu sedimentāciju - ar reprezentabliem attēliem un nogulumu griezumu

A - Smilšains dolomīts ar „putna acs" tekstūru (atzīmēta ar bultu), kas norāda uz plūdmaiņu līdzenumu sedimentācijas vidi (Talsi-263 urbums 201 m dziļumā); B - Dolomīts ar halīta kristālu pseidomorfozēm (atzīmētas ar bultu) karbonātiskajos augšējā plūdmaiņu līdzenumu nogulumos (Dobele-4 urbum 598 m dziļumā); C - Dolomīts ar neregulārām aleirītiskām starpkārtām (Talsi-263 urbums 197-198 m dziļumā); D - Dažādgraudains smilšakmens ar kvarca un laukšpata graudiem (gaišie), dolomīta oolītiem un pseidooolītiem (tumšie). Gan drupu graudi, gan oolīti un pseidooolīti ir labi šķiroti, Šķaune-103 urbums, 380.2 m, Rēzeknes svīta (Stikulis, 1998); E - Oolīta plānslīpējuma mikrofotogrāfija, Liozno urbums (Baltkrievija), aptuveni 383 m, vidusdevona Apakšmorsovas slāņi (Pērnavas svītas analogs pēc ģeoloģiskā vecuma) (V. Kurša foto, Stinkulis, 1998); F - Dolomītmerģelis ar aleirītiska smilšakmens starpkārtām (apakšā), kuru pārsedz dažādgraudains smilšakmens ar dolomīta oolītiem un pseidoolītiem (augšā). Tumšpelēko krāsu piešķir dolomīta oolīti un pseidoolīti ar bagātīgu sulfīdu piejaukumu. Krasais kontakts starp abām fācijām liecina par dolomītmerģeļa izskalošanu dažādgraudainās smilts uzkrāšanās laikā, Ludza-15 urbums, 425,0 m dziļums, Pērnavas svīta (Stinkulis, 1998). Iežu krāsa griezumā tuva oriģinālajai. Pie griezuma vertikālās skalas *dm* – dolomītmerģelis, *dl* – dolomīts, apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmuļu asināmais B attēlā – 2 cm]

The thickness of these laminae is less than few mm up to some cm, reaching maximum 60 cm. In places grey and greenish laminae alternate and reach up to 1,5 cm in thickness.

Structureless, plane-parallel laminated and wavy structured dolostone (F23) is also characteristic to the FA 1.1 of the Pärnu RS (Figure 5.23C). It is more widely distributed in the eastern part of the study area, but occurs also in other drill-cores towards the west, usually in the most upper part of Pärnu Fm, under the brecciated beds of Narva Fm, in such drill-cores as Dobele-4 and Vērgale-45. It appears also at the very top of Pärnu Fm in Pape-95 drill-core and is about 1 m thick. Towards north, F23 occurs in outcrop area at the very top of the succession. In Gārzde-1 drill-core the thickness of F23 reaches 3.5 m, in Valga-10 drill-core this facies is intercalated with dolomitic marl with silty and sandy interlaminae (F22) and associated with nodular facies (F24) at the top of Rēzekne Fm. F23 appears also as 2 m thick layer between fine-grained sandstone beds at the very base of Rēzekne Fm in Värska-6 drill-core. In Alūksne-99 it occurs at the upper part of Rēzekne Fm in thickness of 2 m thick layers. Only in Ludza-15 drill-core (Eastern Latvia) dolostones occur twice: comprising 2.5 m at the top of Rēzekne Fm and above the contact between Rēzekne and Pärnu Formations.

This facies is mainly homogenous, and rare lamination is stressed by dark brown clayey, mica and gypsum admixture. Lamination of different scale can be distinguished: 0.1-3 cm thick light and 0.1-0.5 cm thick dark lamina alternation or very thin (0.1-0.2 mm) laminae. In places wave-lamination is present. Birdseyes (fenestral fabric), which are millimetre-size irregular voids, occur rarely in F24 (Figure 5.23A). Desiccation cracks are typical for the dolostone in Vērgale-45 drill-core, while halite pseudomorphs occur in dolostones of Dobele-4 drill-core (Figure 5.23B). Very often brownish gypsum occurs in F23, most often as interlaminae. In F20 and 23 it occurs as lenses, small veins, in sandsones as cement. Gypsum flakes are oriented parallel to lamination with thickness less than 1 mm, length up to 1 cm (Stinkulis, 1998). Sometimes gypsum lenses have halite crystal shapes, thus possibly representing halite pseudomorphs. According to Stinkulis (1998), gypsum occurs also as thin vertical fine veins with lengh up to 2-3 cm and width less than 1 mm.

Interpretation:

The homogenous texture of dolomitic marls and dolostones indicate deposition from suspension in a low energy carbonate-rich setting in intertidal to supratidal environment (Dalrymple, 1992). During diagenetic processes calcium carbonate deposits were replaced by microcrystalline dolomite (e.g. Hardie, 1987; Lasemi et al., 1989; Lasemi et al., 2012).

In episodes of energy increase carbonate mud was partly eroded and siliciclastic material, as well as carbonate ooids and peloids accumulated. Such activation of water flows could have been caused by wave action and/or tidal currents. The irregular distribution of grains and irregular, wavy character of lamination in F21, 22 and 23 indicate that clastic material and carbonate ooid and peloid input was made by wave action. Laminated dolostone with very thinly laminated or occasional wave-ripple-laminated structure also confirms wave influence in the sedimentation and suggests deposition close to fair weather wave base (Reineck, Singh, 1980).

Lamination of variable thickness indicates different scale periodicity of the sedimentation process, caused by tidal processes. Thickness variation of the laminae reflects daily variation of tidal range: neap-spring cycles (Middleton, 1991; Nio, Yang 1991). The ripple lamination in two different directions also indicates the role of tidal processes to deposition in ebb and flood currents.

Desiccation cracks, tepee structures and association of the beds of this FA with brecciated beds of FA 1.3, are indicators of periodical subaerial environment. Periodic exposure of the deposits results in desiccation and the formation of tepees and mud cracks. It is typical for tidal flats, particularly in the upper intertidal and supratidal settings (Tucker, Wright, 1990; Lasemi et al., 2012).

Tepee structures are common to peritidal deposits and form as a result of desiccation, cementation and crystal growth, thermal expansion, and contraction of partially lithified sediment in arid tidal flats (Kendall, Warren, 1987; Lasemi et al., 2012). V-shaped form in cross-section of the desiccation cracks, and unbended upwards laminae indicate that the cracks formed in subaerial environment. Wave action or tidal currents transporting the coarse clastic material, washed on already lithified dolomitic marl and deposited only in the cracks (Stinkulis, 1998).

Birdseyes are also diagnostic feature to tidal flat environments, commonly form as a result of air or gas bubble formation, desiccation shrinkage, wrinkles in the laminated bacterial deposits or development of trapped air bubbles in the pore spaces during flood tide and subsequent rapid cementation (Shinn, 1983, 1986; Lasemi et al., 2012).

Thin lamination and alternation of coarser and finer material are typical for tidal flats (Reineck, Singh, 1980). Only some cm or tenth of cm thick wave lamination itself is an indicator of shallow settings, above the wave base. This facies contains desiccation cracks, irregular to planar birdseyes, tepee structures, gypsum layers, halite pseudomorphs, which are typical features in arid upper intertidal to supratidal mudflats (Elrick, 1995; Lehrmann et al., 2001; Lasemi et al., 2012).

FA 1.2: Intertidal to supratidal carbonate shoals

This Facies Association is characterized by variably-grained sandstone with dolomite ooids and peloids (F19), as well as dolomitic marl with irregularly distributed coarse and medium sized quartz and feldspar grains (F22), dolomitic marl with silty and sandy interlaminae (F20), trough-cross stratified sandstones (F3), wavy laminated dolostone (F23) and sandstone with siltstone laminae (F14). FA 1.2 is found in the eastern part of the study area. The thickness of this FA varies from 2 to 5 m and reaches its maximum of 35 m occurring in several repetitive cycles in Liozno drill-core (see Figure 5.23.).

The most representative facies of this FA is the variably-grained sandstone with dolomite ooids and peloids (F19; Figure 5.23F). Variably-grained sandstone with dolomite ooids and peloids with the presence of greenish-grey dolomitic marl matrix also exist. F19 comprises 4.5 m at the top of Pärnu Fm in Viļāni-11 drill-cores and 2 m at the base of Pärnu Fm in Ludza-15. To the north the facies is almost 5 m thick and occurs at the top of Pärnu Fm in Tsiistre-327 and Mehikoorma-421 drill-cores. The thickness of the beds reaches up to 3.7 m in Višķi-25 drill-core, comprising most of the Pärnu Fm succession. In Drissa-1ST drill-core it occurs as 3-5m thick beds at the top of Pärnu Fm. It comprises 0.5-1 m thick layers at the top of Rēzekne Fm, as well as in Pärnu Fm in Šķaune-103 drill-core. In Liozno drill-core it repetitively alternates with F20 and F19 and forms the bulk of Pärnu Fm.

The deposits of this FA rarely occurs as interlayers in other deposits. The colour of deposits is grey and light-grey, composed of quartz and feldspar as well as ooids and peloids (Figure 5.23D). The bimodal clastic and carbonate material distribution by the size is characteristic: two fractions dominate, the amount of which varies in different layers (Sinkulis, 1998). The bimodal distribution of grain size determines plane-parallel lamination of deposits with laminae thickness of 0.1-0.5 cm. In places flaser bedding is typical, in Višķi-25 drill-core it comprises almost all 3.7 m thick layer. The bedding is stressed by clayey and silty laminae with thickness less than 1 mm up to 0.2-1 cm. In one beds (Šķaune-103 drill-core, depth 381.6-382.6) cross-stratified 1 m thick sandstone beds occur. Cross-stratification is stressed by alternating of finer and coarser sand grains. Both of these fractions are better sorted than in sandstones with flaser bedding (Stinkulis, 1998). The thickness of coarser laminae is 1-5 mm, finer are 2-8 mm thick, but the sets reach up to 15 cm.

Dolomite ooids and peloids are very typical component of the FA 1.1 and are found in drill-cores in the eastern part of the study area only, besides the amount of the grains increases to east and south-east (Figure 5.23E). Thus, in Liozno drill-core almost entire beds consist of F20, with the exception of the very lower part of the section (Kurshs, 1975; Sorokin, 1981). The amount of ooids and peloids varies in F19 from 10 to 70%. They are quite regularly distributed in the sandstone beds. Two quite distinct fractions can be distinguished in the size of ooids, similar to bimodal distribution of siliciclastic material fractions. According to Stinkulis (1998) the maximal size of carbonate ooids is 0.6-0.8 mm, which is smaller than the size of bigger quartz and feldspar grains.

The transition between variably-grained sandstone with dolomite ooids and peloids (F19) and dolomitic marls (F20) is usually gradual. In the section coarser grains gradually disappear, and the amount of finer grains is decreasing, while the amount of dolomitic marl is increasing. In places the contact is very sharp. The structure of dolomitic marls depends on the amount of clastic grains: more sand occurs, more the structure of dolomitic marls is disintegrated (Stinkulis, 1998). In places at the base of beds of F19 pebbles of dolomitic marls occur.

Interpretation:

The variably-grained sandstone with dolomite ooids and peloids (F19) formed as a result of sedimentation of three different by the origin components: quartz and feldspar grains, carbonate ooids and peloids, as well as in places clayey carbonate material. In relatively longer periods of time dolomitic marl deposited from suspension in a low energy carbonate-rich environment, while sand particles and carbonate ooids and peloids were washed during the episodes of high energy environment. As indicated by wavy lamination, the mixing of these components was possibly caused by the wave action, eroding the primary structures of the dolomitic marls and depositing sandy components. However, the role of wave action was not predominant in formation of FA 1.2. The presence of flaser bedding reflects deposition in tidal currents, which is re-confirmed by the bimodal distribution of grain size. Sandstone form by traction currents and mudstone drapes form during slack-water on tidal flats (Dalrymple, 1992). The cross-stratified sandstones indicate deposition by traction currents in 3-D dunes. The alternation of finer and coarser sand grains in it reflects deposition in periodically waning tidal currents in tidal gullies (Dalrymple, 1992).

Ooids commonly are interpreted to represent direct physiochemical precipitation from seawater (Newell et al., 1960; Bathurst, 1975; Deelman, 1978; Duguid et al., 2010), whose formation is favored by: a) water supersaturated with respect to calcium carbonate; b) a source of nuclei; c) a means of agitation (Davies et al., 1978; Sumner, Grotzinger, 1993; Reeder, Rankey, 2008; Rankey, Reeder, 2009). Unlike siliciclastic tidal sands, which must be reworked from earlier deposits or transported to the depositional system from elsewhere, these sedimentary particles can be (and many are) produced *in situ*, with relatively little net transport (Rankey, Reeder, 2012). As reflected by the concentric structure of the ooids, and its association with variably-grained sandstones, it is suggested that the grains formed in periodically waning and waxing currents (Tucker, Wright 1990; Davies et al., 1978; Stinkulis, 1998). Such periodically changing flows were controlled by tidal processes. This is very nicely illustrated in Liozno drill-core, where ooid-rich beds (F19) rhythmically grade few times with the dolomitic marl deposits (F20).

According to Stinkulis (1998) the increase of content of the dolomite ooids and peloids to the east and south-east of the study area, as well as their thin concentric and radial structure suggest that in this part of the basin, the sedimentary environment was favourable for the growth of these concentric carbonate grains. The ooids are interpreted to be deposited in tidal shoals, similar to many recent tidal carbonate sand shoals of the Bahamas, which are dominated by ooids (Rankey, Reeder, 2012). There ooids are most abundant in agitated waters less than 2 m deep (Newell et al., 1960) and the best-sorted and most ooid-rich sediments occur on the crests of parabolic barforms (Rankey, Reeder, 2012).

Flaser bedded, wavy laminated, as well as cross-bedded sandstones together with dolostone laminae and association of this facies with dolomitic marls indicate the formation of FA 1.2 in tidal shoals on intertidal to supratidal mudflats (Lehrmann et al., 2001). The dominance of the sandy material and the presence of coarser sandy grains, as well as dolomite ooids, in comparison to more fine-grained sediments of FA 1.2., suggests deposition in high energy environment, such as in tidal gullies and tidal shoals on intertidal to supratidal mudflats (Brooks et al., 2003; Rankey et al., 2006).

FA 1.3: Palaeosols

Facies Association 1.3 consists of mainly nodular and karstified facies (F24), represented by carbonate nodules, concretions and networks of calcite veins filled with sandstone and mud clasts interbedded with crinkled, thinly laminated carbonate and siliciclastic mudstones with concave-up structures and desiccation cracks. Usually at the top of Pärnu Fm the FA 1.3 is associated with bioturbated facies (F25) and the brecciated beds of Narva Fm (Figure 5.24.).

FA 1.3 occurs in all parts of the study area and forms from few dm to 3 m thick beds. The FA 1.3 occurs as 1 m thick beds in the middle part of Pärnu Fm in Bīriņi drill-core, as 2 m thick beds in the middle part of Jelgava-2 drill-core. It occurs twice in Dobele-4 drill-core in the middle part of the section, as well as on the top of Pärnu Fm associated with bioturbated facies (F25) and wave-laminated sandstone (F13). In Taurkalne-1 drill-core the bioturbated facies occur three times as 0.3-1 m thick beds in Pärnu Fm. Similarly, it occurs in 3 intervals in Svedasai-252 drill-core with the thickest beds of 1.5 m at the top of Pärnu Fm. It occurs also at the top of Pärnu Fm in Talsi-263 drill-core, intercalated repetitively with laminated mudstone (F16) and forming together about 2 m thick beds. F25 appears also possibly at the base of Pärnu Fm in Gārzde-1 drill-core. Bioturbated facies are found also at the very top of Pärnu Fm as a thin layer in Palanga-319, Tartu-453, Valga-10 and Kaagvere drill-cores. It occurs also at the base of Rēzekne Fm as 1 m thick beds, as well as in three more beds above as thin layers in Mehikoorma-421 drill-core. It occurs as 2-2.5 m thick layers in the middle part, as well as a thin layer in the upper of Pärnu Fm in Värska-6 drill-core. In Tsiistre-327 drill-core it appears as 0.5-1.3 m thick layers in the middle and upper part of Rēzekne Fm. In Alūksne-99 and Madona-93 drill-core it possibly forms few m thick beds in the middle part of Rēzekne Fm. It forms almost 3 m thick beds at the top of Rēzekne Fm in Valga-10 drill-core.

The typical structure of the facies of FA 1.3 is fenestral, mud-cracked beds with common disrupted lamination and nodules (see Figure 5.24.). Irregular gypsum layers are present in some of the beds as thin crinkly and discontinuous laminae. Tepee structures and desiccation cracks are usually associated with F24, occurring in dolomitic marls from mm up to some cm big. These facies are interlayered with numerous erosive-based, cross-stratified sandstone facies (F3) of various thicknesses, and very often with laminated siltstone (F17), heterogenous sandstone and siltstone laminated beds (F15) and siltstone with sand laminae (F16).

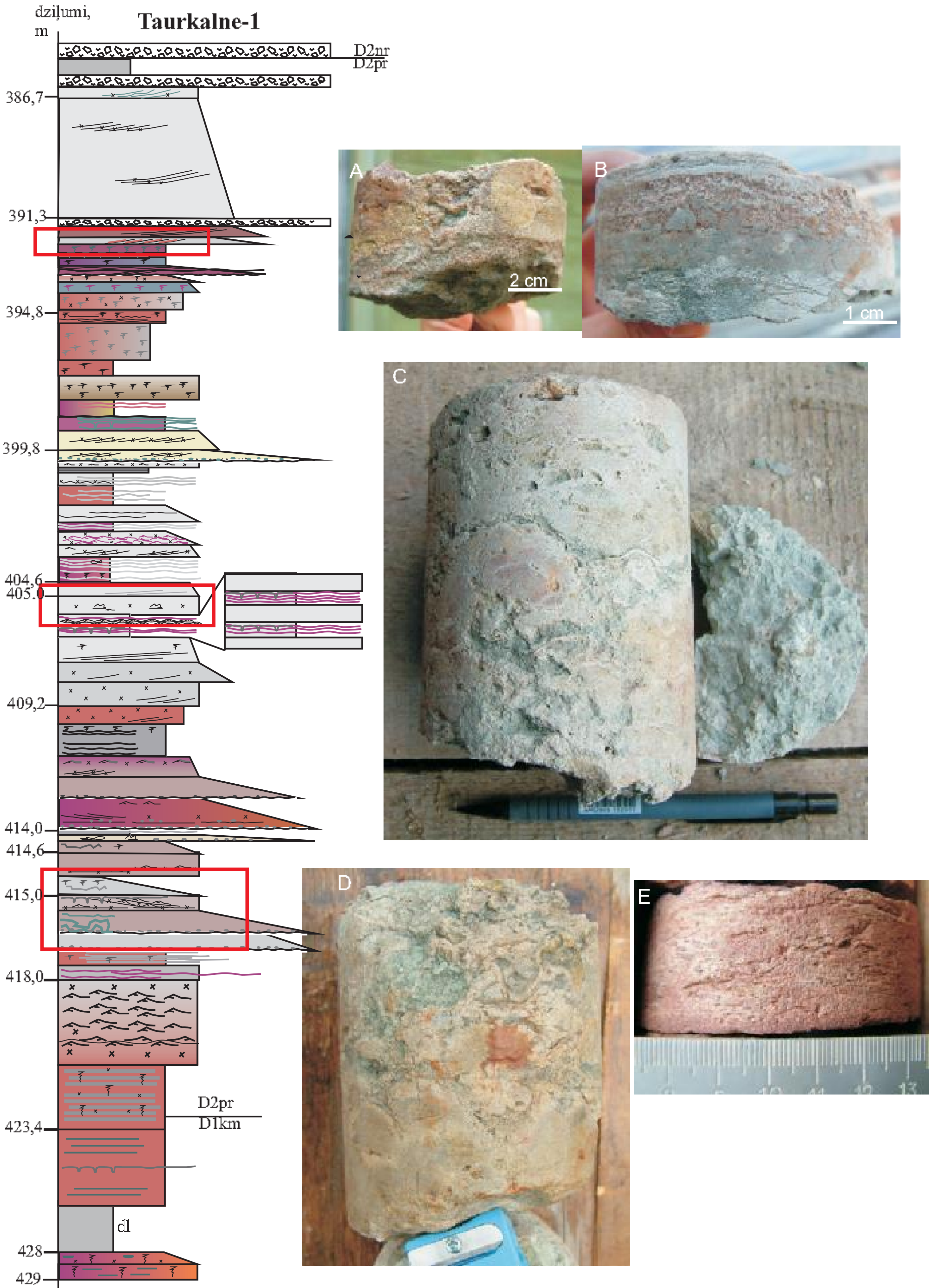

Figure 5.24. See next page
5.24. attēls. Skat. nākamo lappusi

Figure 5.24. Representative photographs and a measured section of the Facies Association 1.3 - palaeosols - silicified crust and karstified beds, crinkle-laminated beds with common features of erosion and subaerial exposure

A - Taurkalne-1 drill-core at 414 m depth; B - Bīriņi drill-core at 387-393 m depth; C - Taurkalne-1 drill-core at 416 m depth; D - Dobele-4 drill-core 595-601 m; E - Jelgava-2 drill-core at 497 m depth. The color of the rocks in the logs is close to the original. Near the vertical scale of the log: *dm* – dolomitic marl, *dl* – dolostone, keys for the horizontal scale of the log see in Figure 5.11. [Scale information: pencil in C measures 15 cm, pen sharpener in D - 2 cm]

5.24. attēls. Fāciju asociācija 1.3. - Paleoaugšņu nogulumi - ar reprezentabliem attēliem un nogulumu griezumu. Nogulumus veido silificēta garoziņa un karstificēti slāņi, slāņi ar krokotu tekstūru, erozijas virmām un atsegšanas subaerālā virsmā pazīmēm

A - Taurkalne-1 urbums 414 m dziļimā; B - Bīriņi urbums 387-393 m dziļumā; C - Taurkalne-1 urbums 416 m dziļumā; D - Dobele-4 urbums 595-601 m dziļumā; E - Jelgava-2 urbums 497 m dziļumā. Iežu krāsa griezumā tuva oriģinālajai. Pie griezuma vertikālās skalas *dm* – dolomītmerģelis, *dl* – dolomīts, apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmulis C attēlā – 15 cm, zīmuļu asināmais D attēlā – 2 cm]

Interpretation:

The nodular and brecciated intervals together with carbonate concretions indicate that FA 1.3 formed during periods of subaerial exposure and nondeposition (Ruskin, Jordan, 2007). The close association of some beds with brecciaed units of Narva Fm, which is interpreted as solution-collapse breccias formed through evaporite dissolution processes (Tänavsuu-Milkeviciene et al., 2009) supports this interpretation. Tepee structures and desiccation cracks could have formed as a result of brecciation of lithified sediment, regardless of their depositional setting (Pratt, 2002). Cemented sandstones with carbonate concretions, nodules, presence of gypsum aggregates and networks of calcite veins are interpreted to be formed in semiarid or arid conditions with low precipitation level (Mack, 1992) as calcretes (Collinson, 1996) in supratidal environment (Warren, Kendall, 1985; Demicco, Hardie, 1994; Pratt, 2002). Modern and ancient arid supratidal deposits generally comprise dolomitic marl which may be interbedded with gypsum and collapse breccias (Lasemi et al., 2012). Butler (1970) suggested that the evaporites and the deformation were the result of evaporation of waters brought to the sabkha during extreme flooding events. Formation of the deposits in arid climate is also supported by the palaeogeographical position of the BDB during the Devonian time (Cocks, Torsvik 2006).

