GEOLOGICA ULTRAIECTINA

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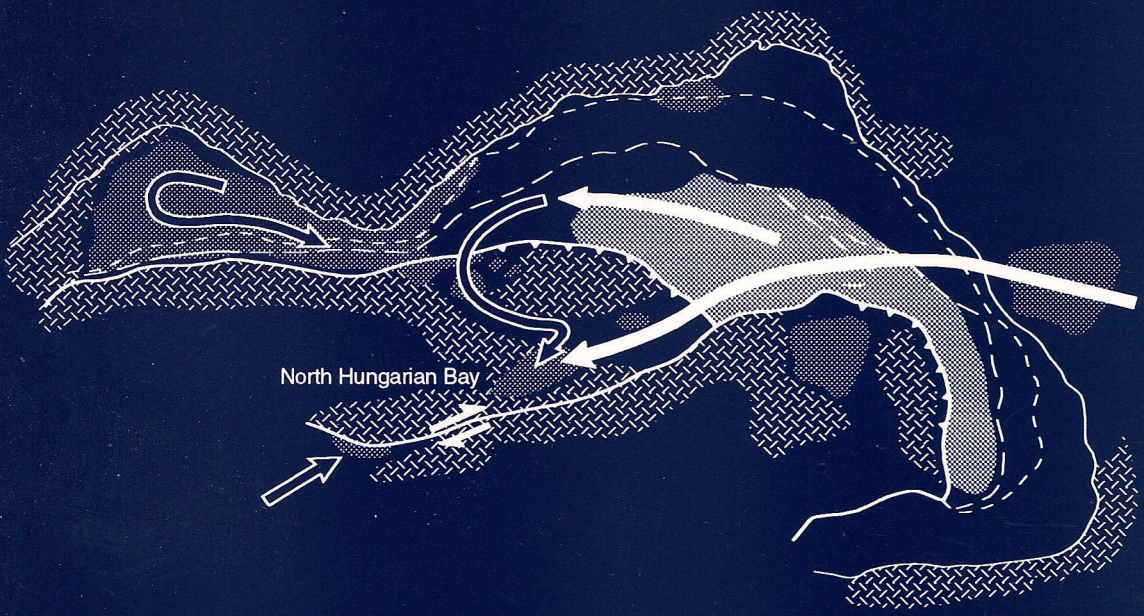
No. 120

THE TIDE-INFLUENCED PÉTERVÁSÁRA SANDSTONE, EARLY MIOCENE, NORTHERN HUNGARY: SEDIMENTOLOGY, PALAEOGEOGRAPHY AND BASIN DEVELOPMENT



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Sztanó, Orsolya
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THE TIDE-INFLUENCED PÉTERVÁSÁRA SANDSTONE, EARLY MIOCENE, NORTHERN HUNGARY: SEDIMENTOLOGY, PALAEOGEOGRAPHY AND BASIN DEVELOPMENT

De vroeg Miocene getijde-beïnvloede Pétervására zandsteen in Noord Hongarije: sedimentologie, paleogeografie en bekkenontwikkeling

Az árapály-formálta Pétervásárai Homokkő szedimentológiája, ősföldrajzi kapcsolatai és az üledékgyűjtő fejlődése (Észak-Magyarország, alsó-miocén)

(met samenvattingen in het Hongaars en het Nederlands)

PROEFSCHRIFT
TER VERKRIJGING VAN DE GRAAD VAN DOCTOR
AAN DE UNIVERSITEIT UTRECHT
OP GEZAG VAN DE RECTOR MAGNIFICUS
PROF. DR. J.A. VAN GINKEL
INGEVOLGE HET BESLUIT VAN HET COLLEGE VAN DEKANEN
IN HET OPENBAAR TE VERDEDIGEN OP MAANDAG
2 MEI 1994 DES MIDDAGS TE 2.30 UUR

DOOR

ORSOLYA SZTANÓ

geboren op 8 mei 1965 te Budapest

Promotores: Prof. Dr. D. Eisma
Prof. Dr. T. Báldi (Eötvös Loránd Universiteit, Budapest, Hongarije)

Co-promotor: Dr. P.L de Boer

CONTENTS

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SUMMARY

Chapter 1

In chapter 1 the Pétervására sandstone is introduced and the aim of this thesis is given. The main purpose of this study was to analyze the sedimentological, palaeogeographical and structural development of the Pétervására Sandstone in the North Hungarian Bay (structurally Hungarian Palaeogene Basin) during the Early Miocene.