Siliciclastic-rich Facies associations

FA 2.1: fluvial and tidally-influenced fluvial deposits of inner estuary

This FA is very common for the outcrop area and is in details described in Chapter 5.1.2. Facies Association 2.1 consists of variably-grained trough cross-stratified sandstone (F3a), trough cross-stratified mudstone conglomerate (F2), mudstone and siltstone conglomerate (F1), structureless sandstone (F6), plane-parallel stratified sandstone (F5) and occasionally cross-stratified sandstone with mud and mica drapes (F4). It occurs also in most of the described drill-cores (Figure 5.25.). The thickness of FA 2.1 is up to 2-5 m, maximum 10 m. It is distributed in all the territory of the study area. FA 2.1 occurs as 7 m thick beds at the base of Pärnu Fm in Tahkuranna drill-core. In two intervals, comprising 0.5-1 m thick beds, it appears in Bīriņi drill-core.

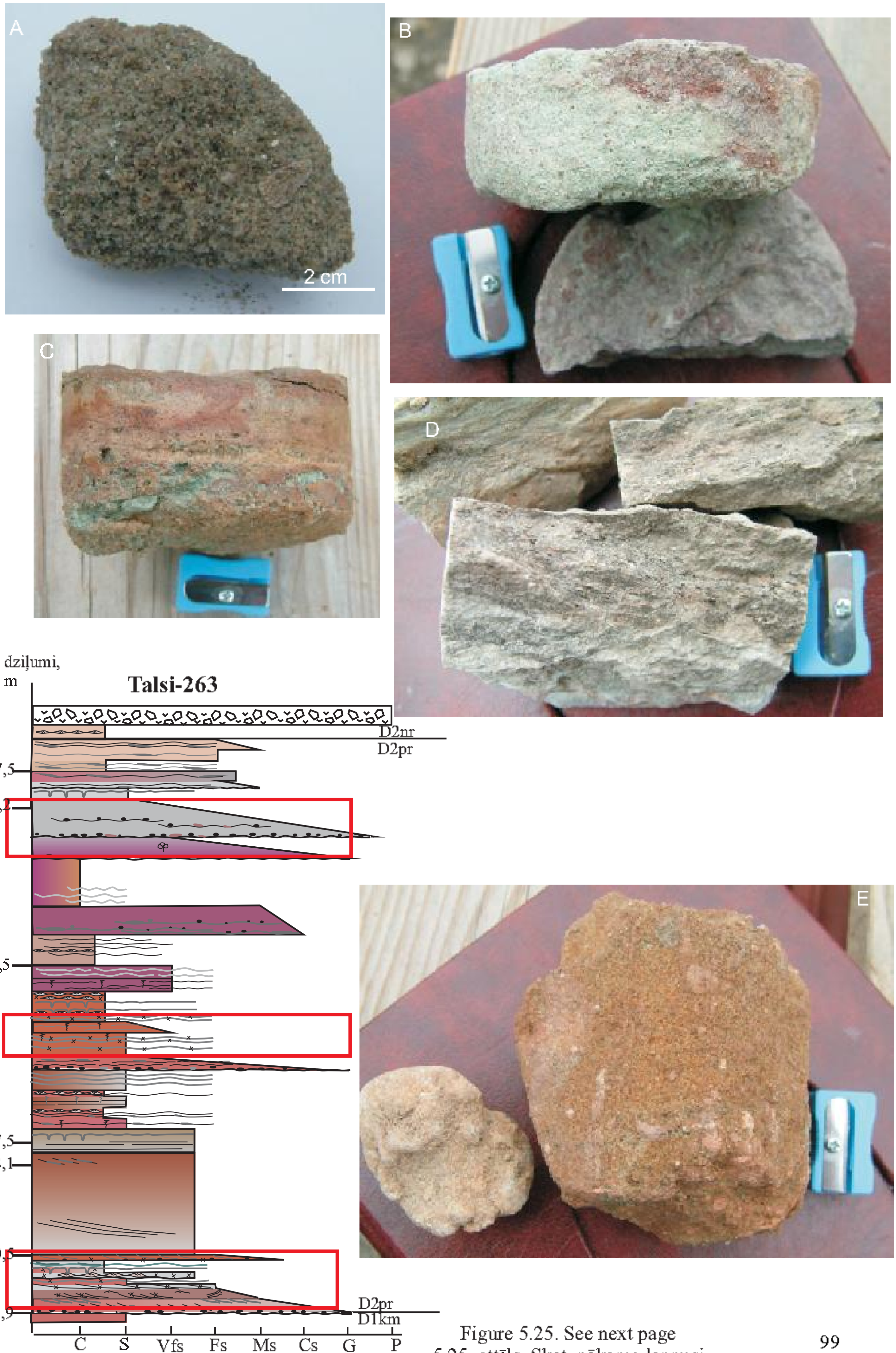

Figure 5.25. See next page
5.25. attēls. Skat. nākamo lappusi

Figure 5.25. Representative photographs and a measured section of the Facies association 2.1 - fluvial and tidally-influenced fluvial deposits of inner estuary

It contains mudstone conglomerate, coarse-grained and variably-grained, unsorted sandstone with fossil fish and plant remains. A - Talsi-263 drill-core at 211 m depth; B - Talsi-263 drill- core at 201 m depth; C - Talsi-263 drill-core at 204 m depth; D - Talsi-263 drill-core at 199 m depth; E - Talsi-263 drill-core at 210 m depth. The color of the rocks in the logs is close to the original. Keys for the horizontal scale of the log see in Figure 5.11. [Scale information: pen sharpener in B, C, D, E measures 2 cm]

5.25. attēls. Fāciju asociācijas 2.1: Estuāra iekšējās daļas fluviālie un plūdmaiņu ietekmētie fluviālie nogulumi ar reprezentabliem attēliem un nogulumu griezumu

To veido māla konglomerāts un rupjgraudains, kā arī dažādgraudains, vāji šķrots smilšakmens ar fosīlo augu un zivju atliekām. A - Talsi-263 urbums 211 m dziļumā; B - Talsi-263 urbums 201 m dziļumā; C - Talsi-263 urbums 204 m dziļumā; D - Talsi-263 urbums 199 m dziļumā; E - Talsi-263 urbums 210 m dziļumā. Iežu krāsa griezumā tuva oriģinālajai. Apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmuļu asināmais B, C, D, E attēlos – 2 cm]

The FA 2.1. occurs in 3 intervals in Talsi-263 drill-core, from 0.3 to 2 m thick units at the base, in the middle part and the topmost part of the Pärnu Fm. In Vērgale-45 drill-core FA 2.1 occurs as a small layer in the upper part of the Pärnu succession. In Mehikoorma-421 drill-core it appears as small interlayer in the middle part of Pärnu Fm. The beds of unsorted sandstone are present at the base of Rēzekne Fm comprising almost 4 m thick beds in Tsiistre-327 drill-core and similarly almost 2 m thick in Tartu-453 drill-core. Several intervals of this FA appear possibly in Pärnu succession in Valga-10 drill-core. In Cīrulīši drill-core fluvial facies occurs at the base of Pärnu Fm and reach 3.8 m in thickness, it appears again as 1.5 m thick beds in the middle part of the succession. FA 2.1 occurs in three intervals in Taurkalne-1 drill-core, in association with palaeosol deposits of FA 1.3. Eastwards, in Atašiene-9 drill-core, fluvial deposits of FA 1.3 comprise the lower part of Rēzekne Fm and reach almost 6 m in thickness, it occurs also at the boundary between Rēzekne and Pärnu Fm. Similarly, it comprises the base of Rēzekne Fm and occurs as 2 thick beds at the base of Pärnu Fm in Drissa-1ST drill-core. In Svedasai-252 drill-core the deposits of FA 2.1 occur at the very base of Pärnu Fm and form about 2.5 m thick units in association with palaeosols of FA 1.3. It possibly occurs as 2 m thick beds in the middle part of Pärnu succession in Jelgava-2 drill-core. Fluvial facies of FA 2.1 comprises about 9 m thick beds at the base of Pärnu Fm in Staicele-4 and possibly 10 m in Ruhnu-500 drill-cores. Therefore, the thickest intervals of FA 2.1 appear in the N-NW, as well as SE part of the study area. Deposits of FA 2.1 are possibly present also in other drill-cores from the study area, however these are not listed here because of the unclear evidence for interpreting these beds as fluvial deposits.

The most typical facies of this FA is the variably-grained trough cross-stratified sandstone (F3; Figure 5.25A and B). The material is composed of poorly-sorted grains are subrounded and angular. In places clay pebbles are typical, as well as small fossil fish and plant remains, and rarely some dolomite peloids are present (Figure 5.25B, D, E). The depositional units are based by significant erosion surfaces with mudstone rip-up clasts. The amount of erosion varies from few dm to several metres, and reaches up to 15 m. In Šķaune-103 drill-core in the depth of 401-403 m trace fossils occur, they are filled in by dolomite. The cross-stratification, as well as plane-parallel lamination is in places stressed by mica drapes. The architecture of beds and more detailed description on composition of this FA see in Chapter 5.1.2 under FA 1: Fluvial deposits and FA 2: Tide-influenced fluvial and tidal channel deposits.

Interpretation:

The dominance of the poorly sorted material, subrounded and angular grains of the sandy material, occurrence of these facies at the channel scours and channel bases, and the association with erosion surfaces indicate deposition in fluvial currents, not far from the source area. The mudstone rip-up clasts at the lower part of the sandbodies are interpreted to represent channel bases. It is confirmed also by the cross-stratified and ungraded structure of sandstones, that suggests deposition in high energy environment from traction currents by migration of 3-D dunes in fluvial channels and rapid sedimentation (Collinson, 1996). However the presence of some cross-stratified units with mica drapes, as well as plane-parallel lamination, indicate episodes of deposition in lower flow regime. Occasional mica drapes might suggest that the fluvial channels were influenced by tidal currents (Nio, Yang, 1991). Thus it is suggested that the deposition took place in fluvial channels and bars, where tidal currents reached the system at spring tides (Nio, Yang, 1991).

FA 2.2: central estuary deposits

Facies Association 2.2 consists of trough cross-stratified sandstone (F3), structureless sandstone (F6), plane-parallel stratified sandstone (F5), cross-stratified sandstone with mud and mica drapes (F4), current ripple cross-laminated sandstone (F7), compound cross-stratified sandstone (F11) and sandstone with siltstone laminae (F14). FA 2.2 is found in the following drill-cores: Valga-10; Taurkalne-1; Atašiene-9; Mehikoorma-421; Värska-6, Bīriņi. The thickness of the FA is usually from few m to 5-6 m, in Värska-6 drill-core it reaches almost 10 m (Figure 5.26.).

The succession consists mainly of fine-grained sandstones with heterolithic strata and relatively higher content of mud drapes and silt/mud laminae in comparison to usually underlying fluvial and tide-influenced fluvial deposits (FA 2.1.), as well as overlying tidal bar deposits (Figures 5.26D and E). Very often the individual beds consist of sandstone and siltstone couplets. In places compound cross-stratified sandstones with mud drapes and lateral accretion sets are present.

Interpretation:

The fine-grained texture of sandstone and its association with mud drapes and silt/mud laminae combined with the compound and inclined structures suggest sedimentation in bars by tidal channels. The regular occurrence of mud drapes and mud/silt laminae together with sandstone (heterolithic cross-strata) suggests deposition of sand by traction currents and mudstone drapes during slack-water episodes on tidal flats in subtidal environment. The fine interlamination of sandstone and mudstone in tidal point bars is interpreted as reflecting deposition during single tidal cycles (Bridges, Leeder, 1976). The inclined heterolithic strata are interpreted as lateral accretion deposits on tidal point bars and indicate that the FA 2.2 was possibly deposited in relatively higher sinuosity channels. The relatively fine-grained texture of this FA, its close association in the section with fluvial and tide-influenced fluvial deposits (FA 2.1) at the bottom and tidal bar deposits at the top (FA 2.3) suggest deposition in a central estuary, where bedload convergence zone prevails, which is characteristic with the fine-grained material (Dalrymple et. al., 1992; Dalrymple, Choi, 2007).

It must be noted that the central estuary FA is described also for the outcrop area and more data if given in Chapter 5.1.2 under FA 2: Tide-influenced fluvial and tidal channel deposits. However, it very much differs from it counterparts found in many drill-cores throughout the basin by its coarse-grained composition of deposits.

FA 2.3: tidal channels and bars

Facies Association 2.3 is found almost in all the drill-cores with significant thicknesses. It is also common or the outcrop area (see Chapter 5.1.2). It forms the thickest units along the northern part of the distribution area of the Rēzekne and Pärnu RS. The thinnest beds of FA 2.3 are present in the western part of the study area: in drill-cores Jelgava-2 and Dobele-4 it appears as 4-5 m thick beds at the bottom of Pärnu Fm. It is not present in Vērgale-45 drill-core. Thus, the thickness of FA 2.3 in Ruhnu-500, Staicele-4, and Valga-10 drill-core reaches about 20 m, while the maximum thickness occurs in Tsiistre-327 drill-core, reaching almost 30 m. The FA usually occurs as erosively based sandstone bodies. It is in details described from the outcrop area. FA 2.3 consists of trough cross-stratified mudstone to siltstone conglomerate (F2), well-sorted, trough cross-stratified sandstones (F3b), cross-stratified sandstone with mud and mica drapes (F4), large scale cross-stratified sandstone (F10) and compound cross-stratified sandstone (F11). Current ripple cross-laminated sandstone (F7), as well as climbing ripple cross-laminated sandstone (F8) is also a typical facies for this FA (see Figure 5.26.).

The FA 2.3. is characterized by some tens cm up to 5 m thick beds with well sorted material, very well rounded grains, common mica and mud drapes, and with mudstone pebbles, that most typically occur at the bottom of the beds (2-4 cm), sometimes fossil fish remains are found and probably fossil plants as well (according to Stinkulis (1998) black flakes in Višķi-25 at the depth of 362 and 362.6 m; see Figures 5.26A, B, C). In some places reactivation surfaces occur, associated with mud drapes.

Below the FA is associated with fluvial and tide-influenced fluvial deposits (FA 2.1) usually with an erosional surface, or with more fine-grained central estuary deposits (FA 2.2). At the top it is associated with usually outer estuary tidal ridges (FA 2.4) or tidal flat deposits (FA 2.5 and FA 2.6). The architecture of the beds, derived from outcrop data, is described in more details in Chapter 5.1.2 under FA 3: Tidal bar deposits.

Interpretation:

The extensive cross-stratification, large scale cross-strata, fine-grained and well-sorted material, mud and mica drapes, and erosional bases, suggest deposition in subtidal channels and bars. Reactivation surfaces are formed as the subordinate current erodes the lee-face of the dune formed by the preceding dominating current (Dalrymple, 1992). Mud drapes are deposited from suspension settling during slack-water periods (Nio, Yang, 1991). The depositional units are based by significant erosion surfaces with mudstone rip-up clasts, interpreted to represent channel bases. The upward-fining large-scale cross-stratified sets (F10) are interpreted to represent bars in the tidal channel since the upwards fining reflects upwards decrease in channel energy, as opposed to dunes that reflect highest current energy at the top (Miall, 1996). This is confirmed by the great thickness of these cross-strata. Recognition of mud couplets and tidal bundling in channel fills of FA 2.3 provides evidence of the subtidal origin of these deposits. Unlike other structures commonly found in tidal deposits mud couplets and tidal bundling caused by channelized tidal currents are unique to tidal deposits (Rahmani, 1988). The presence of mud drapes and relatively finer-grained texture of the beds of this FA, allows to distinguish it from the tidal ridges of outer estuaries (FA 2.4.), that are usually of more coarse-grained texture and contain less mud and mica drapes. Thus, the FA 2.3 is interpreted to be deposited in the proximal parts of outer estuaries.

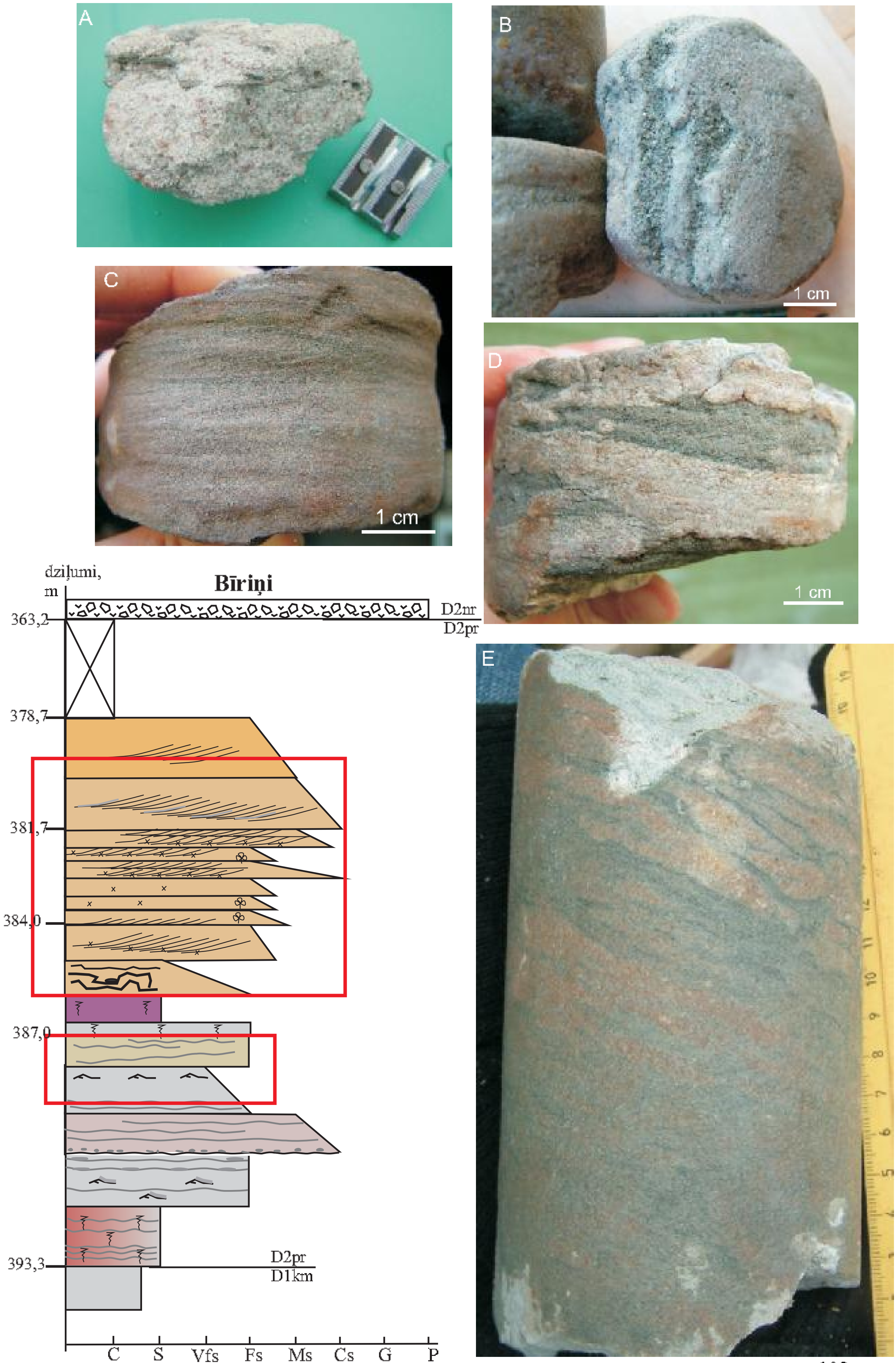

Figure 5.26. See next page.5.26. attēls. Skat. nākamo lappusi

Figure 5.26. Representative photographs and a measured section of the central estuary deposits (Facies Association 2.2) and tidal channel and bar deposits (Facies Association 2.3)

A - Coarse-grained sandstone with mud drapes (Valga-10 drill-core at 253-249 m depth); B - Quartz-rich sandstone with current-ripple lamination stressed by mica (Taurkalne-1 drill-core at 391-386 m depth); C - Ripple-laminated sandstone (Taurkalne-1 drill-core at 420 m depth); D - Slightly-inclined cross-beds with climbing-up ripple-lamination, stressed by mica drapes (Taurkalne-1 drill-core at 416 m depth); E - Irregular and slightly inclined heterolithic srata, interpreted as lateral accretion deposits in tidal point bars of the central estuary (Cīrulīši drill-core at 430-428 m depth). The color of the rocks in the logs is close to the original. Keys for the horizontal scale of the log see in Figure 5.11. [Scale information: pen sharpener in A measures 2 cm]

5.26. attēls. Centrālā estuāra (Fāciju Asociācija 2.2) nogulumi, kā arī plūdmaiņu sēru un grēdu nogulumi (Fāciju Asociācija 2.3) ar reprezentabliem attēliem un nogulumu griezumu

A - Rupjgraudains smilšakmens ar māla kārtiņām (Valga-10 urbums 253-249 m dziļumā); B - Ar kvarcu bagāts smilšakmens ar straumju ripsnojuma pazīmēm, kuru pasvītro vizla (Taurkalne-1 urbums 391-386 m dziļumā); C - Smilšakmens ar straumju ripsnojuma tekstūru (Taurkalne-1 urbums 420 m dziļumā); D - Lēzeni krītošas slīpslāņotas smišakmens sērijas ar kāpjošu ripsnojumu, kuru pasvītro vizlu kārtiņas (Taurkalne-1 urbums 416 m dziļumā); E - Neregulāri slāņoti, lēzeni krītoši heterolītiski nogulumi, kas tiek interpretēti kā laterālās akrēcijas virsmas plūdmaiņu sērē, estuāra centrālā daļā (Cīrulīšu urbums 430-428 m dziļumā). Iežu krāsa griezumā tuva oriģinālajai. Apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmuļu asināmais A attēlā – 2 cm]

FA 2.4: Tidal ridges of outer estuary

Facies Association 2.4 consists of mainly large scale cross-stratified sandstone (F10) and compound cross-stratified sandstone (F11), well-sorted, trough cross-stratified sandstones (F3b), as well as structureless sandstone (F7), trough cross-stratified mudstone to siltstone conglomerate facies (F2), rarely cross-stratified sandstone with mud and mica drapes (F4). FA 2.4 occurs extensively in NW and NE parts of the study area, as well as in places in southern Estonia and northern Latvia (Valga-10, Staicele-4 and Bīriņi cores). Thinner units of FA 2.4 are up to 2 m thick, and they are almost absent in central to southern parts of the study area, such as in Jelgava-2, Dobele-4, Talsi-263, Vērgale-45 drill-cores. The FA 2.4 is missing also in eastern parts of the study area.