Chapter 2

The main stratigraphic, palaeogeographic and structural frame of this thesis is outlined in chapter 2. The Early Miocene is a much debated period in the evolution of the intra-Carpathian area, when a large-scale tectonic reorganization of microplates occurred resulting in large-scale activity of both major and minor tectonic elements. These in turn strongly influenced small-scale basin evolution, sedimentation and preservation of basin-fill deposits. Geographically the area belonged to the Paratethys, a huge inland sea. In the Paratethys both biological and sedimentological events, correlatable over thousands of km, occurred in relation with the large-scale tectonic movements. Thus the fill of the Hungarian Palaeogene Basin is described and the main structural events which influenced its character are shortly discussed.

Chapter 3

Field data, description of outcrop sections, distribution of grain size, bedforms and sediment transport directions of the Pétervására Sandstone are given in chapter 3. Based on this information, four facies belts are determined, characterized by a decrease in grain size and height of the bedforms in the offshore direction. Close to the shoreline conglomerate lobes are interbedded within a sheet-like sand-wave field. The sand waves, up to a height of 12 m, were driven by strong, almost shore-parallel tidal currents. In the offshore direction the sand-wave field declined into a belt of dm-scale dunes and ripples, which finally interfingered with deep-neritic to shallow bathyal siltstones in the more distal areas.

Chapter 4

Large-scale sedimentary phenomena in the Pétervására Sandstone suggests that the paleo-shoreline of the shallow early Miocene sea was defined by the Darnó Fault, part of a long-lived structural zone. Along this zone the Darnó Conglomerate is found. It was formed simultaneously with the Pétervására Sandstone. The sedimentological and petrographical analysis of conglomerates in the Pétervására Sandstone and the coeval Darnó Conglomerate reveals that facies, depositional environments and petrographic composition have much in common. The Darnó Conglomerate was a small fault-controlled fan-delta, producing ophiolite-derived coarse-grained clastics into the Pétervására Sandstone. Conglomerate lobes which occur in the Pétervására Sandstone, were

detached from the fan-delta system and interbedded within the sand-wave field. These lobes were moved by the strong tidal currents, which had been active in the bay. The amount of certain ophiolite-derived heavy minerals and pebbles increases towards the north along the coast, which infers a contemporaneous left-lateral displacement along the Darnó Fault.

Chapter 5

In chapter 5 tidal periods ranging from semi-daily up to yearly are reviewed, This is followed by a discussion of different tidal cyclicities recognized in the large Pétervására sand waves. The lateral thickness variations of foreset laminae indicate a semidiurnal tidal system with well developed daily inequality. Spring-neap, semiannual and annual solar tidal cyclicities can be recognized as well. The collected data allow a quantitative estimation of the tidal range of about 4 m, a spring tidal current velocity of about 0.7-0.9 m/s and both short- and long-term migration velocities of large sand waves of about 0.6-0.8 m/14 days and 10-20 m/year.

Chapter 6

The occurrence of the well developed early Miocene tide-influenced sediments in the North Hungarian Bay far away from the open ocean needs explanation. Moreover, the sedimentary record between the North Hungarian Bay and adjacent Paratethys basins has not been preserved, which makes the route of the tidal waves even more enigmatic. Therefore an analysis of palaeogeographic connections between the North Hungarian Bay and adjacent Paratethys basins is given in chapter 6. It is demonstrated that the connection of the bay with open marine areas must have been through the East Slovakian Basin. Only in that case could tidal waves propagate freely and enter into the bay.

Other examples of tidal deposits of different age in the Alpine region are related to palaeogeographic changes, such as closure and reopening of Paratethyan seaways. All examples demonstrate that tide-influenced deposits are useful tools in palaeogeographic reconstructions.

Chapter 7

A free route for propagation of a tidal wave, discussed in the previous chapter, is a necessary, but not the only requirement for the development of a tide-influenced sedimentary environment. There are many factors, which can reduce or enhance the tidal influence. Those main factors which contributed to the amplification of tidal motions, and to the development of the asymmetrical tidal system in the North Hungarian Bay are shown through a semi-quantitative analysis of the hydro-dynamic conditions in combination with the morphology of the bay The elongate, funnel-shaped morphology of the bay supported the local amplification of tidal waves. Amplification occurred, because the average depth and length of the North Hungarian Bay together with a seaway through East Slovakia provided conditions near resonancy of a semidiurnal tidal wave. Opposed sediment transport directions and asymmetries of the ebb and flood currents between the eastern and the

western side of the bay are interpreted to have been the result of tide-bottom interactions.
A short review of tidal systems at analogous locations demonstrates that analysis of the conditions and the ways of tidal amplification may provide useful information about the development of sedimentary basins.

Chapter 8

In this chapter a sedimentological approach is applied to arrive at a structural reconstruction of basin evolution in northern Hungary. The evolution in time of the shallow marine sedimentary architecture and the presumed effects of the late stage uplift of the flexural basin are discussed. The changes in the Pétervására Sandstone depositional system support the tectonic model recently proposed by Tari el al. (1993).
The second part of the chapter documents and discusses the development of the tide-influenced depositional system in the North Hungarian Bay in response to a marked sea-level fall. This scenario is, as yet, unique between the numerous examples of transgressive tidal systems.

ÖSSZEFOGLALÁS

1. fejezet

Az első fejezet a dolgozatban vizsgált problémát és a kitűzött célokat ismerteti. Elsődleges feladatom a Pétervásárai Homokkő üledékképződési környezetének feltárása volt. Ezen keresztül az Északmagyarországi-öböl (szerkezetileg Északmagyarországi paleogén medence) ősföldrajzi kapcsolatait tártam fel, valamint az üledékgyűjtő fejlődését a korai miocén során.

2. fejezet

A második fejezetben vázolom a dolgozat fő rétegtani, ősföldrajzi és szerkezeti hátterét. A belső kárpáti régió fejlődésének egy sokat vitatott szakasza a korai miocén. Ebben az időszakban a mikrolemezek nagyarányú átrendeződése ment végbe, ami mind a nagyobb, mind a kisebb szerkezeti elemek aktivitásában megnyílvánult. Ez pedig erősen befolyásolta a kisebb üledékgyűjtők fejlődését, magát az üledékképződést, valamint a medencét kitöltő üledékek megőrződésének lehetőségét. Földrajzilag a vizsgált terület a Paratethys hatalmas beltengeréhez tartozott. A Paratethysben az élővilág és az üledékképződés többezer kilométeren át követhető változásai a nagy léptékű szerkezeti mozgásokkal összhangban, részben azok következtében alakultak ki. Ennek megfelelően ismertetem a magyar paleogén medencék üledékeit és a medencefejlődés jellegét meghatározó szerkezeti eseményeket.

3. fejezet

A harmadik fejezetben ismertetem a terepi adatgyűjtés eredményeit, így a Pétervásárai Homokkő legfontosabb szelvényeit, az uralkodó szemcseméret, a rétegformák és a szállítási irányok eloszlását. A terepi adatok alapján négy fáciesegység különíthető el. Ezeket a szemcseméret és a rétegformák méretének csökkenése jellemzi a medence mélyülésének irányában. A parthoz közelebb eső részeken konglomerátum lóbák települnek a lepelszerűen kiterjedt homokhullám mezőbe. Az akár 12 m magasságot is elérő homokhullámokat erős, a parttal közel párhuzamos árapály keltette áramlások mozgatták. A medence mélyebb részei felé a homokhullám mező néhány deciméter nagyságú dűnék, majd homokfodrok övébe ment át. Végül a legtávolabbi területeken mély-neritikus - sekély batiális slírképződményekkel fogazódott össze a Pétervásárai Homokkő.

4. fejezet

A Pétervásárai Homokkő egyes fácieseinek elterjedése arra utal, hogy a korai miocén tenger egykori partvonala a Darnó törés mentén lehetett. E törési zónában található a Darnói Konglomerátum, amely a Pétervásárai Homokkővel egyidőben keletkezett. A homokkőben előforduló konglomerátum testek és az egykorú Darnói Konglomertátum üledékföldtani és kőzettani vizsgálata során beizonyosodott, hogy keletkezési körülményeik és összetételük is sok

közös vonást mutat. A Darnói Konglomerátum egy kis kiterjedésű, a törészóna által meghatározott legyező-deltaként keletkezett. Ezen keresztül jutott az ofiolitos lehordási területről származó durvaszemcsés törmelék a Pétervásárai Homokkő tengerébe. A homokhullám mezőkbe települő konglomerátum lóbák a legyező-delta rendszerről szakadtak le. Mozgásukat az öbölben uralkodó erős árapály keltette áramlások határozták meg. Egyes ofiolitos eredetű nehézásványok és kavicsok mennyisége észak felé, a part mentén nő, amely a Darnó törés üledékképződéssel egyidejű balos oldaleltolódásos jellegére utal.