This FA consists of usually erosionally based, sand-rich, 2-15 m thick packages of sandstone. FA 2.4 is volumetrically dominated by compound cross-stratified sandstone (F11) and large scale cross-stratified sandstone (F10). It consists of quartz-rich, well-sorted, usually medium to coarse-grained sandstone. The amount of mud and mica drapes is smaller than in other FAs. Compared to central to outer estuary tidal bars, the sandstones here are coarser-grained (medium- to coarse- rather than fine- to medium-grained), the sets of cross-strata are generally thicker.

Interpretation:

The erosional base, lateral accretion, coarse grain-size, together with the significant height of the sandbodies, suggest deposition in tidal ridges in a subtidal environment. These bedforms are also called elongate sand bars (Dalrymple et al., 2012). Such large tidal ridges are characteristic for seaward portions of macrotidal environments (Dalrymple et al., 1990). The lower amount of mud and mica drapes in this FA indicates rapid migration of these barforms and low suspended-sediment concentration (Dalrymple et al., 1990; Dalrymple, 1992; Dalrymple et al., 2012). The more quartz-rich composition and better degree of sorting of sandstones suggest that the sediments were supplied by tidal currents (Plink-Björklund, 2005).

FA 2.5: intertidal to subtidal channels and flats

This Facies Association is characterized by plane-parallel laminated sandstone (F5), trough-cross stratified sandstones (F3b), wavy laminated sandstone (F13), sandstone with siltstone laminae (F14), as well as current ripple cross-laminated sandstone (F7), sandstone with deformation structures (F9) and bioturbated facies (F25).

FA 2.5 is found almost in all drill-cores, with the exception of eastern part of the study area, where it is uncommon (it occurs only in Ludza-15 and Atašiene-9, Pärnu Fm), and where carbonate tidal flats prevail. The FA reaches the largest thickness in western and central parts of the study area, such as in Talsi-263, Vērgale-45, Jelgava-2, Taurkalne-1 and Dobele-4 drill-core, where almost entire succession is composed of FA 2.5 associated with FA 2.6. As for the rest of the basin, the thickness varies from few metres to 5 m, maximum 10 m, and the FA is present in association with mainly tidal bar and channel facies of FA 2.3 (Figure 5.27.).

Heterolithic layers of sand- to coarse silt-sized and mud-sized sediments showing flaser to wavy bedding is typical for this FA (Figures 5.27E and F). This millimeter- to centimeter-scale lamination, referred to as rhythmites (Reineck, Singh, 1980) and tidal bedding or heterolithic stratification (Demicco, Hardie, 1994) are very common (Figures 5.27A-C and D-G). Lamination is stressed by the different-sized material or fine material interlayers, usually siltstone or dolomitic marls. The cross-stratification occurs very rarely. Deposits are laterally discontinuous and normally are capped by fine-grained intertidal deposits. The plane-parallel lamination has the laminae thickness less than mm up to 1 cm, rarely 3 cm. According to Stinkulis (1998) in these facies quartz and feldspar grains have bimodal distribution by their size – two fractions with grain size of 0.05-0.2 mm and 0.3-0.8 mm are typical. Deformation structures occur in places as small folds, as well as irregular laminae. Very often entire beds with sandy and silty laminae are wavy and deformed. Bioturbation features are common.

Interpretation:

This millimeter- to centimeter-scale lamination, referred to as rhythmites (Reineck, Singh, 1980) or heterolithic stratification (Demicco, Hardie, 1994) forms due to decreasing tidal current energy and the resulting change in sand to mud ratio in a landward direction, and represents deposition in the lower to middle intertidal settings. Sediment accumulation under these circumstances incorporate the mud and produce what is called *flaser bedding* (Reineck, Wunderlich, 1968) after the German word for “flame”. Widely distributed flaser bedding indicates that sandstone formed by traction currents and mudstone drapes formed during slack-water on tidal flats in a more subtidal settings. Depending on the relative amount of mud versus sand in these successions, wavy bedding or lenticular bedding occur. Because of the alternation of sand and mud on a regular basis, tidal processes are interpreted to control sedimentation. These types of bedding are called heterolithic, because they contain both relatively coarse sediment and fine sediment. They are very common on intertidal environments and are present in some subtidal environments such as estuaries, and the distributaries of deltas. They are easily preserved in the stratigraphic record and are an excellent tidal signature (Dalrymple, 1992, Lasemi et al., 2012). Tidal rhythmites are typically associated with intertidal flats that border marine or estuarine water bodies. Cross-stratification indicates deposition by high energy ebb or flood tidal currents in 3-D dunes or bars.

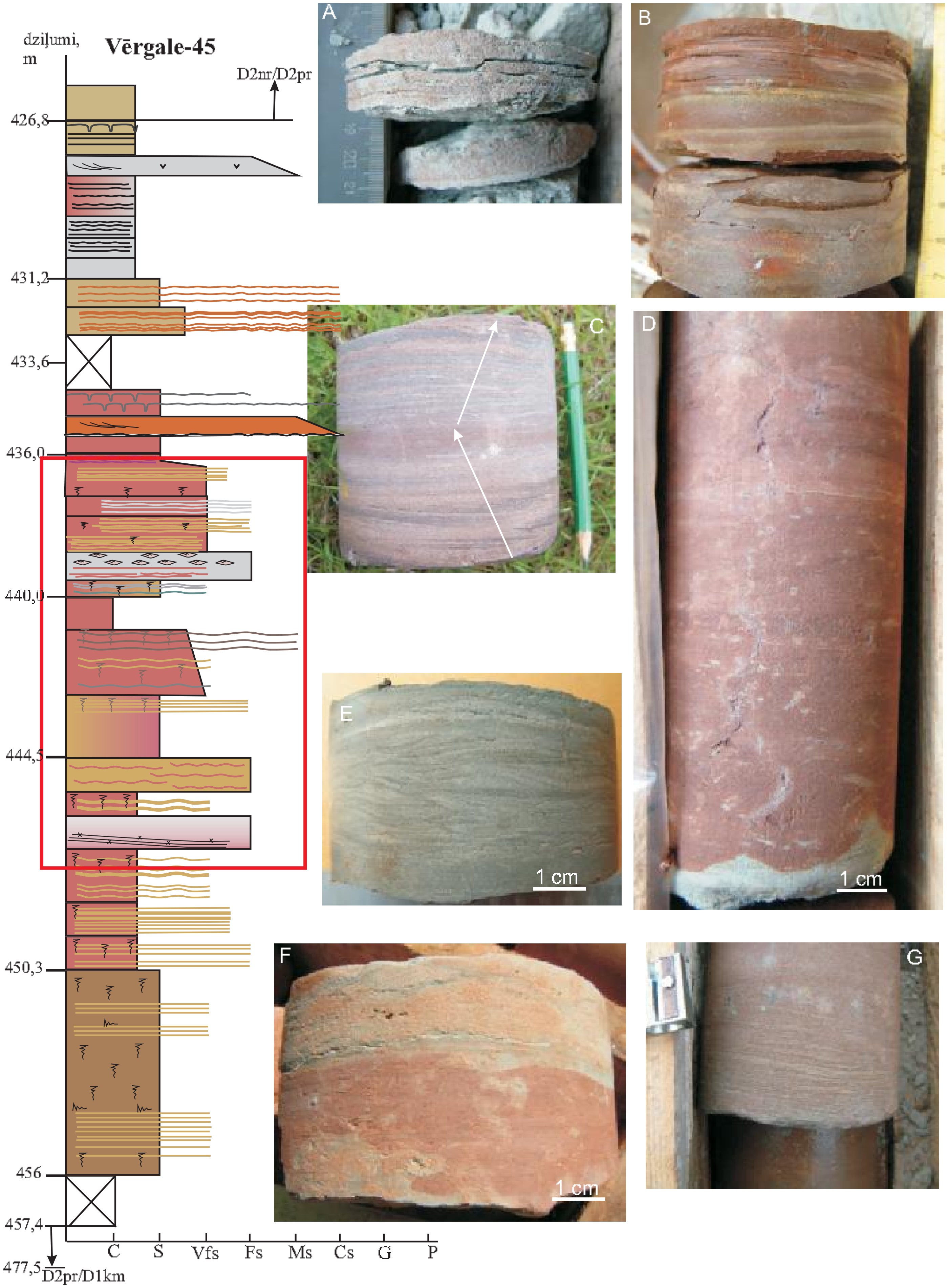

Figure 5.27. See next page
5.27. attēl. Skat. nākamo lappusi

Figure 5.27. Representative photographs and a measured section of Facies Association 2.5 - intertidal to subtidal channels and flats

A - Wavy-laminated sandstone with mud drapes (Jelgava-2 drill-core at 467 m depth); B - Very fine-grained sandstone with coarse sand lamination (Talsi-263 drill-core at 205 m depth); C - Current-ripple laminated sandstone that grades upwards into heterolithic sandstone and siltstone lamination, that grades again into thicker sandy laminae, indicating deposition in waning and waxing currents (Valga-10 drill-core at 263-258 m); D - Fine-grined sandstone with possible desiccation crack and some trace fossils at the bottom of the bed (Jelgava-2 drill core at the depth 471-474 m); E - Wavy-laminated fine-grained sandstone - an indicator of very rare wave activity in the basin (Jelgava-2 drill-core at the depth of 465 m); F - Ripple-laminated light-grey sandstone that erosively overlies very fine-grained reddish sandstone (Valga-10 drill-core at 262 m depth); G - Heterolithic siltstone and sandstone lamination (Valga-10 drill-core at 263-258 m depth). The color of the rocks in the logs is close to the original. Keys for the horizontal scale of the log see in Figure 5.11. [Scale information: pencil in C measures 12 cm; pen sharpener in G measures 2 cm]

5.27. attēls. Fāciju Asociācija 2.5: Vidējā-apakšējā plūdmaiņu līdzenuma un kanālu nogulumi (ar klastisku sedimentāciju) ar reprezentabliem attēliem un nogulumu griezumu

A - Viļņoti slāņots smilšakmens ar māla kārtiņām (Jelgava-2 urbums 467 m dziļumā); B – Ļoti smalkgraudains smilšakmens ar rupjgraudainas smilts starpkārtām (Talsi-263 urbums 205 m dziļumā); C - Smilšakmens ar straumju ripsnojuma tekstūru, kas augstāk pāriet smilšakmens un aleirolīta slāņmijā, kuru atkal nomaina biezāki smilšaini slāņi, liecinot par nogulumu veidošanos periodiski mainīgos hidrodinamiskos apstākļos (Valga-10 urbums 263-258 m dziļumā); D - Smalkgraudains smilšakmens ar, iespējams, žūšanas plaisām un pēdu fosīlijām slāņkopas apakšdaļā (Jelgava-2 urbums 471-474 m dziļumā); E - Viļņoti slāņots smalkgraudains smilšakmens, tā reti sastopamā tekstūra liecina par viļņošanos procesu pakārtotu lomu baseina nogulumu sedimentācijā (Jelgava-2 urbums 465 m dziļumā); F - Gaišpelēks smilšakmens ar ripsnojuma pazīmēm, kas ar izskalojumu uzguļ sīkgraudainam sarkanbrūnam smilšakmenim (Valga-10 urbums 262 m dziļumā); G - Heterolītisks iezis ar smilšamens un aleirolīta slāņmiju (Valga-10 urbums 263-258 m dziļumā). Iežu krāsa griezumā tuva oriģinālajai. Apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmulis C attēlā – 12 cm, zīmuļu asināmais G attēlā – 2 cm]

The plane-parallel laminated sandstone (F5) contains no mud along laminae and is interpreted to be deposited from upper flow regime traction sedimentation by flood currents (Harms, 1975). General absence of subaerial exposure features indicates deposition in intertidal to subtidal settings. Deformation structures, that occur in places as small folds, as well as irregular laminae, probably were produced by wave action.

FA 2.6: intertidal to supratidal mudflat

Facies Association 2.6 consists of homogeneous siltstone and mudstone (F18), laminated siltstone (F17), bioturbated facies (F25), mudstone/siltstone with sand laminae with lenticular bedding (F16), heterogenous sandstone and siltstone lamination (F15), wavy laminated sandstone (F13), and in less degree of sandstone with siltstone laminae (F14) and plane-parallel laminated sandstone (F5). FA 2.6 occurs extensively in NW part of the study area and in central Latvia. It is closely associated with FA 2.5. It is very common in Jelgava-2 and Dobele-4 drill-core, and reaches its maximal thickness in Vērgale-45 drill-core - 3-5 m, and forms the bulk of the section in association with FA 2.5 (Figure 5.28.).

In places pure clayey silt material is distributed, such as in the upper part of Rēzekne Fm (Šķaune-103) and lower part of Pärnu Fm (Ludza-15). The colour is greenish-grey, the only difference with dolostones and dolomitic marls is lack or very small amount of carbonate material and higher amount of clastics. Crinkle-laminated sandstone and mudstone layers ar typical, in which the lenticular bedding prevails (Figure 5.28B). Desiccation cracks occur in places (Figures 5.28C, E, F). The beds are often associated with the karstified FAs on top (Figure 5.28D).

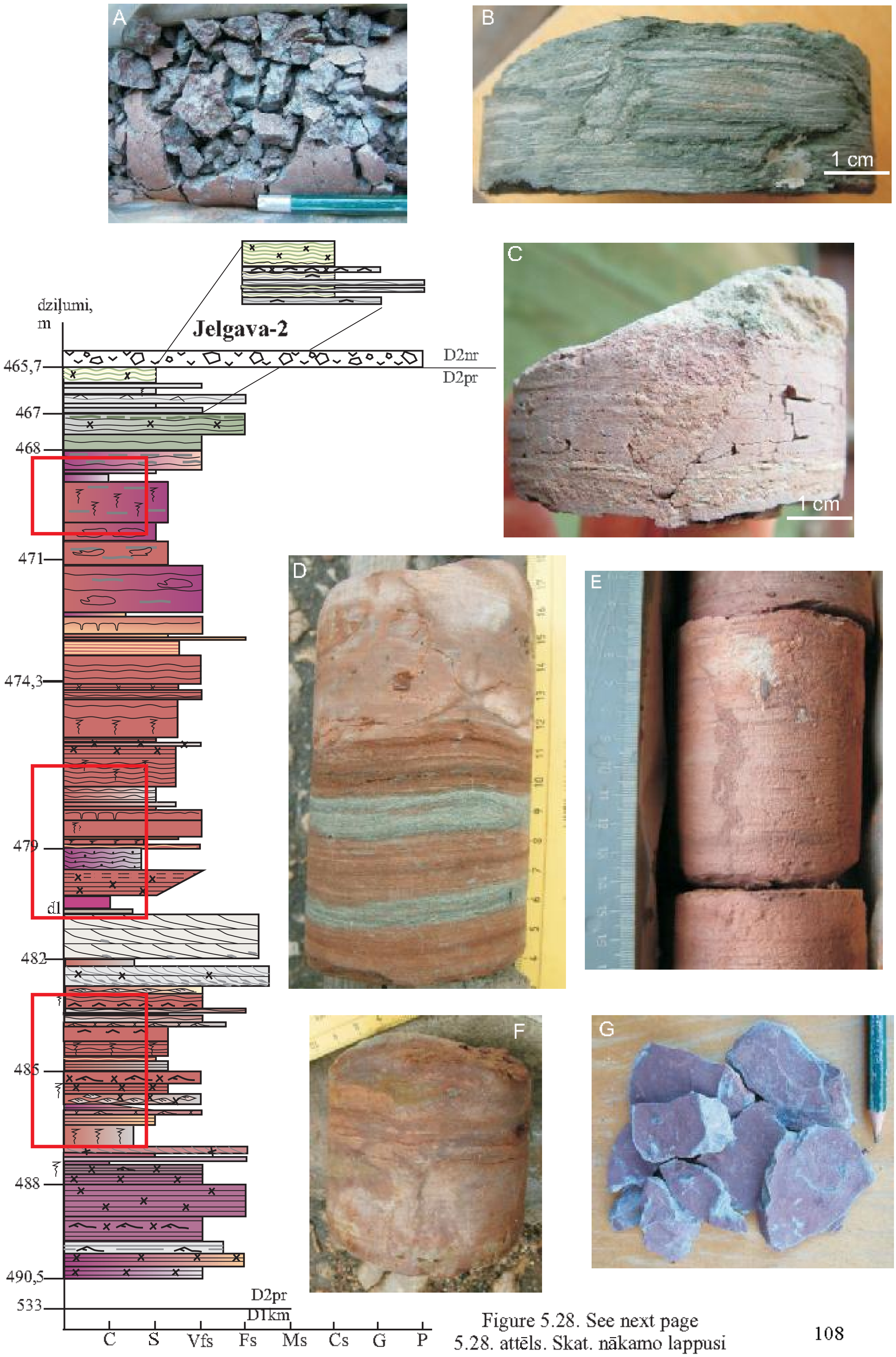

Figure 5.28. See next page
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Figure 5.28. Facies Association 2.6 - intertidal to supratidal mudflats - with representative photographs and a measured section

A - Bioturbated siltstone beds (Taurkalne-1 drill-core at 399-394 m depth); B - Crinkle-laminated sandstone and mudstone (Jelgava-2 drill-core at 463 m depth); C - A desiccation crack in a silty sandstone, filled with coarse, sandy material (Vērgale-45 drill-core at 434 m depth); D - Heterolithic lamination with cemented beds possibly karstified at the top (Talsi-263 drill- core at 202 m depth); E – Very fine-grained sandstone with a desiccation crack (Jelgava-2 drill-core at 479-474 m depth); F - Karstified beds with desiccation features (Talsi-263 drill-core at 210-211 m depth); G - Sandy siltstone with trace fossils (Taurkalne-1 drill-core at 394 m depth). The colour of the rocks in the logs is close to the original. Keys for the horizontal scale of the log see in Figure 5.11. [Scale information: part of the pencil in A measures 7 cm, part of the pencil in G – 4 cm]

5.28. attēls. Fāciju Asociācija 2.6 - Vidējā-augšējā plūdmaiņu līdzenuma nogulumi (ar klastisku sedimentāciju) - ar reprezentabliem attēliem un nogulumu griezumu

A - Alerolīta slāņi ar bioturbācijas pazīmēm (Taurkalne-1 urbums 399-394 m dziļumā); B - Krokoti slāņots smilšakmens un aleirolīts (Jelgava-2 urbums 463 m dziļumā); C - Žūšanas plaisa aleirītiskajā smilšakmenī, kuru aizpilda rupjgraudains smilts materiāls (Vērgale-45 urbums 434 m dziļumā); D - Heterolītiska nogulumu slāņmija ar stipri cementētiem, iespējams, augšdaļā karstificētiem slāņiem (Talsi-263 urbums 202 m dziļumā); E - Sīkgraudains smilšakmens ar žūšanas plaisu (Jelgava-2 urbums 479-474 m dziļumā); F - Karstificēti slāņi ar žūšanas pazīmēm (Talsi-263 urbums 210-211 m dziļumā); G - Smilšains aleirolīts ar pēdu fosīlijām (Taurkalne-1 urbums 394 m dziļumā). Iežu krāsa griezumā tuva oriģinālajai. Apzīmējumus griezuma horizontālajā skalā skat. 5.11. attēlā. [Informācija mērogam: zīmuļa daļa A attēlā – 7 cm, zīmuļa daļa G attēlā – 4 cm]

Deposits are in places bioturbated by organisms leaving vertical to sub-horizontal traces (F25). In places very many rounded and elongated aggregates are filled with possibly dolomitic cement, and are oriented in different directions, which do not match with the main structure of deposits. The size of these aggregates is 0,6-3 cm, and they are possibly trace fossils (see Figures 5.28A, G).

Interpretation:

The thick pure mudstone beds indicate deposition in low-energy environment, possibly in intertidal ponds. However the majority of these beds containing laminae of sandy material, indicate rather alternating deposition from suspension and from traction currents, represented by tidal currents in intertidal to supratidal environment. The mudstone layers thus represent deposition by the waning current during the slack water period of a tidal cycle. Lenticular bedding (F16), which is very typical for this FA, indicates deposition from reversing tidal currents (Reineck, Singh, 1980) and represents an intertidal environment (Dalrymple, 1992). When the tidal currents are too slow to produce ripples, they produce sand layers from suspension that alternate with mud layers (Dalrymple, 1992), as seen in plane-parallel laminated sandstones and mudstones of F15.

Plane-parallel laminated mudstone with desiccation cracks (F17) is interpreted to have formed on top of bar surfaces. Desiccation features can develop on intertidal surfaces but may also form in various other environments that have no tidal influence such as fluvial flood plains. However desiccation features that are associated with tidal conditions and with other tidalites can also serve as a distinctive tidalite signature (Davis, 2012). In arid climate due to high salinity and evaporite formation, burrowers and browsers are practically absent in the upper intertidal deposits, whereas, the lower intertidal facies are intensely bioturbated, as indicated by facies 25.

The wide distribution of lenticular beds and laminated mudstones with sandstone layers suggests deposition on intertidal to supratidal mudflats, where the current energy is lower, and subaerial exposure is frequent.

5.2.2. Basin evolution

The Rēzekne and Pärnu RS are characterised by very distinct and laterally variable facies. Open sea situated west of the epicontinental basin and was partly separated from it by the northwest-southeast barrier - northern part of the Belarus-Mazurian anteclise (Kurshs, 1975; 1992). During the transgression in Rēzekne-Pärnu time, it was also connected to the basin in Moscow syneclyse in the east (Kurshs, 1975). In previous studies, the deposits of Rēzekne and Pärnu RS were interpreted to have accumulated either in shallow marine environments or fluvial channel or distributary fills and delta deposits (Kurshs, 1992; Kleesment, 1997; Plink-Björklund, Björklund, 1999).

The vertical cyclic pattern in transition in facies and fining-upward trend in sedimentary packages/stratigraphic units is considered to reflect a retrograding estuarine succession and subtidal to intertidal to supratidal sandy to muddy and carbonate tidal flat system which fringe the shoreline. According to tidal signatures found all over the distribution area of Rēzekne and Pärnu succession, and the regional geological trend, it is suggested that the deposition took place in macrotidal, tide-dominated estuarine system and on adjacent tidal flats (see more in Discussion, the Chapter 6.5). It is worth to note, that palaeocurrent directions may change from ebb to flood dominance over distances as short as a few hundred metres and the complexity of palaeocurrents is consistent with highly variable local net sand transport directions observed in modern tidally influenced coastlines (Ashley, 1990), which is very typical for deposits of Pärnu RS. The fact, that in modern environments the direction of bedform migration in response to tidal currents may range from ebb to flood and inclusive approximately shore-parallel migration (Ashley, 1990) is taken into account in palaeocoastline definition and palaeoenvironment reconstruction. In view of the orientation of the tidal systems in Rēzekne and Pärnu basins, the well developed SW-SE palaeocurrent distribution of dunes and sandbar bedforms suggests NW-NE tending shoreline. Shallow water (up to supratidal zone) developed at the basin margins and gradually deeper water (subtidal zone) towards the south of the study area. Strong river influx from the north reduced the salinity in the basin and carried in large amounts of sand-rich material, therefore the northern part of the basin is predominated by sandy deposits. Freshwater influx did not reach the southern parts of the basin, therefore dolomite and gypsum cement often occur there in the similar sandstones (Kurshs, 1975; Stinkulis, 1998). Sandy dolomitic marls spread in eastern and south-eastern part of the basin. Carbonate sedimentation prevails in this part of the basin, containing variety of facies, which are not found in other parts of the study area. The dominant type of sediments there are dolomitic marls and sandy dolomitic marls with inclusions of gypsum, which accumulated in supratidal and intertidal mudflats. Typical feature of these sediments are dolomite ooids and peloids. Internal fabric of ooids, including concentric laminae of sulphide minerals, suggest that these grains formed under changeable low and high agitation (Stinkulis, 1998), caused by tides. Further to the east in the territory of Belarus (Drissa-1ST and Liozno drill-cores) deposits contain still less clastic admixture, and are represented by the clayey dolomites. The gypsum lenses and intercalations often occur in the clayey carbonate sediments here. Eastwards from the study area gypsum layer appears and thicken to southeast, but farther to east even beds of halite occur (Tikhomirov, 1995). The amount and thickness of siliciclastic deposits increase again in the central part of the basin. Extensive non-carbonate tidal flats with tidal rhythmites that developed mainly in western and central parts of the basin indicate that the environment of deposition was typically intertidal. The amount of desiccation cracks suggests deposition even in supratidal settings. These environments typically associate with intertidal flats that border marine or estuarine water bodies. This is in contrast to what is suggested by Kurshs (1975), that deeper basin existed in the western part of Pärnu time basin in Baltic States.

The source area for this zone was situated in the north, and the material was transorted by fluvial and tidal currents, very occasionally by wave activity. As indicated by fluvial and tidal channel deposits in SE part of the basin, it is not excluded, that the clastic material was provided also from the southerly situated source area in Rēzekne time (base of Šķaune-103, Drissa-1ST un Liozno cores). During time of transgression and lowered sediment influx from the source area the territory was flooded and estuaries and tidal flats formed with predominant mudflat formation. Wave action had a subordinate role in sediment transport and deposition in Rēzekne and Pärnu basins. Few indicators of wave influence to the deposition of sediments in the basin exist, such as occasional wave ripple lamination and wavy, irregular lamination, however no deeper water sedimentary structures, indicating active wave action, such as swaley and humocky cross-stratification are found. This again reconfirms, that the sediments throughout the basin were deposited in very shallow-water environments. Although wave-dominated coasts and wave-built sedimentary bodies are very common on the seaward flanks of tide-dominated estuaries (Yang et al., 2005, 2007; Daidu, 2012; Tessier, 2012), the lack of wave generated structures in Rēzekne and Pärnu basins confirms that tidal processes controlled the deposition throughout the basin. Occurrence of subaerial exposure is indicated also by the wide variety of palaeosols. It must be noted, that the palaeosols occur in different intervals of Rēzekne and Pärnu succession, but the trend of their association with the topmost parts of beds is very typical, usually they are associated with: 1) fluvial channel deposits; 2) cover the top of major tidal bedforms, such as bars; 3) are intercalated with supratidal to intertidal mudflat deposits. These palaeosol intervals mark the last phase in the channel infill, and can be used as indicative for episodes of exposure only if occur in same intervals in few adjacent drill-cores, such as for example in Dobele-4 and Jelgava-2 drill-cores, the succession of which is interpreted as intertidal to supratidal mudflats.

The volume of the depositional units of Rēzekne and Pärnu RS differs from one part of the basin to another. The thickest units of Rēzekne and Pärnu RS are found in the northern part of the distribution area, while the thinnest occur in the south. One of the parameters determining estuary volume is the coastal plain gradient (Boyd et al., 1992). According to Steel et al. (2012) the thinnest of most of these transgressive intervals probably reflects a lack of significant relief articulation during rapid transgressive drowning or a low sediment supply, whereas thicker transgressive accumulations reflect a higher sediment supply and a steeper transgressive trajectory. A rise of sea level on a flat coastal plain passive margin generates a long estuary with a large volume. According to Steel et al. (2012), the transgressive tidal accumulations in incised valleys commonly reach significant thicknesses. Thus, presumably, the thickest beds of the succession accumulated in estuaries, while the thinnest on tidal flats. Moreover, estuaries are highly efficient sediment traps and their deposits have high preservation potential because of their location within palaeovalleys (Dalrymple et al., 1992). Several estuaries have been interpreted to form in Rēzekne and Pärnu basins, situated along the palaeocoastline. As indicated in correlation chart of the adjacent drill-cores (see Appendixes) the development of estuaries during the Rēzekne and Pärnu time was preceded by incision of the channels in the Silurian beds about 15 m, and this pattern is seen throughout the cross section.

Depositional rates, preservation potentials and stratigraphy

In tidal systems, especially such sand-rich ones as sequences of Rēzekne and Pärnu RS, preservation potential of deposits and time gaps in the stratigraphic records must be taken into account. The preservation potential of tidal deposits is high where tidal flats are aggrading and prograding (Davis, 2012). This study demonstrates that tidal deposits of the Rēzekne and Pärnu RS accumulated rapidly and were preserved widely. Moreover, where preserved in the stratigraphic record, tidalites represent accumulation during very short time intervals (Klein, 1998). As a result the sedimentary record is partial, since accretion is usually discontinuous, and erosion often removes part or all of the record. This limits considerably the range of conditions favorable for preservation of deposits to areas where there is a good supply of sediment and most often a rising relative sea level in the subtidal zone (Nio, Yang, 1991) in areas sheltered from waves and storms (Dalrymple, 1992), such as the Rēzekne and Pärnu basins. Estuaries are defined as representing exclusively transgressive depositional systems, as geomorphologic feature that can be formed by the drowning of incised-valley system and which "become filled and cease to exist as soon as sea-level rise slows" (Dalrymple et al., 1992). Estuaries and their deposits are thus diagnostic of the sequence-stratigraphic late lowstand to transgressive system tracts (Meyer, 1998).

In stratigraphic context, three to four major sedimentary cycles were distinguished in the sequence of the Rēzekne and Pärnu RS in the Baltic States (see Figure 5.17. and the Appendixes). The disconformity at the base of the Rēzekne or Pärnu RS formed as a result of a relative sea-level fall and fluvial incision during the lowstand. Thus, this erosional surface interpreted as a sequence boundary, mark the base of Rēzekne and Pärnu RS. The tidal ravinement surface separates lowstand (LST) deposits from the above tidal bar deposits, representing retrogradational transgressive systems tract (TST). A lags of shell debris appear to be common feature of modern tidal channel and estuary deposits (Allen, 1991) and is usually associated with the tidal ravinement surface. However the absence of shell debris within tidal channels as suggested by Richards (1994) may reflect the removal of shell debris by diagenetic processes. Tidal ravinement surface may be produced by a number of many different types of tidal channel and does not necessarily separate distinctive facies in central and western parts of the basin, thus it is difficult to distinguish them. Moreover, these surfaces are rarely found outside the valley system (Reading, 1996), so it is very difficult to use ones for regional correlation. On the contrary, since they have a high chance of preservation in the valleys, they are detected in the outcrop area. Towards the end of deposition of Pärnu time the rate of relative sea level rise exceeded that of accommodation space filling, resulting in shallow marine conditions during deposition corresponding to the Narva time.

In overall sequence stratigraphic context, Rēzekne and Pärnu RS represent the deposits of the lowstand and the subsequent transgression periods. The large thickness of the tidal deposits suggests a state of near-equilibrium between sedimentation rate and relative sea-level rise. This relative sea level rise was due to combination of progressive basin subsidence and a eustatic sea-level rise in Rēzekne and Pärnu time (Marshall et al., 2007; Haq, Schutter 2008). Tidal deposition was terminated when the rate of relative sea-level rise exceeded that of filling of accommodation space in the Narva time.

Tide-dominated estuaries and tidal flats

The tide-dominated channel fill and bar deposits and the associated marginal tidal flat deposits of Rēzekne and Pärnu basins are interpreted to have been deposited in estuaries. This suggestion is made based on the scale of incision at the base of channels, on the infill architecture and on the presence of transgressive tidal deposits. The lack of barrier islands and lagoonal sediments suggest deposition in wide-mouthed, tide dominated estuaries (Mellere, Steel, 1996). Facies trends within estuarine fills vary depending on the amount and distribution of sediment deposited by river, wave- and tidal currents, a balance that can vary over time as ratios of accommodation to sediment supply change during estuary filling (Allen, Posamentier, 1993). Estuarine deposits are in general aligned oblique to perpendicular to the shoreline particularly if they tend to be tidal (Reinson, 1992). The estuaries of Rēzekne and Pärnu RS, that fringe the shoreline and are aligned perpendicular to palaeocoastline, are interpreted to be of macro-tidal origin, to development of which the major role was played by tidal processes. The possible tidal ranges in Rēzekne and Pärnu basins are discussed in Chapter 6.5.

Transport pattern in estuaries develops a bedload convergence zone within the middle portion of all estuaries, called "straight-meandering-straight", which generate grain-size trends in the sand fraction: 1) a seaward decrease in sand grain size through the river-dominated, inner part of estuaries; 2) landward decrease in sand size in the outer part of estuaries (Dalrymple et al., 1992; Dalrymple et al., 2004). As stated by Dalrymple et al. (1992) longterm transport of bedload is seaward in the river-dominated zone, whereas coarse sediment moves up estuary in the marine-dominated zone as a result of wave and/or flood-tidal current action. Thus, the central zone is an area of net convergence and typically contains the finest-grained bedload sediments regardless of whether the estuary is wave-or tide-dominated. Richards (1994) discusses that mesotidal estuaries develop well-preserved tripartite sand-mud-sand structure. A macrotidal, coarse-grained, sand-rich and complete estuarine fill of tide dominated estuarine systems of Rēzekne and Pärnu RS, especially in its northern distribution area, lack the well developed tripartite sand-mud-sand facies. Only in relatively few cores the fine-grained central estuary deposits are present. For more discussions see Chapter 6.1 on tide-dominated sand-rich estuaries. As a result, the depositional record of estuaries of Rēzekne and Pärnu basin is dominated by coarse-grained subtidal sand bars and estuarine channels that pass landward into sand tide-influenced fluvial to tidal channel and to fluvial channels in palaeocoastline zone, which corresponds to outcrop area. This central part is dominated by sands that occur in the tidal channels along the length of the estuary, while muddy sediments accumulate primarily in marginal tidal flats along the sides of the estuary. Elongated sand ridges of outer estuaries basinwards are interpreted from drill-core material in northern Latvia and southern Estonia. They contain minor amounts of mud, such as thin mud laminae and flasers, which accumulated in estuary inlets, possibly only in sheltered coastal embayments (Stupples, 2002). The ridges are erosively-based and composed of well-sorted, medium to coarse-grained sands with large-scale cross-stratification. Erosional surfaces with mud rip-up clasts are common. The upper portion of this transgressive estuarine succession is generally removed by tidal channel erosion. The presence of these ridges in addition confirms the tidal dominance to the deposition, while in wave-dominated estuaries barriers at the mouth of estuaries usually develop (Tessier, 2012).

It is more difficult to identify muddy intertidal sequences in the rock record than their sandier counterparts due to the lack of channel fill sequences and the very gradual gradients in sedimentological properties (Hinton, 1998). The close association of muddy sequences in succession of the Rēzekne and Pärnu RS with the channel fill tide-influenced and tidal estuarine sequences, allows to interpret muddy sequences of Rēzekne and Pärnu basin as deposited on tidal flats. Also fining-upward cycles is a supporting evidence for deposition on prograding tidal flats (Middleton, 1991). Such fining-upward sequence, common for tidal flats, which are building seaward, may also be identified as due to the lateral migration of fluvial systems (Hinton, 1998). However all range of tidal signatures present in deposits of Rēzekne and Pärnu RS, allows to interpret these fining-upward sequences are produced by migration of tidal channels, and by development of supratidal flats over intertidal and subtidal sandy flat and channel deposits.

The density of bioturbation in a bed varies inversely with the rate of sedimentation so that, where sediment supply is low and invertebrate life abundant, primary structures such as bedding may be completely destroyed (Miall, 2000). Beds formed by rapid sedimentation may show little or no penetration by organisms, as in the northern part of the Rēzekne and Pärnu RS distribution area. The most bioturbated and thus preserved layers are found within supratidal to intertidal mudflat settings. For intertidal flats, the sediment is delivered by tidal currents from the adjacent subtidal environments. Such intertidal deposits are still vulnerable to erosion by wind, tides and storms (Davis, 2012). Therefore most of the sequence shows scours and local unconformities. According to Middleton (1991) special hydraulic conditions are necessary for the preservation of clear evidence for tides: 1) subtidal environment, probably in a laterally migrating tidal channel with relatively weak, strongly asymmetrical tidal currents; 2) the predominant tidal current is able to transport sand and the subordinate current very little, so that there is little erosions of the cross-strata formed by the predominant tidal phase; 3) an environment rich in suspended mud is necessary so that each tidal sand layer can be clearly separated from the next. When such conditions are present, as in the sequence of Rēzekne and Pärnu RS, the cyclic nature of the deposits constitutes very convincing evidence of tidal origin.

6. Discussion

6.1. Recognition criteria of sand-rich tide-dominated estuaries

Some criteria for recognition of sand-rich tide-dominated estuaries in the rock record are proposed here. The distinction of sand-rich tide-dominated estuaries with respect to facies and stratigraphy of sediment infill is discussed here.

It must be pointed out that the examples of tide-dominated estuaries described from the rock record remain relatively rare (Archer et al., 1994; Mellere, 1994; Khin, Myitta, 1999; Rossetti, 1998; Shanmungam et al., 2000; Plink-Björklund, 2008). Despite the fact that many sedimentological and sequence stratigraphical analysis describe facies and deposition in estuarine environments, most interpretations refer to the wave-dominated estuary model or to mixed-energy wave and tide-dominated estuary model based on the Gironde estuary (Tessier, 2012). Few examples of ancient sediment successions have been interpreted as exclusively tide-dominated estuaries (Archer et al., 1994; Mellere, 1994; Gjelberg, Steel, 1995; Shanmugam et al., 2000; Hori et al., 2001; Plink-Björklund, 2005; 2008; Kitazawa, 2007; Pontén, Plink-Björklund, 2009). Majority of this research deal with the studies of fine-grained tide-dominated estuaries, while little material exist on sand-rich estuarine systems. Sand-filled estuaries appear to be difficult to recognize, especially in the rock record. Without a thorough knowledge of the palaeogeography and without sufficient experience in recognizing the influence of tides in siliciclastic sediments, one might interpret an estuarine sandy channel-fill as a fluvial channel-fill or a distributary channel of a fluvial-dominated delta. This is one of the reasons why the sequence of Pärnu Fm in the Baltic States has been previously interpreted as fluvial channel or distributary fills and delta deposits (Kurshs, 1975, 1992; Kleesment, 1997). It is necessary to note that the upper part of Pärnu RS is one of the sandiest parts in the Devonian sequence - the proportion of sandstones reaches 70-90 % (Kurshs, 1975).

The following discussion will address how sand-rich tide-dominated estuaries can be recognized in stratigraphic succession and how they can be distinguished from similar environments/features that are deposited under non-tidal conditions, as well as from their fine-grained counterparts. Some of the sedimentary structures and sequences can form in multiple depositional environments, such as fluvial, deltaic or shallow marine. At the same time, in some cases sediments also accumulate under tidal conditions, but there are no tidal signatures preserved, such as in estuaries, tidal flats, or tidal inlets. The lack of tidal signatures may be due to energy conditions with little or no sediment transport, or may be due to the influence of waves that rework any evidence of tidal activity (Davis, 2012). Because of these circumstances, it is certain that beds which are interpreted as containing tidal sediments, may represent only a portion of those sediments that has been actually accumulated under tidal conditions.

Davis (2012) points out that monolithic tidalites occur, but are not common, in the geological rock record. He concludes that it is likely that tidal deposits in monolithic depositional environments do not leave tidal signatures, and are therefore not recognized in ancient rock record. He notes that, however, few exceptions exist. One of such exceptions is Precambrian Baraboo and the Cambrian Jordan sandstones in Wisconsin (Pape et al., 2003), were tidalites are entirely composed of well-sorted sandstone.

In this study it is illustrated that tidalites of Pärnu Fm in the northern part of its distribution area are mainly monolithic facies (see more Chapter 5.1.1) and argued that even in such environments, where monolithic sedimentation prevails, tidal signatures are abundant and can be distinguished in the rock record.

A number of indicators reflecting tidal control on the deposition are preserved in modern and ancient stratigraphic sequences. These tidal signatures are distinguished by cyclic deposition and are important indicators in reconstructing the ancient depositional environments. In sand-rich tidal environments, such as estuaries, the bulk of tidal signatures are preserved within cross-strata formed by migrating bedforms where currents move over non-cohesive sediments of sand or gravel (Davis, 2012). Therefore special types of cross-strata containing specific tidal signatures are produced, such as tidal bundles, reactivation surfaces, grain size upwards-decrease within individual cross strata in the cross-sets, reversals in palaeocurrent directions in bedforms of different scales (see more for details in Chapter 5.1.4).

The currents during flood and ebb tide move and deposit relatively coarse sediments, typically sand. During slack water periods, fine suspended sediment settles. The result is an alternation of sand and mud forming tidal bedding or rhythmite (Reineck, 1967), that is usually heterolithic, displaying more than one sediment type. Tidal rhythmites are typically associated with intertidal flats that border marine or estuarine water bodies (Davis, 2012). In case of Pärnu Fm, in northernmost part of its distribution area, few heterolithic tidal sediments are preserved, but are found mainly in drill cores further south. The whole succession is almost entirely sand-dominated. The rhythmites of Pärnu Fm in outcrop area, represented by marginal tidal flat deposits, are monolithic and contain mainly fine to very-fine grained sand. It must also be noted, that rhythmic sediments can accumulate in fluvial and lacustrine environment and especially in fluvial deltaic environments (Harris et al., 1993; Dalrymple et al., 2003). Therefore in order to clearly interpret the sedimentary environment in which tidal signatures occur, the placement of the sequence in the context of entire depositional system is essential. Since deposits of marginal tidal flats (FA 4) of Pärnu Fm are closely associated with tide-dominated deposits of outer estuary (FA 3), there is no doubt about its origin in tidally-controlled environment.

The same condition is applicable to another very typical tidal signature in facies of Pärnu Fm - tidal bundles, when each bundle is a couplet of typically heterolithic cross-strata that develop from the migration of 3-D dunes. However, again in facies of Pärnu Fm in northernmost part of its distribution area, tidal bundles appear to be monolithic and contain almost entirely sandy sediments. The tidal bundles accumulate in a sequence of trough cross-strata that shows rhythmic changes in individual bed thicknesses from spring to neap during each tidal cycle (Visser, 1980). These bundles are excellent examples of tidal sediments and show high preservation potential, because they develop in channels where burial can be rapid (Davis, 2012). In such bedforms there is also a high potential of reactivation surfaces to be present within the bundle sequence, especially during spring tides, which are very prominent in tidal bundle sequence of sandy deposits of Pärnu Fm. In the absence of fine-grained material (silt- and mudstone), which usually forms tidal bundles and associated mud/mica drapes, tidal bundle sequences in sandstones of Pärnu Fm in outcrop area very often exhibits grain size decrease from coarse and medium-grained sand to fine and very fine-grained sand and silt at individual cross-strata scale within tidal bundle sequence (for more details see Chapter 5.1.4). Thus, these can be considered the coarse-grained counterparts of the heterolithic tidal rithmites that prevail in fine-grained systems.

Few key points that characterize the stratigraphy of sediment fill of sand-rich tide-dominated estuaries are highlighted here. For more on sequence stratigraphy of the succession see in Chapter 6.3.

Many studies related to tide-dominated estuary sediment infill refer to sequence stratigraphy concepts allowing the distinction of system tracts (lowstand, transgressive and highstand) and key surfaces, such as sequence boundary, the transgressive surface, the tidal and wave ravinement surfaces and maximum flooding surface (Tessier, 2012). Due to diversity of interpretations there is no clear definition of the statigraphy for sand-rich tide-dominated estuaries. Moreover, in such sand-rich system, as the northern part of Pärnu basin, where several episodes of incision and infill took place, it appears to be difficult to differentiate the systems tracts and correctly locate major stratigraphic surfaces. As discussed by Tessier (2012), it is worth noting, that both in modern settings and in the rock record, sediment infill in tide-dominated estuaries show a large diversity in terms of geometry and relative proportion of facies within the preserved systems tracts.

The sequence boundary in most cases is defined at the bottom of a fluvial valley incised during the previous relative sea level drop. The lowstand systems tract (LST) is usually assigned to fluvial deposits preserved in the bottom of the valley. In most cases, it is very reduced in volume, represented only by remnants of fluvial deposits, reworked during the subsequent transgression by powerful tide/wave currents (Tessier, 2012). In only few cases part of the LST consists of tide-influenced fluvial deposits (Gjelber, Steel, 1995; Plink-Björklund, 2005). The basal erosion surface of Pärnu Fm erodes into older, mainly Silurian rocks in the northern part of its distribution area according to drill-core data in the adjacent cores. However, the full lateral extent of the erosion surfaces is unknown. The continuous fluvial and tide-influenced fluvial to tidal channel deposits on the base of the erosional surfaces occur across the entire cross-section and northern part of the distribution of Pärnu Fm (SU 2, see Chapter 5.1.2). These deposits grade basinwards and upwards into tide–dominated deposits . Fine-grained deposits are missing in this part of the section in the study area. At the base of SU 2, single fluvial or tidally influenced fluvial to tidal channels, as well as multiple vertically stacked fluvial or tidally influenced fluvial to tidal channels occur. This prominent and widespread erosional surface that is overlain by coarse-grained fluvial channel deposits and mark the base of the Pärnu Fm, is interpreted as sequence boundary, because: 1) these erosion surfaces are traceable in in all drill cores; 2) the depth of incision is significant (local erosional topography up to 15 m) and 3) these erosion surfaces are overlain by fluvial deposits that mark a significant basinward facies shift in comparison to mainly carbonate shallow marine sediments of the Silurian age. Sequence boundaries formed during the fall of relative sea-level to its lowest position. Similar to tide-dominated estuarine deposits of the Eocene Central Basin, Spitsbergen (Plink-Björklund, 2005), the fluvial and tidally influenced fluvial channels reflect deposition in rivers that became tidally influenced at their river mouth, below the mean high tide. Therefore the successions of coarse-grained fluvial and tidally influenced fluvial deposits that shift abruptly basinward across older sequences are interpreted as lowstand deposits (LST). The upwards transition from fluvial to tidally influenced fluvial deposits indicates deposition during the initial sea level rise while the tripartite composition of the three stratigraphic units indicates fluctuations of the sea level and three episodes of incision and infill.

The transgressive systems tract (TST) is usually described as a bulk of tide-dominated estuary infill (Zaitlin et al., 1994). The fluvial channels are generally overlain by tide-dominated estuarine deposits. At the top of the fluvial channel deposits there is a marked transgressive surface that signifies valley drowning and the landward migration of the bayline, separating the fluvial lowstand deposits from the estuarine transgressive deposits. It is marked by a significant erosional surface. The transgressive surface separates fluvial aggradation from landward-stepping estuarine deposition in basinward portions of the system.

These tide-dominated estuarine deposits have a landward-stepping trend as they onlap the basal fluvial deposits in a landward direction. At the landward end of the exposed estuaries (outcrop area), the landward-stepping is implied by a landward shift of the tidally influenced fluvial channels. Tidally influenced fluvial channels are replaced by tidal channel sand bars at gradually higher stratigraphic levels. At the seaward end of the exposed estuaries (outer segment), the landward stepping is implied by the landward shift of tidal sand bars and tidal flat deposits. The outer segment of estuaries is usually very sand-dominated and is composed mainly of tidal sand ridges outside the studied outcrop area.

These tidal sand ridges deposits occur on highly erosional basal tidal ravinement surfaces that, in some cases, cut through the whole underlying estuarine and fluvial deposits and merge with the sequence bounding erosional surface (similar to the deposits found in drill-cores in Central Latvia). In most sequences the tidal ravinement surfaces occur within the landward stepping estuarine deposits, some distance above the basal fluvial segment. Similar to tide-dominated estuarine deposits of the Eocene Central Basin, Spitsbergen (Plink-Björklund, 2005), the transgressive deposits volumetrically dominate the depositional sequence. However, unlike the study of Plink-Björklund (2005), where in contrast to the lower, fluvial segment, the middle transgressive segments are mud-prone, the TST deposits of Pärnu Fm are sandy and contain only mud/mica drapes within cross-strata of the bedforms, such as sand dunes and bars. The only mudstones, also containing carbonates, were deposited on marginal tidal flats. These fining-upwards depositional units occur at the very top of the succession and are associated with the bars of the outer estuaries. It should be noted that no signs of bioturbation are found in the sand rich estuarine environment of the Pärnu Fm due to highly stressed environment (Dalrymple, Choi, 2007). However tidal flat deposits in seaward portion of the estuaries (outer segment) tend to be more bioturbated. The landward-stepping estaurine succession that accumulated above the lowstand deposits is interpreted as transgressive estuarine deposits. A relative sea-level rise is indicated by a landward shift of all depositional environments, a landward onlap, and the development of the tidal ravinemet surfaces. The highstand systems tract deposits correspond to Narva RS of the Middle Devonian succession in Baltics, and are not described in this study, but studied in detail by Tänavsuu-Milkeviciene et al. (2009).

In such sand-rich environments as Pärnu basin, where rapid deposition prevails, it is very important to consider the preservation potential of the deposits (see also Chapter 5.2.2). What is preserved in the stratigraphic rock record as tidalite represents the minimum of tidal environments that existed in the geological past. One of the most common environments, where this situation prevails, is the transition from fluvial to tidal domination in estuaries (Dalrymple, 1992; Davis, 2012), which are widely represented in the northernmost part of the Pärnu RS distribution area, including the outcrop area. According to Davis (2012) tidal channels tend to be among the best preserved, whereas the upper intertidal zone is the most poorly preserved. It must be noted, that in this study, tidal channels are very well preserved in the rock record, and can be studied in outcrops and in cores, however tidal deposits of the intertidal zone occur in other parts of Pärnu basin with the exception of most northern part of its distribution area, closer to the palaeocoastline where deposition in sand-rich estuaries prevails.

6.2. Tide-dominated estuaries versus tide-dominated deltas

The deposits of tide-dominated estuaries and tide-dominated deltas can be remarkably similar in the geological rock record. Coarse-grained estuaries are especially difficult to distinguish from fluvial channel-fills or distributary channels of tide-dominated deltas. As demonstrated by long-lasting debates about the interpretation of the depositional environment, accurate facies models are still missing to allow the distinction between different tide-dominated environments as preserved in the rock record (Tessier, 2012). A very careful facies analysis combined with detailed studies of vertical and lateral facies transitions are essential in order to determine tide-dominated estuarine deposits from tide-dominated delta deposits (see discussions in Walker, 1992; Dalrymple et al., 1992; Boyd et al., 2006; Dalrymple, 2006; Dalrymple, Choi, 2007). By developing clear facies models based on distinctive combination of sedimentary processes, it is possible to correctly identify and differentiate these two environments (Boyd et al., 2006).

Pärnu RS in its northernmost distribution area (outcrop area) has been previously interpreted as fluvial channel or distributary fills and delta deposits (Kurshs, 1975, 1992; Kleesment, 1997). In this study it has been re-interpreted as tide-dominated estuary. The below-mentioned criteria allow interpreting the deposits of Pärnu Fm in the outcrop area as a tide-dominated estuary and distinguish it from tide-dominated delta succession. These criteria can be very well compared with the discussions on depositional environments of other sequences of the BDB, especially with regards to deposition of tide-dominated deltas (Pontén, Plink-Björklund, 2005; Tänavsuu-Milkeviciene, Plink-Björklund, 2009; Tänavsuu-Milkeviciene et al., 2009).

- The deposits of Pärnu estuary in the study area and in adjacent drill cores show evident overall upward-fining grain size trend repeating in three cycles (represented by three stratigraphic units). Based with unconformity on underlying Silurian marine carbonate deposits within stratigraphic units, coarse-grained deposits of fluvial channels and bars are overlain by finer grained, well-sorted sandstones with mud drapes, which are capped by fine to very fine-grained, dolomitic sandstones at the very top of the succession. The deposits of Pärnu succession are caped by overlying shallow marine mixed siliciclastic-carbonate deposits of Narva Fm (Tänavsuu-Milkeviciene, Plink-Björklund, 2009). On the contrary, deposits of tide-dominated deltas show overall upward-coarsening trend, reflected by delta progradation basinwards, typical for other parts of the Middle to lowermost Upper Devonian siliciclastic succession in the Baltic States: upper part of the Narva Fm, Aruküla Fm and Gauja Fm (Pontén, Plink-Björklund, 2007; Tanavsuu-Milkieviciene, Plink-Björklund, 2009; Tanavsuu-Milkieviciene et al., 2009).
- The deposits of Pärnu Fm in the study area and in adjacent drill cores show coarser grain size trend at both ends of the estuarine system, thus reflecting the two sediment sources: fluvial and tidal. Coarser deposits are typical for the northern (landward) part of the distribution area, while most fine-grained deposits, including significant mud layers, occur in the central part of the system. Another coarse-grained size deposit peak corresponds to more basinwards oriented reaches of the estuary, based on data derived from cores in southern Estonia and northern and NE Latvia. In contrast to estuarine fills, for tide-dominated deltas overall basinward fining grain-size trend is typical, which is driven by the fluvial sediment source (Boyd et al., 2006; Dalrymple, Choi, 2007; Tänavsuu-Milkeviciene, Plink-Björklund, 2009), since tide-dominated deltas are defined as progradational sediment bodies that protrude from the shoreline, where tidal currents rework river supplied sediments basinward (Willis, 2001).
- The above-mentioned criteria on describing two sediment sources in the Pärnu estuary, are reflected also by the major sediment transport directions. Deposits of Pärnu estuary show bidirectional palaeocurrent distribution, especially at its central part and also outer (basinward) reaches. On the contrary, tide-dominated deltas will

in general have more basinward oriented palaeocurrent distribution (Boyd et al., 2006; Tänavsuu-Milkeviciene, Plink-Björklund, 2009).

- Sharp base and "blocky" character of the tidal bars. According to Steel et al. (2012) tidal signals in regressive and transgressive succession are not greatly different, though tidal bars and sand ridges tend to be more thickly developed in estuaries, and some facies successions (e.g., tidal bars fining upwards into tidal flat and supratidal muds) are more common in estuaries, whereas in deltas tidal bars tend to cap upward-thickening parasequences. According to Dalrymple et al. (1992), dune and bar deposits may form a significant part of the geological record of estuaries and are extensively developed in estuaries as a result of broad range of conditions under which they form. Rahmani (1988) states, that lateral accretion bedding is a characteristic architectural element of deposition of point bars. Estuarine bars compared to their fluvial counterparts tend to be asymmetric in the direction of the dominant current. They have lower-angle down-current bed dip angles, accrete mostly lateral to the direction of the dominant flows, and tend to fine upwards from the bar fronts to overlying bar top deposits (Dalrymple, Rhodes, 1995). This is the most common stacking pattern feature for the bar units in the outcrop study area (FAs 2 and 3). Each of the cross-stratified sets that forms the point bars, consists in detail of a cyclic bundling of sandier and muddier laminae suggesting a regular and orderly fluctuation in energy conditions common to tidal processes. According to Dalrymple and Rhodes (1995), as well as Yang and Nio (1985), preservation of bundle sequence attests to high depositional rates. For example, this is the case for lateral accretion on the inner bank of a curved tidal channel. The latter, together with above-mentioned criteria, reflects deposition in tide-dominated bars of estuaries.
- With regards to sequence stratigraphic analysis, the Pärnu succession in the study area is interpreted as an incised valley estuary. Tidal estuary deposits are associated with the fills of valleys flooded during periods of relative sea level rise (Dalrymple et al., 1994), and erosional surface, formed during valley incision is characteristic. The erosional unconformity, formed during valley incision is characteristic for Pärnu Fm and is traced regionally in the drill cores. In the most northern distribution area of the Formation, the erosional surface is overlain by poorly sorted fluvial channel deposits suggesting that sediments have been formed due to a valley cut into underlying Silurian carbonate rocks. The tripartite character of the succession suggests several episodes of incision and infill. On the contrary, deltas are usually not associated with incisions (Plink-Björklund, 2005; Boyd et al., 2006). The more detailed description on valley incision is given in Chapter 6.3, discussing the regional trend of the depositional environment of the whole Rēzekne and Pärnu basins.
- In sequence-stratigraphic terms, estuaries more commonly occupy the transgressive system tract while deltas more commonly occupy the regressive and highstand system tracts (Boyd et al., 2006). Tidal bars of Rēzekne and Pärnu RS in its more basinward reaches are based by tidal ravinement surface, and in the most landward position (the studied outcrop area), are underlain by estuarine and fluvial facies. The estuary formed during flooding (early transgressive stage) of a previously created incised valley with a pronounced unconformity at its base.

6.3. Basin topography and incised-valley estuaries

The Rēzekne and Pärnu RS of the BDB represents lowstand to transgressive succession of tide-dominated esturaine deposits associated with adjacent tidal flat complex. Few definitions of estuaries exist (Dalrymple et al., 1992). The most common classification used in sedimentology, which is also used in this study is the one defined by Dalrymple et al. (1992) as "the seaward portion of a drowned incised valley". Incised valley has been defined (Zaitlin et al., 1994) as a "fluvially eroded, elongate topographic low that is typically larger than a single channel form, and is characterized by an abrupt seaward shift of depositional facies across a regionally mappable sequence boundary at its base. The fill typically begins to accumulate during the next base level rise and may contain deposits of the following highstand and subsequent sea-level cycles". Incised valleys pass landward into fluvial-channel systems that feed the valleys (Zaitlin et al., 1994).

Deposition in Rēzekne and Pärnu basins took place in an environment comprising a shallow epicontinental tidal sea (see Discussion in Chapter 6.5). The presence of tidal currents implies that Pärnu basin was connected with open ocean basin and that the tidal currents formed in the open oceans and propagated into the adjacent shelf seas. Today, most tide-dominated estuaries are associated with tide-dominated shelves, such as the English Channel and the China sea and more generally to shelves, that are large enough to amplify the oceanic tidal wave. Elongated bays and gulfs are then favorable coastal configurations for extreme amplification of tidal waves that propagate on shelves. Thus, coastal configuration is a critical factor in controlling tidal dynamics (Tessier, 2012). Several estuaries are interpreted to be developed along the palaeocoastline of the Rēzekne and Pärnu basins based on detailed facies analysis and spatial and lateral distribution of these facies derived from outcrop, as well as drill core data. At a first glance it may seem unlikely that major rivers entered the basin every few tens of km along the coastline. However, it should be noted that the land-ocean interface in BDB was likely different from most coastlines that exist today. In the absence of rooted land plants, point sources of sediment supply were unlikely. Instead, the land-ocean interface was most probably in the form of braided network with tidal modifications taking place in rivers and on tidal flats (Eriksson, Simpson, 2012). Moreover, the relatively flat topography of the BDB meant that large areas could have been influenced by water-level changes, both from the seaward and the landward end. Also, a global palaeo-ocean spanned most of the planet during Palaeozoic Era and strong tidal resonances could have occurred (Archer, 1996). Funnel-shaped estuaries within low-relief basins with strong basin resonances would also favour tidal conditions resulting in a variety of tidal facies (Archer, Greb, 2012) common for northern part of the basin. Also, the relatively big size of the landmass of Euramerica may have resulted in the common occurrence of big palaeorivers (Potter, 1978; Archer, Greb, 1995).

The estimation of the total depth of erosion (eventual incised valley depth), is not completely clear because of the relatively narrow ouctrop area, as well the lack of data in the drill cores, which often due to their coarse-grained composition are poorly preserved in cores. Nevertheless the character of the erosion surfaces at the base of the Regional Stage and its regional distribution up to 15 m, as well as the distribution and configuration of the estuaries along the palaeocoastline, suggest that the estuarine deposits backfilled incised valleys, that had been eroded during the preceding sea-level falls (see more details in Chapters 5.1.3 and 5.2.2).

The criteria for recognizing incised valley estuaries outlined by Zailtin et al. (1994) can be applicable for this study:

1) The basal erosion surfaces, that occur througout the study area and mark the lower boundary of Rēzekne and Pärnu RS, are regional incisions. This high relief erosion surface records deep incision into mainly marine carbonate deposits of Silurian age during a lowstand in relative sea-level. The deepness of the incision is another critical aspect to consider regarding bedrock inheritance. Since it controls directly the accommodation space, the incision deepness determines the stage of infill of an estuary with respect to sediment supply, that is its degree of maturity from unfilled to completely filled (Dalrymple et al., 1992). However, this is important for both tide- and wave-dominated estuaries. For tide-dominated estuaries, the incision depth governs the potential for preservation of the infill (Tessier, 2012);

2) The basal fluvial deposits of the LST exhibit a significant basinward facies shift;

3) The base of the incised valleys are in places correlated with fine-grained sediment horizons of the intertidal areas in the northenmost part of the distribution area (along the palaeocoastline);

4) The estuaries infills onlap landwards the valley walls and form more significant packages of tide-dominated deposits along the palaeocoastline. Moreover, tide-dominated estuaries require confinement in a narrow, funnel-shape geometry (see Dalrymple et al., 1992; Zaitlin et al., 1994), suggesting that the whole thickness of the individual depositional sequences may have been confined within the valleys. Another criteria may be that because deposits above the erosion surfaces record mainly ebb-dominated paleoflows, it suggests that sediments must have been supplied mostly from more landward areas, rather than originating as marine sands rework into estuary mouth (Dalrymple at al., 1990).

Definition of estuary type and the history of sea level fluctuation is dependant largely on the identification and correct interpretation of the bounding discontinuities (Reinson, 1992). The tripartite character of the valley incision typical for Rēzekne and Pärnu RS, reflects three major episodes of the valley incision and infill. The thickness of each depositional sequence/stratigraphic unit varies from 5 to 25 m. Moreover, the thickest sequences tend to be associated with sequence boundaries in the northernmost part of the distribution area, which is along the palaeocoastline. Tidal resonance or at least tidal amplification has a low potential to occur in valley that is inundated too rapidly, or if tidal amplification occurs, it does not last enough time for a tide-dominated estuary to develop. This can be a reason why majority of the sequences are largely eroded, such as for example SU 1 and SU 3 from the outcrop area. The infill of all types of estuaries depends also on balance between the rate of sea level change and sediment supply, which has been changing in time.

The lack of fluvial deposits in some of the drill-cores, especially those situated basinwards, or relatively thin packages of fluvial deposits overlying the erosion surface at the base of the sequences, may indicate that the sediments were afterwards eroded by strong tidal currents during maximum tidal resonance following the transgression. Tide-dominated estuaries are associated with powerful tidal currents and therefore potentially deep tidal scouring. As a consequence of a deep tidal ravinement surface, preservation potential of underlying LST deposits is low (Tessier, 2012). The valley shape determines the possibility for tides to be amplified as transgression takes place in the valley. Funnel-shaped valleys and more generally valleys with high length/width ratio, primarily favour hypersynchronous behavior of the tidal wave, and thus tide-dominated estuary occurrence. Moreover, narrow valleys are assumed to promote tide-dominated estuary development rather than wave-dominated estuaries whereas broad valleys might not provide enough construction to create strong tidal currents, causing them to be wave-dominated (Nordfjord et al., 2006). They are very similar to range of estuaries along the western coast of France (for details on analogues see Chapter 6.4).

6.4. Ancient and modern analogues of Rēzekne and Pärnu basins

Estuaries which occupy drowned valleys are common along modern transgressive coasts and were presumably equally abundant during past transgressions. They are highly efficient sediment traps and their deposits have high preservation potential because of the location within palaeovalleys (Dalrypmle et al., 1992).

Many studies on modern and ancient estuaries exist (Rahmani, 1988; Dalrymple et al., 1990; Allen, 1991; Allen, Posamentier, 1993; Dalrymple, Zaitlin, 1994; Richards, 1994; Kvale et al., 1995; Fenies, Tastet, 1998; Lanier, Tessier, 1998; Plink-Björklund, 2005, 2008; Tessier, 2012). However examples of tide-dominated estuaries in the rock record are not widely described. Moreover, the estuaries like the one of Pärnu Fm are characteristic by its compositionally coarse-grained material, are not as widely described as their more fine-grained analogues (Rahmani, 1988; Lanier et al., 1993; Richards, 1994; Plink-Björklund, 2005, 2008).

Holocene analogues for Pärnu estuaries are widely developed along east coast of Australia and west coast of France (Woodroffe et al., 1993; Garnaud et al., 2003; Chaumillon et al., 2008). Cross-bedded sandstones that comprise the major portion of Pärnu Fm have analogues in the form of subtidal sand bars in many Holocene settings, such as bay Mont-Saint-Michel, Bay of Fundy, Gironde estuary, estuaries of Netherlands (Figure 6.4.1.).

Gironde estuary in SW France described by Allen (1991) is one of the modern models for macrotidal estuarine systems. However, unlike Pärnu estuary, it is mixed-energy wave- and tide-dominated (Tessier, 2012). Nevertheless, analogues of facies and architecture of the beds of tidal channel deposits of Pärnu Fm are observed in estuarine tidal bars from Gironde estuary (Fenies, Tastet, 1998; Figure 6.1F). Here ebb-oriented 2D and 3D dunes superimpose the surface of the bar, where abundant clay pebbles, eroded from supratidal marsh, are incorporated in the cross-sets of the beds. These facies are similar to the coarse-grained channelized facies in the lower part of the SU 2 in the outcrop area (FA 1). Moreover, dunes, climbing on the surface of the bar are both flood- and ebb-oriented, which shows certain similarities with the superimposed forms (dunes and ripples) climbing up the large macroforms (FAs 2 and 3) of Pärnu Fm in outcrop area, also reflecting polymodal palaeocurrent distribution.

A modern counterpart of Pärnu RS cross-bedded facies is subtidal shoal in the Eastern Scheldt estuary, SW Netherlands, the succession of which is similar to the middle unit of Pärnu estuary (SU 2). It illustrates the tripartite succession with coarse, cross-bedded sandstones channel thalweg deposits in the lower part, tide-influenced fluvial to tidal channel deposits in the middle part and mainly fining-upward tidal bar deposits in the upper part. It is also similar to the Holocene tidal channel section described by Yang and Nio (1989) in the Eastern Scheldt. Yang and Nio (1989) describe such a sequence of the section in tide-influenced environment as a result of a lateral migration of tidal channels, which is similar to the interpretation of the upper part of Pärnu Fm succession in outcrop area (FAs 2 and 3). Moreover, the topmost part of the tidal channel section in the Eastern Scheldt consists mainly of fine-grained sandstones, siltstones and intercalated mud beds, similar to FA 4 for Pärnu Fm in outcrop belt.

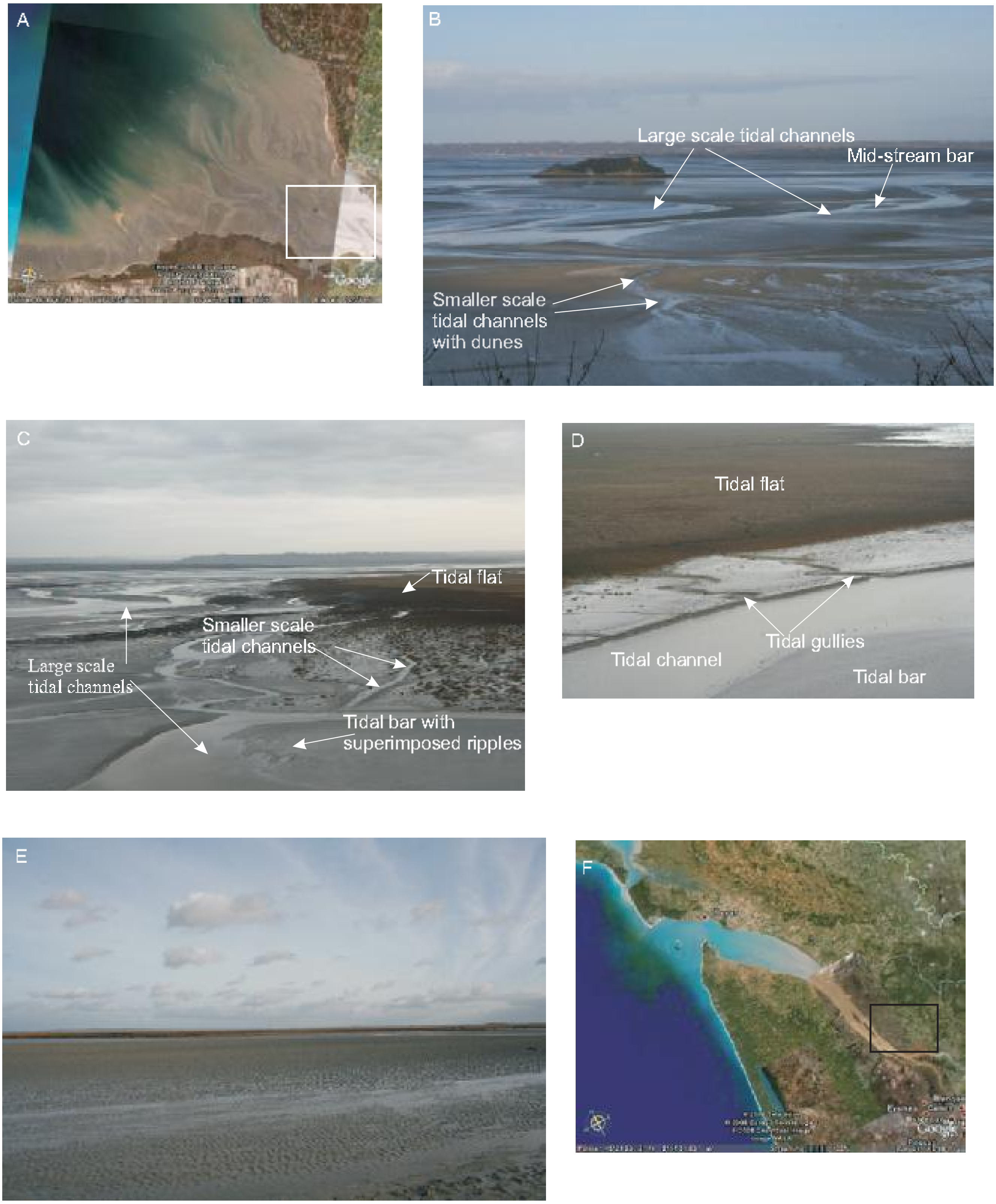

Figure 6.1. Several examples of modern tide-dominated estuarine depositional environments as analogues to estuaries and tidal flats of the Baltic Devonian Basin of Pärnu and Rēzekne time

A-E - Macrotidal estuary of Mont St Michel, NW France; B-D - Fluvial-tidal transition zone of inner estuarine depositional environment of Mont St Michel; E – Marginal tidal flats with current ripples exposed during the ebb, Mont St Nichel; F: Gironde estuary – another example of the macrotidal estuary, Western France

6.1. attēls. Daži mūsdienu Pērnavas un Rēzeknes laikposmu Baltijas devona baseina estuāru un plūdmaiņu līdzenumu vistuvākie analogi

A-E - Augstas plūdmaiņu amplitūdas estuārs - San Mišela, ZR Francija; B-D - Fluviālu-plūdmaiņu apstākļu pārejas zona Sant Mišelas estuāra iekšējā daļā; E – Malas zonas plūdmaiņu līdzenums ar straumju ripsnojumu bēguma laikā, San Mišela; F - Žironde estuārs – cits augstas plūdmaņu amplitūdas estuāra piemērs, Francijas rietumi

Rahmani, 1988, describes a 15 m succession in the Eastern Scheldt estuary, in which cross-bedded sands, 6 m thick, represent the lower part of the channel fill, while the upper part, about 9 m thick, consists of abandonment stage deposits with plane-parallel lamination. The cross-bedded sands show bimodal-bipolar palaeoflow, with the ebb being the dominant current. Sandstone contains well preserved mud couplets and tidal bundling, very similar to the sequences of Pärnu Fm. Data from a number of outcrops of recent and Tertiary deposits of the Rhine and Meuse Rivers in the Netherlands have been described by Van der Berg et al. (2007) with a focus on diagnostic criteria for fluvial-tidal transition zone. Few similar criteria have been described and compared also with Pärnu Fm deposits in outcrop area (for details see Chapter 5.1.4).

Cobequid Bay-Salmon River Estuary (Bay of Fundy) described by Dalrymple et al. (1990) is another modern analogue of Pärnu RS estuaries. Dalrymple and Rhodes (1995) describe similar to tide-dominated deposits of Pärnu RS gently-dipping, lateral accretion bars, formed by superimposed dunes, which migrate obliquely to the channels in the tidal environment in the Bay of Fundy. The authors indicate, that the deposits fine upward due to decrease in the current speed from the channels to the bar crests, which is common in the central portion of the straight-meandering-straight estuary zone - similar to deposits of the central zone of the Pärnu estuaries.

Bay Mont-Saint-Michel in NW France is another modern example of a macrotidal estuary described by Lanier and Tessier (1998). Morphologically, the evolution of Bay of Mont-Saint-Michel from its head to the mouth is quite well the tripartite model by Dalrymple et al. (1992). The morphosedimentary model of the Mont-Saint-Michel can also be applicable to Pärnu estuaries, except that Mont-Saint-Michel estuary has mixed siliciclastic- carbonate sediments (Tessier, 2012). Moreover, here, along the estuarine zone, extended intertidal and marsh areas are developed with very fine-grained sand to silt sediments similar to supratidal and intertidal flats adjacent to estuaries (Tessier, 1998). This indicates the major difference with Pärnu estuaries, which are characterized by more coarse-grained composition of beds and the occurrence of supratidal mudflats instead of marshes in prevailing arid tropical climate conditions with lack of vegetation (Kurshs, 1992). Another major difference is with regards to the lack of elongated sand ridges within the outer portions (mouth) of the estuary, which on the contrary is very common for tide-dominated estuaries of Pärnu basin. Mont-Saint-Michel is located along the open coast and as such, wave dynamics from the English channel preclude the formation of elongate sand ridges system (Lanier, Tessier, 1988), which is not the case for Pärnu basin, where wave influence to the deposition is minimal. Moreover, sediments in this area of the Mont-Saint-Michel estuary are medium- to fine-grained sands, which do not allow large dune formation. Thus, the mouth of the estuary consists essentially of a wide braided-bar system (Lanier, Tessier, 1998).

However, the major depositional trend, as well as sedimentary environments at the local scale within the tidal-fluvial transition zone and proximal reaches of the outer estuary of Mont-Saint-Michel (see Figure 6.1.A-E), reflect similar facies and tidal signatures with deposits of Pärnu estuaries, including rhythmical variations of the shape of climbing ripple lamination, described in Chapter 5.1.4. Although environment of periodic rapid accumulation of sediments is favorable for the development of climbing-ripple lamination (Reineck, Singh, 1980), very few studies describe the presence of climbing ripples in tide-influenced environments (Yokokawa et al., 1995; Lanier, Tessier, 1998; Van den Berg et al., 2007). Climbing ripple lamination in Bay Mont-Saint-Michel is common for the fluvial-estuarine transition zone in an area with a well developed straight-meandering-straight morphology. Ebb-dominated climbing ripple lamination at Mont-

Saint-Michel, similar to those derived in the outcrop 7, is associated with tidal bars in proximal reaches of the outer estuary (FA3, outcrop area).

From ancient estuarine examples the study of Eocene Central Basin, Spitsbergen, by Plink-Björklund (2005) gives facies description and high-resolution sequence stratigraphic analysis on tide-dominated estuary facies in rock record. Another study by Plink-Björklund (2008) on Chimney Rock Tongue, Upper Cretaceous, Campanian, Utah–Wyoming, USA, also provides example of tide-dominated estuarine successions. Rossetti (1998) reports on Cujupe Formation, Upper Cretaceous to Lower Tertiary São Luis Basin, Brazil, by giving facies analyses and depositional reconstruction of an estuary, that illustrates tide-influenced facies, which infill an incised valley. However unlike Pärnu Fm, the facies succession is indicative to a wave-dominated estuary environment. Gjelberg and Steel (1995) describe tidal estuary that consists of the similar facies as Pärnu RS from the Helvetiafjellet Formation (Barremian-Aptian) in Spitsbergen. The facies here consist of similar mainly fine- to medium-grained sandstone, except in the basal part where coarser sandstone with pebbles and thin conglomerates occur, marked by erosion surface. These deposits in the lowermost part of the section interpreted as fluvial deposits, are preserved in the bottom of a small incised valley, analogical to FAs of SU 2 in outcrop area. According to authors' interpretation, an estuary developed as a consequence of a relative sea-level rise. The estuarine deposits reflect high sediment input to an environment dominated both by fluvial and tidal processes, whereas the bulk of Pärnu Fm in the study area consists of tidal-fluvial transition and tidal bar deposits. Large-scale beds in the middle part of this unit represent lateral accretion of tidal sand bars, similar to deposits of FA 2 and 3 of Pärnu succession in the outcrop area.

6.5. Tidal ranges and tidally controlled epicontinental basins

This chapter deals with a discussion on tidal ranges in the Rēzekne and Pärnu basins and provides an insight in the role of tidal processes in epicontinental (epeiric) seas, such as the BDB. Tidal range is the elevation difference between the high tide and the succeeding low tide. Sedimentary environments are generally classified according to their tidal ranges as microtidal (0-2 m), mesotidal (2-4 m) and macrotidal (>4 m) (Johnson, Baldwin, 1996). This is important in reconstructing depositional environments of ancient tidalites. In the geological record, a combination of observations from regional and outcrop scale is essential to distinguish tidal ranges, but even that does not give a precise understanding of tidal ranges which were present in the geological history.

In the geological past, large areas of the continental interiors were covered by shallow epicontinental seas (Klein, Ryer, 1978), such as the BDB including in Rēzekne and Pärnu times (Kurshs, 1975; 1992). Understanding of these basins is limited due to the absence of suitably scaled Holocene analogs for ancient epicontinental seas. Shallow, large seas that may provide relatively close modern analogies are the East China and Yellow seas, the Arafura Sea, the Java Sea, the North Sea and the Hudson Bay (Klein, Ryer, 1978). Sedimentation on modern shelves is considered to be dominant either by tides or waves and storms (Boyd et al, 1992). The principal problem with drawing modern analogies is that the Holocene shallow marginal shelf seas mark the flooding of the ocean into continental shelves, whereas ancient epicontinental seas were isolated from the open ocean, had low angle bottom topography (Wells et al., 2005a) and covered larger areas than their modern counterparts. It is also worth to note, that the Earth/Moon relationship has changed over geological time as the distance between Earth and Moon increased (Kvale et al., 1999). The Moon is currently retreating from Earth at a rate of 3.82 cm a^{-1} (Dickie et al., 1994). This also possibly caused changes in tidal range, most likely a decrease on a global scale, which complicates drawing the analogies with modern tidal environments. Although shallow Holocene continental shelf seas differ from ancient epicontinental seas in above important aspects, tidal circulation pattern of these modern seas serve as a clue to the water circulation pattern of ancient shallow basins, in particular, the correlation of shelf width with tidal amplitude and velocity of tidal currents (Klein, Ryer, 1978).

The lack of modern analogues has led to a debate on the role of tidal processes in ancient epicontinental seas. One theory suggests that these seas were atidal due to rapid landward attenuation of tidal wave energy (Shaw, 1964). Later Irwin (1965) and Boggs (2001) stated that tidal waves could not penetrate over the shallow depths and great sizes of these seas. In contrast, another theory suggests that these seas were tide-dominated, such as the early studies by Klein and Ryer (1978) that gave both modern and ancient examples of tide-dominated conditions in epicontinental seas. There were also efforts in the past to determine the tidal range under which specific tidalite sequence have been deposited (Klein, 1971, 1975). More recently few studies refer to tide-dominated epicontinental seas, due to a positive correlation between shelf width and tidal range observed in some modern settings (Wells et al., 2005a; Wells et al., 2005b; Allison, Wells, 2006; Yoshida et al., 2007; Wells, 2010; Mitchell et al., 2011) and using sedimentological analysis of tidal facies (Ponten, Plink-Björklund, 2005; Tänavsuu-Milkeviciene, Plink-Björklund, 2009; Tänavsuu-Milkeviciene et al., 2009; Plink-Bjorklund, 2012).

Recently, numerical modeling of tides provides a quantitative means to investigate tidality in the geological record (Slater, 1985; Slingerland, 1986, Ericksen, Slingerland, 1990; Austin, Scourse, 1997; Wells et al., 2005a; Wells et al., 2005b; Allison, Wells, 2006). Majority of these studies suggest broad micro- to mesotidal ranges with localized areas of macrotidal activity. The most recent studies by Wells et al. (2005a), Wells et al. (2005b), Wells (2010), Mitchell et al. (2011) use a model with unstructured finite element tetrahedral mesh to predict tides in both ancient and modern 10-200 m deep basins. A number of these computer simulations on tidal dynamics in larger epicontinental seas predict regionally microtidal conditions (Wells et al., 2005a and b), despite the abundant evidence of tidalites in geological record. The authors assumed that indeed some of these seas were tideless at the regional scale, with only local amplification in small coastal embayments. According to Allison and Wells (2006), this can be a result of a function of local funneling and resonance in an embayment, and/or possibly, closer proximity of the Moon to Earth, or some other mechanisms which are yet unknown. However, the extent to which these models apply to all epicontinetal seas is under discussion and few arguments against tidal process simplifications in the above studies exist (see discussion by Higgs, 2006). The result of the study for the Upper Carboniferous epicontinental basin for Northern Europe (Wells et al., 2005a) describe that the vast epicontinental seas of the geological past behaved more like large salty lakes than present-day coastal environments. While according to Mitchell et al. (2011), tide and associated bed shear stress model results of the Early Jurassic Laurasian Seaway of Northwest Europe suggest that flow construction can have a considerable impact on even the largest epicontinental seas, and can occur thousands of kilometers from the deep water of the open ocean.

Few approaches have been proposed for determining palaeotidal ranges in sedimentary rock (Davis, 2012). However today good methods in determining palaeotidal range do not exist other than detailed study of well-preserved, complete stratigraphic sequence from the base to the top of the intertidal zone (Davis, 2012) in combination with the record of the regional geological setting.

In this study it is argued that tide-dominated shallow epicontinental seas existed in the past, and the Rēzekne and Pärnu basins are examples on how tides actually controlled the deposition in this shallow basin. Such suggestion is made based on the combination of cyclic stratigraphic sequences at bed scale and the regional geological setting.

It is known that morphosedimentary data may provide information as to whether an area is wave or tide dominated and if tide-dominated, whether micro-, meso- or macrotidal environments prevail (Hinton, 1998). Sedimentological analysis of the tidalites may be used to determine the paleotidal regime in areas where also good palaeogeographic reconstructions are possible. Thus in the geological record observations from regional and local scales must be combined in order to distinguish tidal ranges (Dalrymple, 1992). Tidal range may be broadly approximated from deposit morphology and sandbody geometry at the regional scale, and from tidal signatures at the outcrop or local scale (Johnson, Baldwin, 1996). Therefore:

1) At outcrop scale tidal deposits are recognized by variety of tidal facies and signatures (for more details see Chapter 5.1.4). However, not necessarily, the wide distribution of tidal signatures would suggest a macrotidal palaoenvironment, therefore regional implications should be also taken into account. In modern tidalite sequence there is a general tendency to equate a complete sequence of sediments at the base of the intertidal environment to those at the upper portion, corresponding to supratidal environment with the actual tidal range (Klein, 1972; Evans, 1975; Knight, Dalrymple, 1975). In these conceptual models the thickness of the complete stratigraphic sequence is equal to the tidal range (Figure 6.2). However, such a model is hypothetical and must include the entire potential tidal sequence (Davis, 2012). Moreover, a complete sequence of these models is almost never preserved in the stratigraphical record. Discrepancies from true tidal range may be caused by basin subsidence, compaction and erosion (Klein, 1972). Many erosional surfaces do not allow distinguishing complete depositional sequences in the deposits of Pärnu Fm. Most commonly the base is difficult to determine and the top portion of such sequences is missing due to erosion. More detailed sequences are however, observed in drill-cores of Rēzekne and Pärnu RS from Eastern Latvia, where carbonate sedimentation prevails and intertidal to supratidal flat deposits are dominant. The base of such sequences is somewhat difficult to recognize because of multiple erosional surfaces, while the top is easier to identify due to presence of desiccation features, such as mud cracks, tepee structures and palaeosols, which are characteristic features of supratidal environments (Brooks et al., 2003; Rankey et al., 2006). The best preserved sequences observed in drill-cores in Eastern Latvia and Belarus illustrate sequences of 5-7 m thickness, which are assumed to be the ones best preserved within all succession. Tidal rhythmites usually form in macro-tidal settings, but tidal current speeds must be low enough to preserve the delicate sand-mud couplets without eroding them (Wells et al., 2005a). This in addition reconfirms that the environment in northern part of the basin was macrotidal, as almost no tidal rhythmites are found there or those found are eroded. On the contrary, most of them are preserved in eastern and southern part of the basin, where tidal flat sedimentation prevailed.

2) At a regional scale, the tidal range influences shoreline morphology and this is determining the geometry of sandbody (Johnson, Baldwin, 1996). Strong tidal amplification occurs when tidal waves propagate into the broad and shallow shelf sea, consequently rising tidal range shoreward (Daidu, 2012). It is undoubtedly easier to identify ancient macrotidal environments than microtidal ones. Most micro- and mesotidal areas are wave dominated, whereas some mesotidal and most macrotidal areas are tide dominated. However, if wave action is limited due to topographic sheltering, or the tidal current speeds are increased by a topographic construction, tidal dominance can locally occur in microtidal areas (Dalrypmle, 1992). Combination of little wave influence, amplification of tidal energy, especially due to the resonance effects as the basin passes

into the tidal amplification window, are the main controls on the occurrence of tide-dominated or strongly tide-influenced successions (Dalrymple, Choi, 2007; Willis, 2005; Yoshida et al., 2007). It is known, that tide-dominated estuaries and tidal flats form along coastlines experiencing high tidal ranges (Dalrymple, 1992). The deposits of these environments provide the bulk of the Rēzekne and Pärnu successions. Tidal flat deposits are the main depositional environments in microtidal settings because of the limited tidal range, however they may be extensive in mesotidal environments. Similarly, tidal channels are likely to be more prevalent in macro- than in microtidal environment, because of the stronger tidal currents generated by larger tidal range. Tidal range is also highly affected by resonance and shoaling in localized, shallow water and coastal embayments, such as the palaeocoastline area of the Pärnu basin. It formed during relative sea level lowstands and the early stages of subsequent transgressions when incised valleys were cut and flooded producing a complex, embayed coastline at the northern part of the basin. Such embayments also increase the possibility of tidal amplification by funneling and shoaling (Dalrymple, 1992).

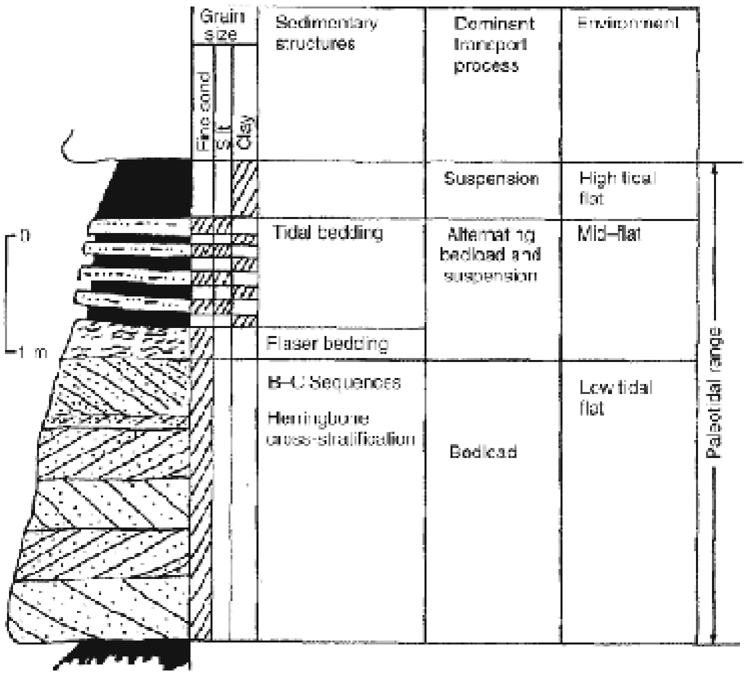

Figure 6.2. A generalized sequence used for determining tidal ranges in rock record
The thickness of the interval representing subtidal, intertidal and supratidal sediments, coincides with the mean tidal range. In ancient deposit equivalent, the thicknes of similar sequences gives a quantitative measurement of palaeotidal range (after Davies, 2012; Klein, 1971)

6.2. attēls. Vienkāršots nogulumu griezums, ko izmanto plūdmaiņu amplitūdu noteikšanai senajos nogulumos
Intervāla biezums, kurā ietilpst apakšējā, vidējā un augšējā plūdmaiņu līdzenumu nogulumi, atbilsts plūdmaiņu amplitūdai. Seno nogulumu analogos atbilstošu slāņkopu biezums norāda uz plūdmaiņu amplitūdu (pēc Davies, 2012; Klein, 1971)

Tidal evidence in facies of the Rēzekne and Pärnu RS, that are wide-spread through the whole territory of its distribution area together with bedform morphology and palaeogeographical data (Kurshs, 1992), strengthens the argument for at least locally effective tides and macro-tidal environment in these epicontinental basins. Tidal range also varies with changes in relative sea level. Tide-dominated environments are more common in inner-shelf reaches during transgressions in open-shelf settings and during lowstands in epicontinental seas (Yoshida et al., 2007), such as is the Rēzekne and Pärnu basins. The recent computer simulations and models also predict that the largest tidal ranges were generated in localized coastal embayments, mainly in the early stages of transgression, such as in the Rēzekne and Pärnu basins. Although Plink-Björklund (2012) states that since BDB was a very shallow-water basin with depth of 20-40 m throughout its evolution, it has experienced significant wave energy dissipation and tidal energy amplification even independently of the sea level or sediment supply changes. The gradient of palaeocoastline was low, which allowed tides to penetrate landwards. It is clear that local shoreline morphology varied greatly, but it is evident that tidal currents were locally strong enough to transport medium-grained sand in simple dunes and bars as in most part of the basin. Therefore, it is suggested that the overall tidal regime in Rēzekne and Pärnu basins was in order of high meso- to macrotidal. It is likely to have been macrotidal (>4 m tidal range) on the northern, north-eastern and north-western coastlines, though on the southeastern side of the basin there is likely to have been a meso- to macrotidal regime (2-4 m tidal range).

Conclusions

This thesis reports on a depositional history of the Devonian Rēzekne and Pärnu time basins in the Baltic States, based on facies analysis and detailed vertical and lateral correlation of the successions derived from the outcrops and drill-core data. It reveals a much more complex depositional history of the Rēzekne and Pärnu basins compared to the results of previous studies, which suggested deposition of Rēzekne and Pärnu successions of the Baltic palaeobasin in shallow marine and deltaic environments (Kurshs, 1975, 1992; Kleesment, 1997).

- In this study the Rēzekne and Pärnu successions are re-interpreted as tide-dominated transgressive estuarine and tidal flat system. Several estuaries possibly existed along the palaeocoastline. The vertical cyclic pattern in transition of facies and fining-upward trend for the entire succession throughout the basin is considered to reflect a retrograding estuarine succession and subtidal to intertidal to supratidal sandy to muddy (including carbonate) tidal flat system which fringe the shoreline.
- Nine documented facies associations reflect deposition in a shallow epicontinental basin: intertidal to supratidal carbonate mudflats, intertidal to supratidal carbonate shoals, palaeosols, fluvial and tidally-influenced fluvial deposits of inner estuary, central estuary deposits, tidal channels and bars, tidal ridges of outer estuary, intertidal to subtidal channels and flats, and intertidal to supratidal mudflat. Four of them (fluvial and tidally-influenced fluvial deposits of inner estuary, central estuary deposits, tidal channels and bars, intertidal to subtidal channels and flats) have been identified and described in detail in the outcrop area.
- The data derived from the outcrop area represents an excellent example of sand-rich ancient estuary, which is little described in the rock record. Few indications for recognition of sand-rich estuaries in rock record have been given.
- The succession in the outcrop belt and adjacent drill-cores represents relatively thick beds of a fluvial-tidal transition zone, which is one of the most interesting zones in terms of facies associations and differs hydraulically from the upstream fluvial- and downstream tide-dominated estuarine zone. The recognition of tidal signatures in fluvial-tidal deposits is important as it enables the determination of the farthest extent of a transgression, especially taking into account the great length of the fluvial-tidal transition zone in many low-gradient rivers, such as were presumaby rivers entering the BDB. Thanks to detailed recognition criteria of tidal signatures in fluvial-tidal transition zone, as well as in sand-rich estuaries, new field data have demonstrated tidal dominance in the Rēzekne and Pärnu time basins. The distinctive structures described as indicative of the fluvial-tidal zone and proximal parts of the outer estuary can not be considered as exclusively diagnostic for the recognition of these zones. These criteria here should be used as a complex of diagnostic features and have a comparative value.
- The outcrop area (and adjacent drill-cores) has the highest sand content in comparison to other parts of the basin and is interpreted to have been closer to the palaeocoastline, and thus to the terrestrial areas occupied by rivers. The orientations of the bedforms are determined by cross-stratification dip measurement data. Fluvial transport directions towards SE–SW suggest that

the palaeoshoreline was roughly NW-NE trending, however the exact position of rivers is not known. Shallow water (up to supratidal) at the basin margins prevailed, with gradually deeper water (subtidal) towards the south.

- Sediment was derived from two main source areas: a fluvial drainage area in the present north and a marine source (by tidal currents) in the present south. However the sediment input from south by fluvial drainage is not excluded as well, as indicated by some coarse-grained composition of beds from the drill-cores in eastern part of the study area. The deposition was in general controlled by tidal currents throughout the basin. Wave action had a subordinate role in sediment transport and deposition in Rēzekne and Pärnu basins.
- In order to differentiate tide-dominated estuaries from tide-dominated deltas in the rock record, it has been thus crucial to focus not only on detailed facies analysis and distinguishing tidal signatures, but to emphasize the overall stacking pattern of the succession. In this study Rēzekne and Pärnu RS is re-interpreted as a tripartite transgressive tide-dominated estuarine succession and associated tidal flat complex based on: (1) sharp base of the succession, (2) fining-upward vertical stacking pattern of stratigraphic units that reflects upward-increasing tidal influence, (3) lateral facies changes that reflect a coarse-fine-coarse seaward grain-size trend, (4) sharp base and "blocky" character of the tidal bars and sand ridges, (5) common bidirectional palaeocurrent distribution, reflecting two sediment sources.
- In the outcrop area Pärnu Fm is subdivided into 3 stratigraphic units. SU 1 is represented by tide-influenced fluvial to tidal channels and bars erosively overlain by channelized, immature cross-stratified conglomerate and sandstone beds which mark the base of SU 2 and are in places transitionally, but in most cases erosively, overlain by variably-grained channel thalweg fill and bar deposits with less variability in tidal signatures. The basal conglomerates and sandstones of the lowermost unit are poorly-sorted and contain intra-basinal clay clasts, as well as fossil plant and fish remains. Texturally mature fine-grained cross-stratified sandstones erosively overlay beds of fluvial-tidal transition zone. These beds are well-sorted and contain variable tidal signatures. In the southwestern reaches of the outcrop belt these deposits are overlain erosively by SU 3, representing again a new cycle with fluvial-tidal transition zone deposits. The topmost part of the section consists mainly of very fine-grained and carbonate-rich deposits. With each transgression, the facies of transitional sediments moved up and down the lower river reach, thus marking the extent of a transgression.
- The three stratigraphic units of the studied succession in Rēzekne and Pärnu RS represent three major episodes of incision and infill. In general, the distinctive unconformity between the Pärnu Fm and the underlying Silurian carbonate succession in the outcrop area, combined with the character of vertical variation in facies, suggests that the deposits of the Pärnu Fm formed in an incised valley, followed by progressive infill by fluvial processes during lowstand. As indicated by the drill-cores, the depth of incision was approximately 15 m. The bulk of the succession was deposited in the tide-dominated estuarine environment during flooding (early transgressive stage) of a previously created incised valley with a pronounced unconformity at the base. Tidally dominated facies above, capped by tidal flat deposits of

topmost part, passing into more carbonate deposits of Narva Fm, reflect subsequent transgression and infill of the estuaries.

- Early opinion about the role of tides in ancient epicontinental basins, including the BDB, as an epicontinental basin, was that co-oscillating tides entering the basin could not propagate large distances within the seaway because of rapid attenuation of tidal wave energy. However, evidence accumulated from facies data, is opposing this suggestion. High mesotidal to macrotidal ranges are suggested to be dominant in the BDB of Rēzekne and Pärnu times.

References

Adams J.E, Frenzel H.N., 1950. Capitan barrier reef, Texas and New Mexico. *Journal of Geology, 58*, pp. 89–312

Allen P.G., 1991. Sedimentary processes and facies in Gironde estuary: a recent model for macrotidal estruarine systems. *In* Smith D.G., Reinson G.E., Zaitlin B.A., Rahmani R.A. (eds.), *Clastic tidal sedimentology*. Canadian Society of Petroleum Geologists, Memoir, *16*, pp. 29-40.

Allen G.P., Posamentier H.W., 1993. Sequence stratigraphy and facies model of an incised valley fill: Gironde Estuary, France. *Journal of Sedimentary Petrology*, *63*, pp. 378-391.

Allison P.A., Wells M.R., 2006. Circulation in large ancient epicontinental seas: what was different and why? *Palaios*, *21*, pp. 523-515.

Archer A.W., 1996. Panthalassa: paleotidal resonance and a global paleocean seiche. *Paleoceanography, 1*, pp. 625–632.

Archer A.W., Feldman H.R., Kvale E.P., Lanier W.P., 1994. Comparison of drier and wetter-interval estuarine roof facies in the Eastern and Western Interior coal basins, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology*, *106*, pp. 171-185.

Archer A.W., Greb S.F., 1995. An Amazon-scale drainage in the early Pennsylvanian of Central North America. *Journal of Geology*, *103*, pp. 611–628.

Archer A. W., Greb S.F., 2012. Hypertidal Facies from the Pennsylvanian Period: Eastern and Western Interior Coal Basins, USA. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 421-436.

Ashley G.M., 1990. Classification of large-scale sub-aqueous bedforms: a new look at an old problem. *Journal of Sedimentary Petrology*, *60*, pp. 160–172.

Ashley G.M., Southard J.B., Boothroyd J.C., 1982. Deposition of climbing-ripple beds: a flume simulation. *Sedimentology, 29*, pp. 67-79.

Austin W.E.N., Scourse, J.D., 1997. Evolution of seasonal stratification in the Celtic Sea during the Holocene. *Journal of the Geological Society, London,154,* pp. 249–256.

Bathurst R.G.C., 1975. Carbonate sediments and their diagenesis. *Developments in sedimentology, 12,* Elsevier. 658 pp.

Butler G.P., 1970. Holocene gypsum and anhydrite of the Abu Dhabi, Trucial Coast, Persian Gulf: An alternative explanation of origin. *Proceedings of 3rd International Salt Symposium.* Northern Ohio Geological Society, Cleveland, pp 120–152.

Boersma J.R., 1969. Internal structure of some tidal mega-ripples on a shoal in Westerschelde estuary, the Netherlands: report of a preliminary investigation. *Geologie en Mijnbouw 48, 4*, pp. 409-414.

Boggs S.N., 2001. *Principles of sedimentology and stratigraphy*. 3rd edition. Prentice-Hall, Englewood Cliffs.

Boyd R., Dalrymple R.W., Zaitlin B.A., 1992. Classification of clastic coastal depositional environments. *Sedimentary Geology, 80*, pp. 139–150.

Boyd R., Dalrymple R.W., Zaitlin B.A., 2006. Estuarine and incised-valley facies models. *In* Posamentier H.W.,Walker R.G. (eds.), *Facies models revisited.* SEPM Special Publication, *84*, pp. 171-235.

Brangulis, A. J., Kuršs, V., Misāns, J., Stinkulis Ģ., 1998. *Latvijas ģeoloģija. 1:500 000 mēroga ģeoloģiskā karte un pirmskvartāra nogulumu apraksts.* Rīga, Valsts ģeoloģijas dienests, 70 lpp.

Bridge J.S., Jalfin G.A.,Georgieff S.M., 2000. Geometry, lithofacies and spatial distribution of cretaceous fluvial sandstone bodies, San Jorge basin, Argentina: outcrop analog for the hydrocarbon-bearing Chubut Group. *Journal of Sedimentary Reaserch, 70,* pp. 341-359.

Brooks G.R., Doyle L.J., Suthard B.C., Locker S.D., Hine, A.C., 2003. Facies architecture of the mixed carbonate/siliciclastic inner continental shelf of west-central Florida: implications for Holocene barrier development. *Marine Geology, 200,* pp. 325–349.

Chaumillon E, Proust J-N., Menier D., Weber N., 2008. Incised valley morphologies and sedimentary fills within the inner shelf of the Bay of Biscay (France): a synthesis. *Journal of Marine Systems, 72,* pp. 383–396.

Cocks L.R.M., Torsvick T.H., 2005. Baltica from the late Precambrian to mid-Palaeozoic times: The gain and loss of a terrane's identity. *Earth Science Reviews, 72,* pp. 39-66.

Cocks L.R.M., Torsvick T.H., 2006. European geography in a global context from the Vendian to the end of the Palaeozoic. *In* Gee D.G., Stephenson R.A. (eds.), *European Lithosphere Dynamics.* Geological Society of London, Memoir 32, pp. 83–95.

Collinson J.D., 1996. Alluvial sediments. *In* Reading, H.G. (eds.), *Sedimentary environments: processes, facies and stratigraphy*. Oxford, U.K., Blackwell Science Ltd, Oxford, 3-rd edition, pp. 20-54.

Daidu F., 2012. Open coast tidal flats. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 187-230.

Dalrymple R.W., 1984. Morphology and internal structure of sandwaves in the bay of Fundy. *Sedimentology, 31*, pp. 365-382.

Dalrymple R.W., 1992. Tidal depositional systems. *In* Walker R.G., James N.P. (eds.), *Facies Models: Response to Sea-level Change*. Geological Association of Canada, St Johns, Canada, pp. 195-218.

Dalrymple R.W, 2006. Incised valleys in time and space: an introduction to the volume and an examination of the controls on valley formation and filling. *In* Dalrymple, R.W., Leckie, D.A., Tillmann, R.W. (eds.), *SEPM Special Publication, 85*, pp. 5-12.

Dalrymple R.W., Choi K.S., 2004. Proximal-distal trends in tide-dominated depositional systems (abstract): *Tidalites-2004, 6th International Conference on tidal sedimentology*. Copenhagen, Abstract volume, pp. 38-40.

Dalrymple R.W., Choi K.S., 2007. Morphologic and facies trends through the fluvial-marine transition in the tide-dominated depositional system: A schematic framework for environmental and sequence-stratigraphic interpretation. *Earth-Science Reviews, 81*, pp. 135-174.

Dalrymple R.W, Boyd R., Zaitlin B.A., 1994. *Incised-valley systems: origin and sedimentary sequences*. *SEPM Special Publication, 53*, p. 391.

Dalrymple R.W., Knight R.J., Zaitlin B.A., Middleton G.V., 1990. Dynamics and facies model of a macrotidal sand-bar complex, Cobequid Bay-Salmon River Estuary (Bay of Fundy). *Sedimentology, 37*, pp. 577-612.

Dalrymple R.W., Makay D.A., Ichaso A.A., Choi K.S., 2012. Processes, morphodynaics and facies of tide-dominated estuaries. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 79-108.

Dalrymple R.W., Rhodes R.N., 1995. Estuarine dunes and bars. *In* Perillo G. (eds.), *Geomorphology and Sedimentology of Estuaries. Developments in Sedimentology*, Elsevier Science, New York, pp. 359-422.

Dalrymple R.W., Zaitlin B.A., Boyd R., 1992. Estuarine facies models: conceptual basis and stratigraphic implications. *Journal of Sedimentary Petrology, 62(6),* pp. 1130-1146.

Dalrymple R.W., Baker E.K., Harris P.T., Hughes M.G., 2003. Sedimentology and stratigraphy of a tide-dominated, foreland-basin delta (Fly River, Papua New Guinea). *In* Hasan Sidi F., Nummedal D., Imbert P., Darman H., Posamentier H.W. (eds.),

Tropical Deltas of Southeast Asia - Sedimentology, Stratigraphy, and Petroleum Geology. SEPM Special Publication, 76, pp. 147–173.
Dalrymple R.W., Zaitlin B.A., 1994. High-resolution sequence stratigraphy of a complex, incised valley succession, the Cobequid Bay–Salmon River estuary, Bay of Fundy, Canada. *Sedimentology, 41*, pp. 1069–1091.
Dalrymple R.W., 2010. Introduction to siliciclastic facies models. *In* James N.P., Dalrymple R.W. (eds.), *Facies models 4*. Geological Association of Canada, St. John's, pp 59–72.
Davis R.A., 2012. Tidal signatures and their preservation potential in stratigraphic sequences. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 35-56.
Davies P.J., Bubela B., Ferguson J., 1978. The formation of ooids. *Sedimentology 25*, pp. 703–729.
Deelman J.C., 1978. Experimental ooids and grapestones: carbonate aggregates and their origin. *Journal of Sedimentary Petrology, 48*, pp. 503–512.
Demicco R.V., Hardie L.A., 1994. Sedimentary structures and early diagenetic features of shallow marine carbonates. *Atlas Series 1,* 265 p.
de Raaf J.F.M., Boersma J.R., 1971. Tidal deposits and their sedimentary structures (seven examples from Western Europe). *Netherlands Journal of Geosciences, 50,* pp. 479–504.
Dickie J.O., Bender P.L., Faller J.E., Newhall, X.X., Wiant, R.L., Williams, J.G., Yoder C.F., 1994. Lunar laser ranging: A continuing legacy of the Apollo program. *Science, 265,* p. 482–490.
Duguid S.M.A, Kyser T.K., James N.P., Rankey E.C., 2010. Microbes and ooids. *Journal of Sedimentary Research, 80*, pp. 236–251.
Els B.G., Mayer J.J., 1998. Coarse clastic tidal and fluvial sedimentation during a large late archean sea-level rise: the Turffotein subgroup in the Vredefort structure, South Africa. *In* Alexander C.R, Davis R.A., Henry V.J. (eds.) *Tidalites: Processes and Products. SEPM Special Publication, 61*, pp. 155-165.
Elrick, M. 1995. Cyclostratigraphy of Middle Devonian carbonates of the Eastern Great basin. *Journal of Sedimentary Research, 65,* pp. 61–79.
Emery D., Myers K. (eds.), 1996. *Sequence stratigraphy*. Blackwell Science, 297 pp.
Ericksen M.C., Slingerland R., 1990. Numerical simulations of tidal and wind-driven circulation in the Cretaceous Interior Seaway of North America. *Geological Society of America Bulletin, 102,* pp. 1499–1516.
Eriksson K.A, Simpson E., 2012. Precambrian tidal facies. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 397-420.
Evans G., 1975. Intertidal flat deposits of the Wash, western margin of the North Sea. *In* Ginsburg R.N. (eds.) *Tidal deposits*. Springer, New York, pp. 13–20.
Fenies H., Tastet J-P., 1998. Facies and architecture of an estuarine tidal bar (the Trompeloup bar, Gironde Estuary, SW France). *Marine Geology, 150*, pp. 149-169.
Fenies H., Resseguier A., Tastet J., 1999. Intertidal clay-drape couplets (Gironde estuary, France). *Sedimentology, 46*, pp 1-15.
Flemming B.W., 2012. Siliciclastic back-barrier tidal flats. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 231-268.
Gailīte L.I., Kuršs V., Lukševiča L., Lukševičs E., Pomeranceva R., Savaitova L., Stinkulis Ģ., Zabele,A. 2000. *Legends for geological maps of Latvian bedrock.* State Geological survey, Riga, 103 p.
Garnaud S., Lesueur P., Clet M., Lesourd S., Garlan T., Lafi te R., Brun-Cottan J-C., 2003. Holocene to modern fine-grained sedimentation on a macrotidal shoreface to inner-shelf setting (Eastern Bay of the Seine, France). *Marine Geology, 202*, pp. 33–54.

Gjelberg, J., Steel, R., 1995. Helvetiafjellet Formation (Barremian-Aptian), Spitsbergen: characteristics of a transgressive succession. *In* Steel, R., Felt, V.L., Johannesen, E.P., Mathieu, C. (eds.), *Sequence Stratigraphy on the Northwest European Margin.* Norwegian Petroleum Society (NPF). *Special Publication, 5*, pp. 571-573.

Golubcov V.K., Mahnach A.S., 1961. *Faciji teritoriji Belorusiji v paleozoje i rannem mezozoje*. Minsk, p. 183 (Russian).

Harms J.C., Southard J.B., Spearing D.R., Walker R.G., 1975. Depositional environments as interpreted from primary sedimentary structures and stratification sequences. *Society of Economic Palaeontologists and Mineralogists Short Course 2*, 161 p.

Hardie L.A., 1987. Dolomitisation: a critical view of some current views. *Journal of Sedimentary Petrology, 57*, pp.166–183.

Harris, P.T., Baker, E.K., Cole, A.R., Shorts, S.A., 1993. A preliminary study of sedimentation in the tidally dominated Fly River Delta, Gulf of Papua. *Continental Shelf Research, 13*, p. 441–472.

Haq B.U., Schutter S.R., 2008. A chronology of palaeozoic sea-level changes. *Science, 322,* pp. 64-68.

Higgs R., 2006 Discussion on large sea, small tides: the Late Carboniferous seaway of NW Europe. *Journal of the Geological Society, London, 163*, pp. 893–895.

Hinton A.C., 1998. Tidal changes. *Progress in Physical Geography, 22*, pp. 282–294.

Hori K., Saito Y., Zhao Q., Cheng X., Wang P., Sato Y., Li C., 2001. Sedimentary facies of the tide-dominated paleo-Changjiang (Yangtze) estuary during the last transgression. *Marine Geology*, *177*, pp. 331–351.

Irwin M.L., 1965. General theory of epeiric clear water sedimentation. *Bulletin of American Association of Petroleum Geologists, 49*, pp. 445–459.

Johnson H.D., Baldwin C.T., 1996. Shallow clastic seas. *In* Reading, H.G., (eds.), *Sedimentary Environments: processes, facies and stratigraphy*, 3rd. Blackwell Science, Oxford, pp. 232–280.

Jones B.G., Martin G.M., Senapati N., 1993. Riverine-tidal interactions in the monsoonal Gilbert River fandelta, northern Australia. *Sedimentary Geology, 83,* pp. 319-337.

Jopling A.V., Walker R.G., 1968. Morphology and origin of ripple-drift cross-lamination with examples from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology, 38*. pp. 971-984.

Kaufmann B., 2006. Calibrating the Devonian time scale: a synthesis of U-Pb ID-TIMS ages and conodont stratigraphy. *Earth Science Reviews, 76*, pp. 175-190.

Kalamees K., 1988. Some Middle Devonian plants from Estonia. *Proceedings of the Academy of Sciences of the Estonian SSR*. Geology, *37*.

Kendall C.G., Warren J., 1987. A review of the origin and setting of tepees and their associated fabrics. *Sedimentology, 34,* pp. 1007–1027.

Khin K., Myitta, 1999. Marine transgression and regression in the Miocene sequences of northern Pegu (Bago) Yoma, Central Myanmar. *Journal of Asian Earth Sciences, 17,* pp. 369–393.

Kitazawa, 2007. Pleistocene macrotidal tide-dominated estuary-delta succession, along the Dong Nai River, Southern Vietnam. *Sedimentary Geology*, *194*, pp. 115–140.

Kleesment A., 1997. Formation of the territory, Devonian sedimentation basin. *In* Raukas A., Teedumäe A. (eds.), *Geology and Mineral Resources of Estonia.* Estonian Academy Publishers, Tallinn, pp. 205-206.

Kleesment A., 2009. Roundness and surface features of quartz grains in Middle Devonian deposits of the East Baltic and their palaeogeographical implications. *Proceedings of the Estonian Academy of Sciences*. Geology, *58*, pp. 71-84.

Kleesment A.,Mark-Kurik E., 1997. Sedimentary cover, Devonian. Formation of the territory, Devonian sedimentation basin. *In* Raukas A., Teedumäe A. (eds.), *Geology and Mineral Resources of Estonia*. Estonian Academy Publishers, Tallinn, pp. 112-113.

Klein, G.D., 1971. A sedimentary model for determining paleotidal range. *Geological Society Association Bulletin, 82,* pp. 2585–2590.

Klein G.D., 1972. Determination of paleotidal range in clastic sedimentary rocks. *In 24th international geological Congress*. Montreal Compte Rendus, Section 6, Montreal, pp 397–405.

Klein G.D., Ryer T.A., 1978. Tidal circulation patterns in Precambrian, Paleozoic and Cretaceous epeiric and mioclinal shelf seas. *Geological Society of America Bulletin 89,* pp. 1050–1058.

Klein G.D., 1975. Paleotidal range sequences, Middle Member, Wood Canyou Formation (late Precambrian), east California and western Nevada. *In* Ginsburg R.N., (eds.), *Tidal deposits*. Springer, Heidelberg, pp 171–177.

Klein G.D,, Ryer T.A., 1978. Tidal circulation patterns in Precambrian, Paleozoic and Cretaceous epeiric and mioclinal shelf seas. *Geological Society of America Bulletin, 89*, pp. 1050–1058.

Klein G.D., 1998. Clastic tidalites: A partial retrospective view. *In* Alexander C.R, Davis R.A., Henry V.J. (eds.) *Tidalites: processes and products*. SEPM (Society for Sedimentary Geology) Special Publication, 61, pp 5–14.

Kreisa R.D., Moiola R.J., 1986. Sigmoidal tidal bundles and other tide-generated sedimentary structures of the Curtis Formation, Utah. *Geological society of America Bulletin, 97*, pp. 381-387.

Knight R.J., Dalrymple R.W., 1975. Intertidal sediments from the south shore of Cobequid Bay, Bay of Fundy, Nova Scotia, Canada. *In* Ginsburg R.N. (eds.), *Tidal deposits*. Springer, New York, pp 47–55.

Kurshs V., 1975. *Litologija i mineralnije resursi terrigennogo devona Glavnogo Devonskogo Polja*. Zinātne, Riga, 219 s. (Russian).

Kurshs V., 1992. *Devonskoje terigennoje osadkonakoplenije na Glavnom Devonskom Pole*. Zinātne, Riga, 208 s. (Russian).

Kurshs V., Stinkulis G. 1998. Middle Devonian clastic and carbonate transitional facies in the eastern Baltic (abstract): *15th International Sedimentological Congress*, Alicante, Abstracts, pp. 488-490.

Kvale E., 2012. Tidal constituents of modern and ancient tidal rhythmites: criteria for recognition and analyses. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 1-18.

Kvale E.P., Johnson H.W., Sonett C.P., Archer A.W., Zawistoski A., 1999. Calculating lunar retreat rates using tidal rhythmites. *Journal of Sedimentary Research, 69*, pp. 1154-1168.

Kvale E.P., Archer A.W., 1990. Tidal deposits associated with low-sulfur coals, Brazil formation (lower Pennsylvanian), Indiana. *Journal of Sedimentary Petrology, 60,* pp. 563-574.

Kvale E.P., Cutright J., Bilodeau D., Archer A., Hohnson H.R., Pickett B., 1995. Analysis of modern tildes and implications for ancient tidalites. *Continental Shelf Research, 15*, pp. 1921-1943.

Lanier W.P., Feldman H.R. and Archer A.W., 1993. Tidal sedimentation from a fluvial to estuarine transition, Douglas Gropu, Missourian, Kansas. *Journal of Sedimentary Petrology, 63(5),* pp. 860-873

Lanier W.P., Tessier B., 1998. Climbing-ripple bedding in the fluvial-estuarine transition: a common feature associated with tidal dynamics (modern and ancient analogues). *In*

Alexander C.R, Davis R.A., Henry V.J. (eds.) *Tidalites: Processes and Products. SEPM Special Publication, 61*, pp. 109-116.
Lasemi Y., Yahani D., Amin-Rasouli H., Lassemi Z., 2012. Ancient carbonate tidalites. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 567-708.
Lasemi Z., Boardman M.R., Sandberg P.A., 1989. Cement origin of supratidal dolomite, Andros Island, Bahamas. *Journal of Sedimentary Petrology, 59,* pp. 249–257.
Leclair S.F., Bridge J.S., 2001. Quantitative interpretation of sedimentary structures formed by river dunes. *Journal of Sedimentary Research, 72 (5),* pp. 713-716.
Lehrmann D.J., Wan Y., Wei J., Yu Y.-Y., Xiao J., 2001. Lower Triassic peritidal cyclic limestone: an example of anachronistic carbonate facies from the Great Bank of Guizhou, Nanpanjiang Basin, Guizhou province, South China. *Palaeogeography. Palaeoclimatology. Palaeoecology, 173,* pp. 103–123.
Lukševičs E., Stinkulis Ģ., Mūrnieks A., Popovs K., 2012. Geological evolution of the Baltic Artesian Basin. *In* Dēliņa A., Kalvāns A., Saks T., Bethers U., Vircavs V. (eds.), *Highlights of groundwater research in the Baltic Artesian Basin.* Rīga, University of Latvia, pp. 7-52.
Lyarskaja L., 1974. *Pogranichnije nizhne-srednedevonskije otlozhenija Pribaltiki. In* Regionaljnaja geologija Pribaltiki. Riga, s. 45-55. (Russian).
Lyarskaja L., 1981. *Rezeknenskij i pjarnusskij gorizonti. In* Devon i karbon Pribaltiki. Riga, s. 66-92 (Russian).
Mark-Kurik E., Põldvere A. 2012. Devonian stratigraphy in Estonia: current state and problems. *Proceedings of the Estonian Academy of Sciences.* Geology, *61(1),* pp. 33-47.
Mack G.H., 1992, Paleosols as an indicator of climatic change at early Late Cretaceous boundary, southwestern New Mexico. *Journal of Sedimentary Petrology, 62*, pp. 483-494.
Markevich V.P., 1960. The concept of facies. *International Geological Research, 2,* pp. 376-379.
Marshall J.E.A., Astin, T.R., Brown, J.F., Mark-Kurik, E., Lazauskiene, J., 2007. Recognizing the Kačak Event in the Devonian terrestrial environment and its implications for understanding land–sea interactions. *In* Becker, R.T., Kirchgasser, W.T., (eds.), *Devonian Events and Correlations: Geological Society of London, Special Publication 278,* pp. 133–155.
Mellere D., 1994. Sequential development of an estuarine valley fill: the Twowells tongue of the Dakota Sandstone, Acoma Basin, New Mexico. *Journal of Sedimentary Research, 4*, pp. 500–515
Mellere D., Steel R.J., 1996. Tidal sedimentation in Inner Hebrides half grabens, Scotland: the Mid-Jurassic Bearreraig Sandstone Formation. *In*: Geology of siliciclastic shelf seas, Geological Society Special Publication, 117, pp. 49-79.
Miall A.D., 1992. Alluvial deposits. *In* Walker R.G., James N.P. (eds.), *Facies Models: Response to Sea-level Change*. Geological Association of Canada, St Johns, Canada, pp. 119-136.
Miall A.D., 1994. Reconstructing fluvial microform architecture from two-dimensional outcrops: examples from castlegate Sandstone, Brook Cliff, Utah. *Journal of Sedimentary Research, 64 (2),* pp. 146-158.
Miall A.D., 1996. *The geology of fluvial deposits, sedimentary facies, basin analysis and petroleum geology*. Springer Verlag Heidelberg, Berlin, 582 p.
Miall A.D., 2000. *Principles of sedimentary basin analysis.* Springer, 3-rd updated and enlarged edition, 616 p.

Middleton G.V., 1973. Johannes Walthers law on the correlation of facies. *Geological Society of America Bulletin, 84,* pp. 979-988.
Middleton G.V., 1991. A short historical review of clastic tidal sedimentology. *In* Smith D.G., Reinson G.E., Zaitlin B.A., Rahmani R.A (eds.), *Clastic tidal sedimentology.* Canadian Society of Petroleum Geologists, Memoir 16, pp 9-15.
Mitchell A.J., Ulicny D., Hampson G.J., Allison P.A., Gorman G.J., Piggott M.D., Wells M.R., Pain C.C., 2010. Modeling tidal current-induced bed shear stress and palaeocirculation in an epicontinental seaway: the Bohemian Cretaceous Basin, Central Europe. *Sedimentology, 57*, pp. 359–388.
Mitchell A.J., Allison P.A., Gorman G.J., Piggott M.D., Pain C.C., 2011. Tidal circulation in an ancient epicontinental sea: the Early Jurassic Laurasian Seaway. *Geology 39*, pp. 207–210.
Moore R.C., 1949. Meaning of facies. *Geological Society of America Memoir 39*, pp. 1-34.
Mowbray T. De., Visser M.J., 1984. Reactivation surfaces in subtidal channel deposits, Oosterschelde, SW Netherlands. *Journal of Sedimentary Petrology 54,* pp. 811-824.
Narbutas V. 1984. *Krasnocvetnaja formacija nizhnego devona Pribaltiki i Podoliji.* Viļņa, 136 pp. (Russian).
Nemec W., Steel R., Lonne I., 2003. *Sedimentary facies analysis: from processes to system tract.* Lecture compendium, UNIS, Svalbard, 100 p.
Newell N.D., Purdy E.G., Imbrie J., 1960. Bahamian oolitic sand. *Journal of Geology, 68,* pp. 481–497.
Nichols G., 1999. *Sedimentology and stratigraphy.* Blackwell Publishing, 355 p.
Nikishin A.M., Ziegler P.A., Stephenson R.A., Cloetingh S.A.P.L., Furne A.V., Fokin P.A., Ershov A.V., Bolotov S.N., Korotaev M.V., Alekseev A.S., Gorbachevd V.I., Shipilove E.V., Lankreijerc A., Bembinova E.Y.U., Shalimova I.V., 1996. Late Precambrian to Triassic history of the East European Craton: dynamics of sedimentary basin evolution. *Tectonophysics, 268,* pp. 23–63.
Nio S.D., Yang C.S., 1991. Diagnostic attributes of clastic tidal deposits: a review. *In* Smith D.G., Reinson G.E., Zaitlin B.A, Rahmani R.A (eds.), *Clastic tidal sedimentology.* Canadian Society of Petroleum Geologists, Memoir *16*, pp. 3-28.
Nordfjord S., Goff J.A., Austin J.A., Gulick S.P.S., 2006. Seismic facies of incised-valley fills, New Jersey continental shelf; implications for erosion and preservation processes acting during latest Pleistocene-Holocene transgression. *Journal of Sedimentary Research, 12,* pp. 1284–1303.
Owen G., 1996. Experimental soft-sediment deformation: structures formed by the liquefaction of unconsolidated sands and some ancient examples. *Sedimentology 43 (2),* pp. 279–293.
Pape C.H., Cowan C.A., Runkel A.C., 2003. Tidal bundle sequencesc in the Jordan Sandstone (Upper Cambrian), southeastern Minnesota, USA: evidence for tides along inboard shorelines of the Sauk epicontinental sea. *Journal of Sedimentary Research, 71,* pp. 354–366.
Payton C.E., 1977. Seismic Stratigraphy - Applications to Hydrocarbon Exploration. *American Association of Petroleum Geologists Memoir, 26.* Tulsa.
Paškevičius J., 1997. *The Geology of the Baltic Republics.* Vilnius University, Geological Survey of Lithuania, Vilnius, 387 p.
Plink-Björklund P., 2005. Stacked fluvial and tide-dominated estuarine deposits in high-frequency (fourth-order) sequences of the Eocene Central Basin, Spitsbergen. *Sedimentology, 52*, pp. 391-428.
Plink-Björklund P., 2008. Wave-to-tide facies change in a Campanian shoreline complex, Chimney Rock Tongue, Wyoming–Utah, USA. *In* Hampson G., Steel R., Burgess P.,

Dalrymple R.W. (eds.), *Recent Advances in Models of Siliciclastic Shallow-Marine Stratigraphy.* Society for Sedimentary Geology Special Publication, 90, pp. 265–291.

Plink-Björklund P., Björklund L., 1999. Sedimentary response in the Baltic Devonian Basin to post-collisional events in the Scandinavian Caledonides. *GFF, 121,* pp. 79-80.

Plink-Björklund P., 2012. Effects of tides on deltaic deposition: Causes and responses. *Sedimentary Geology, 279*, pp. 107-133.

Polivko I., 1977. *Paleogeografija i paleotektonika rannego i srednego devona Srednej Pribaltiki.* Riga, 163 s. (Russian).

Pontén A., Plink-Björklund P., 2007. Depositional environments in an extensice tide-influenced delta plain, Middle Devonian Gauja Formation, Devonian Baltic Basin. *Sedimentology*, *54*, pp. 969-1006.

Pontén A., Plink-Björklund P., 2009. Regressive to transgressive transits reflected in tidal bars, Middle Devonian Baltic Basin. *Sedimentary Geology*, *218*, pp. 48–60.

Posamentier H.W., Jervey M.T., Vail P.R., 1988. Eustatic controls on clastic deposition 1. – Conceptual framework. *In* Wilgus C.K., Hastings B.S., Kendall S.G., Posamentier H.W., Ross C.A., Van Wagoner J.C. (eds.), *Sea level changes - an integrated approach.* Society of Economic Palaeontologists and Mineralogists Special Publication 42, pp. 109-124.

Potter P.E., 1978. Significance and origin of big rivers. *Journal of Geology*, *86*, pp. 13–33.

Pratt, B.R., 2002. Tepees in peritidal carbonates: origin via earthquake-induced deformation, with example from the Middle Cambrian of western Canada. *Sedimentary Geology, 153,* pp. 57–64.

Purser B.H., Tucker M.E., Zenger D.H., 1994. Problems, progress and future research concerning dolomites and dolomitization. *In* Purser B.H., Tucker M., Zenger D., (eds.), *Dolomites. A volume in honour of Dolomieu.* IAS Special Pblication 21, Blackwell Science, pp. 3-20.

Rahmani R.A., 1988. Estuarine tidal channel and nearshore sedimentation of the Late Cretaceous epicontinental sea, Drumheller, Alberta, Canada. *In.* de Boer P.L et. al. (eds.), *Tide-influenced sedimentary environment and facies.* Reidel Publishing Company, pp. 433-471.

Rankey E.C., Riegl B., Steffen K., 2006. Form, function and feedbacks in a tidally dominated ooid shoal, Bahamas. *Sedimentology, 53,* pp. 1191–1210.

Rankey E.C., Reeder S.L., 2012. Tidal Sands of the Bahamian Archipelago. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 537-566.

Rankey E.C., Reeder S.L., 2009. Holocene ooids of Aitutaki Atoll, Cook Islands, South Pacific. *Geology, 37,* pp. 971–974.

Reading H.G., 1986. *Sedimentary environments and facies.* Oxford, U.K., Blackwell Science Ltd, Oxford, 2-nd edition, 615 p.

Reading H.G., 1986. *Sedimentary environments: processes, facies and stratigraphy*. Oxford, U.K., Blackwell Science Ltd, Oxford, 3-rd edition, 688 p.

Reading H.G., Collinson J.D., 1996. Clastic coasts. *In* Reading, H.G. (eds.), *Sedimentary environments: processes, facies and stratigraphy.* Oxford, U.K., Blackwell Science Ltd, Oxford, 3-rd edition, pp. 154-231.

Reeder S.L., Rankey E.C., 2008. Interactions between tidal flows and ooid shoals, northern Bahamas. *Journal of Sedimentary Research, 78*, pp. 175–186.

Reineck H-E., 1967. Layered sediments of tidal fl ats, beaches and shelfbottom of the North Sea. *In* Lauff G.H., (eds.), *Estuaries.* American Association for the advancement of science publication 83. American Association for the Advancement of Science, Washington DC, pp. 191–206.

Reineck H.-E., Singh I. B., 1980. *Depositional Sedimentary Environments.* Second edition. Springer-Verlag: Berlin, Heidelberg, New York, 551 p.
Reineck H-E., Wunderlic F., 1968. Classification and origin of flaser and lenticular bedding. *Sedimentology, 11,* p. 99–104.
Reinson G.E., 1992. Transgressive barrier island and estuarine systems. *In*: Walker R.G., James N.P. (eds.), *Facies models: response to sea level change*. Geological Associations of Canada, St. John's, pp. 179–194.
Richards M.T., 1994. Transgression of an estuarine channel and tidal flat complex: the Lower Triassic of Barles, Alpes de Haut Provence, France. *Sedimentology*, *41*, pp. 55-82.
Roberts D. 2003. The Scandinavian Caledonides: event chronology, palaeogeographic settings and likely modern analogues. *Tectonophysics, 365*, 283–299.
Rossetti D de F., 1998. Facies architecture and sequential evolution of an incised-valley estuarine fill: the Cujupe Formation (Upper Cretaceous to Lower Tertiary), São Luis Basin, Northern Brazil. *Journal of Sedimentary Research, 68*, pp. 299–310.
Ruskin B.G., Jordan T.E., 2007. Climate change across continental sequence boundaries: paleopedology and lithofacies of Iglesia Basin, northwestern Argentina. *Journal of Sedimentary Research, 77,* pp. 661–679.
Savvitova L., 2002. Latvijas augšdevona stratigrāfiskā iedalījuma revīzija. Atskaite. LU Ģeoloģijas institūts. 24 lpp.
Shanely K.W., McCabe P.J., Hettinger R.D., 1992. Tidal influence in Cretaceous fluvial strata from Utah, USA: a key to sequence stratigraphic interpretation. *Sedimentology, 39,* pp. 905-930.
Shanmugam G., Poffenberger M., Toro Alava J., 2000. Tide dominated estuarine facies in the Hollin and Napo ('T' and 'U') formations (Cretaceous), Sacha Field, Oriente Basin, Ecuador. *American Association of Petroleum Geologists Bulletin, 84*, pp. 652–682.
Shaw A.B., 1964. *Time in stratigraphy*. McGraw-Hill, New York, 365 p.
Shinn E.A., 1986. Modern carbonate tidal flats: their diagnostic features. *Colorado School of Mines Quarterly, 81,* pp. 7–35.
Shinn E.A., 1983. Birdseyes fenestrae, shrinkage pores, and loferites: a reevaluation. *Journal of Sedimentary Petrology, 53,* pp. 619–628.
Slater R.D., 1985. A numerical model of tides in the Cretaceous Seaway of North America. *Journal of Geology, 93,* pp. 333–345.
Slingerland R., 1986. Numerical computation of co-oscillating palaeotides in the Catskill epeiric Sea of eastern North America. *Sedimentology, 33*, pp. 487–497.
Steel R.J., Plink-Björklund P., Aschoff J., 2012. Tidal deposits in the Campanian Western Interior Seaway, Wyoming, Utah and Colorado, USA. *In* Davis R.A., Richard A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 437-472.
Stinkulis Ģ., 1998. Latvijas devona klastisko-karbonātiežu un kaļķakmeņu-dolomītu pārejas zonu sedimentoloģija un mineraloģija. *Doktora darbs*, Latvijas Universitāte, Rīga, 228 lpp.
Stinkulis Ģ., Blāķe D., Upeniece I., Lukševičs E., Zupiņš I. 2011. Clays of the Lode Formation, slump depressions and unique assemblage of fossils in the Liepa (Lode) clay pit. *In* Stinkulis Ģ., Zelčs V. (eds.), *The Eighth Baltic Stratigraphical Conference. Post-Conference Field Excursion Guidebook.* University of Latvia, Riga. pp. 25-32.
Stinkulis Ģ., Spruženiece L., 2011. Dolocretes as indicators of the subaerial exposure episodes in the Baltic Devonian palaeobasin. *In* Lukševičs E., Stinkulis Ģ., Vasiļkova J. (eds.), *The Eighth Baltic Stratigraphical Conference*. Abstracts, Riga, pp. 62-63.
Stupples P., 2002. Tidal cycles preserved in late Holocene tidal rythmites, the Wainway Channel, Romney marsh, southeast England. *Marine Geology, 182,* pp. 231-246.

Sumner D.Y., Grotzinger J.P., 1993. Numerical modeling of ooid size and the problem of Neoproterozoic giant ooids. *Journal of Sedimentary Petrology, 63,* pp.974–982.

Sorokin V., 1981. *Devon i kabon Pribaltiki.* Riga, Zinātne, 502 s. (Russian).

Tanavsuu-Milkieviciene K., Plink-Björklund P., 2009. Recognizing tide-dominated versus tide-influenced deltas: Middle Devonian starta of the Baltic Basin. *Journal of Sedimentary Research, 79*, pp. 887-905.

Tanavsuu-Milkieviciene K., Plink-Björklund P., Kirsimae K., Ainsaar L., 2009. Coeval versus reciprocal mixed carbonate-siliciclastic deposition, Middle Devonian Baltic Basin, eastern Europe: implications from the regional tectonic development. *Sedimentology, 56*, pp. 1250-1274.

Teichert C., 1958. Concept of facies. *American Association of Petroleum Geologists Bulletin, 42*, pp. 2718-2744.

Tessier B., 1998. Tidal cycles: annual versus semi-lunar records. *In* Alexander C.R., Davis R.A., Henry V.J. (eds.), *Tidalites: Processes and Products*, SEPM Special Publication, 61, pp. 69-74.

Tessier B., 2012. Stratigraphy of tide-dominated estuaries. *In* Davis R.A., Dalrymple R.W. (eds.), *Principles of tidal sedimentology*, Springer, pp. 109-128.

Tessier B., Gigot P., 1989. A vertical record of different tidal cyclicities: an example from Miocene Marine Molasse of Digne (Haute Provance, France). *Sedimentology, 36,* pp. 767-776.

Tikhomirov S., 1995. *Etapi osadkonakoplenija devona Russkoj platform i obschije voprosi razvitija i strojenija stratisferi. Moskva*, 444.p. (Russian).

Torsvik T.H., Rehnstrom E.F., 2003. The Tornquist Sea and Baltica-Avalonia docking. *Tectonophysics, 362*, pp. 67–82.

Tucker, M.E., Wright, V.P., 2001. *Carbonate Sedimentology*. Blackwell Science, Oxford, 482 pp.

Van den Berg J.H., Boersma J.R., van Gelder A., 2007. Diagnostic sedimentary structures of the fluvial-tidal transition zone. Evidence from deposits of the Rhine and Meuse. *Netherlands Journal of Geosciences, 86,* pp. 287–306.

Van Wagoner J.C., Mitchum R.M., Campion K.M., Rahmanian V.D., 1990. Siliciclastic sequence stratigraphy in well logs, cores and outcrops. *American Association of Petroleum Geologists Methods in Exploration Series 7,* 55 p.

Visser M.J., 1980. Neap-spring cycles reflected in Holocene subtidal large-scale bedform deposits: A preliminary note. *Geology, 8,* pp. 543-546.

Walker R.G., 1976. Facies models 1. General introduction. *Geoscience Canada, 3,* pp. 21-24.

Walker R.G., 1992. Facies, facies models and modern startgraphic concepts. *In* Walker R.G., James N.P. (eds.), Facies models: response to sea level change: St. Jhon's, Newfoundland, Geological Association of Canada, pp.1-14.

Warren, J.K., Kendall, C.G.S.T.C., 1985. Comparison of sequences formed in marine sabkha (subareal) and salina (subaqueous) settings – modern and ancient. *AAPG Bulletin, 69,* pp. 1013–1023.

Wegmann E., 1963. L'expose original de la notion de facies par Amand Gressly (1814-1865). *Sciences de la Terre, 9(1),* pp. 83-119.

Wells M.R., Allison P.A., Hampson G.J., Piggott M.D, Pain C.C., 2005a. Modelling ancient tides: The Upper Carboniferous epieric seaway of Northwest Europe. *Sedimentology, 52,* pp. 715–735.

Wells M.R., Allison P.A., Piggott M.D., Hampson G.J., Pain C.C., Gorman G.J., 2010. Tidal modeling of an ancient tide dominated seaway, part 2: the Aptian Lower Greensand seaway of Northwest Europe. *Journal of Sedimentary Research, 80*, pp. 411–439.

Wells M.R., Allison P.A., Piggott M.D., Pain C.C., Hampson G.J., De Oliviera C., 2005b. Large sea, small tides: The Upper Carboniferous seaway of Northwest Europe. *Geological Society of London Journa, 162,* pp. 1–5.
Willis B.J., 2005, Deposits of tide-influenced river deltas. *In* Giosan, L., Bhattacharya, J.P., (eds.), *River Deltas - Concepts, Models, and Examples. SEPM, Special Publication 83,* pp. 87–129.
Willis B.J., Sharon G.L., 2003. Formation of deep incisions into tide-dominated river deltas: implications for the stratigraphy of the Sego sandstone, Book cliffs Utah, USA. *Journal of Sedimentary Research, 73(2)*, pp. 246-263.
Willis B.J., Bhatacharya J.P., Gabel S.L., White C.D., 1999. Architecture of a tide-influenced delta in the Frontier Formation of central Wyoming, USA. *Sedimentology, 46*, pp. 667-688.
Woodroffe C.D., Mulrennan M.E., Chappell J., 1993. Estuarine infill and coastal progradation, southern van Diemen Gulf, Northern Australia. *Sedimentary Geology, 83*, pp. 257–275.
Yang C.S., Nio S.D., 1985. The estimation of palaeohydrodinamic processes from subtidal deposits using time series analysis methods. *Sedimentology. 32,* pp. 41-57.
Yang C.S., Nio S.D., 1989. An ebb-tide delta depositional model – a comparison between the modern Eastern Scheldt tidal basin (southwest Netherlands) and the Lower Eocene Roda sandstone in southern Pyrenees (Spain). *Sedimentary Geology, 64*, pp. 175-196.
Yang B.C., Dalrymple R.W., Chun S.S., 2005. Sedimentation on a wave-dominated, open-coast tidal flat, southwestern Korea: summer tidal flat – winter shoreface. *Sedimentology*, *52*, pp. 235–252.
Yang B.C., Dalrymple R.W., Gingras M.K., Chun S.S., Lee H.J., 2007. Up-estuary variation of sedimentary facies and ichnocoenoses in an open-mouthed, macrotidal, mixed-energy estuary, Gomso Bay, Korea. *Journal of Sedimentary Research, 77,* pp. 757–771.
Yoshida S., Steel R.J., Dalrymple R.W., 2007. Changes in depositional processes – an ingredient in a new generation of sequence-stratigraphic models. *Journal of Sedimentary Research, 77*, pp. 447–460.
Yokokawa M., Kishi M., Masuda F., Yamanaka M., 1995. Climbing ripples recording the change of tidal current conditions in the middle Pleistocene Shimosa Group, Japan. *In* Flemming B.W., Bartholomä A. (eds.), *Tidal Signatures in Modern and Ancient Environment.* IAS, Special Publication 24, Blackwell Science, pp. 301-311.
Zaitlin B.A., Dalrymple R.W., Boyd R., 1994. The stratigraphic organization of incised-valley systems. *In* Dalrymple R.W., Boyd R., Zaitlin B.A. (eds.), *Incised-valley systems: origin and sedimentary sequences.* Society for Sedimentary Geology special publication 51, pp 45–60.