5. fejezet

Az ötödik fejezetben az árapály tevékenységre jellemző ciklusosságot tekintem át a félnapitól az éves terjedelműig. Ezt a Pétervásárai Homokkő nagy homokhullámaiban felismerhető különböző időtartamú árapály ciklicitás tárgyalása követi. A mellsőlemezek oldalirányu vastagságváltozásai szemidiurnális árapály hatására utalnak, jól fejlett napi egyenlőtlenséggel. A holdciklusaival összefüggő vakár-szökőár, és a nap körüli keringéssel kapcsolatos féléves és éves ciklicitás szintén felismerhető. A mellsőlemezek vastagságváltozásaiból a rendszer egyes fizikai változóinak számszerű becslésére is lehetőség nyílt. Az apály és a dagály közötti szintkülönbség kb. 4 m, az árapály keltette áramlások maximális sebessége (szökőár idején) kb. 0.7-0.9 m/s, a homokhullámok vándorlásának mértéke kb. 0.6-0.8 m lehetett 2 hét alatt, illete 10-20 m évente.

6. fejezet

A biztosan árapály-formálta Pétervásárai Homokkő megjelenése a nyílt óceántól ugyancsak messze elhelyezkedő Északmagyarországi-öbölben magyarázatot igényel. További problémát jelent, hogy az Északmagyarországi-öböl és a szomszédos Paratethys medencék közötti összeköttetést bizonyító üledékek csak nagyon szórványosan maradtak fenn. Ezért a dagályhullám utjának felderítése érdekében, az öböl és a környező Paratethys medencék ősföldrajzi kapcsolatait elemzem a hatodik fejezetben. Bebizonyítom, hogy az egyetlen olyan tengeri összeköttetés, amely lehetővé tette a dagályhullám akadálymentes haladását és eljutását az Északmagyarországi-öbölbe a Keletszlovákiai medencén keresztül vezett.

Az árapály-formálta üledékképződés további különböző korú példái az Alpi térségben mind ősföldrajzi változásokkal, Paratethys átjárók kinyílásával vagy bezáródásával kapcsolatben jöttek létre. Az ismertett példák mindegyike azt bizonyítja, hogy ősföldrajzi rekonstrukciók során sikerrel használhatók az árapály-formálta üledékek.

7. fejezet

Az árapály-uralta üledékképződési környezet kialakulásának szükséges, de nem elégséges feltétele az előző fejezetben tárgyalt, a dagályhullám szabad haladását biztosító tengeri átjáró. Sok olyan további tényező van, amelyek felerősíthetik vagy elnyomhatják az árapály hatását. A hetedik fejezetben azokat a tényezőket vizsgálom amelyek hozzájárultak az árapály hatás felerősödéséhez

és az Északmagyarországi-öbölben tapasztalt aszimmetrikus áramlási rendszer kialakulásához. Ehhez az öböl morfológiáját és az ezzel kapcsolatos hidrodinamikai viszonyokat elemzem. A dagályhullám helyi felerősödését segítette az öböl elnyúlt tölcsér-szerű alakja. A dagályhullám felerősödését okozta az is, hogy az Északmagyarországi-öböl és a Keletszlovákiai-átjáró átlagos vízmélysége és együttes hossza megközelítette a szemidiurnális (félnapos periodusú) dagályhullám rezonanciájának feltételeit (azaz, a beérkező és az öböl fejénél visszaverődő hullám interferencia révén egymást erősítette). Az öböl nyugati és keleti felén észlelt ellentétes irányú és aszimmetriájú apály és dagály áramlás pedig, az árapályhullám és az aljzat kölcsönhatásakor gerjesztett, cirkuláris maradék áramlás hatására jött létre.

Egyéb, hasonló árapály-formálta üledékek példáinak rövid áttekintése azt bizonyítja, hogy az árapály tevékenység feltételeinek gondos elemzésével az üledékgyűjtők fejlődésére (vízmélység, hosszúság, szélesség) is nyerhetők adatok.

8. fejezet

Ebben a fejezetben üledékföldtani megközelítéssel jutok el az észak-magyarországi alsó-miocén üledékgyűjtő szerkezeti fejlődésének rekonstrukciójához. A sekélytengeri üledékek felépítményének időbeli változásait és a flexurális-medence fejlődésének kései szakaszában bekövetkező kiemelkedés hatását vizsgálom. A Pétervásárai Homokkő üledékképződési rendszerében bekövetkező változások úgy tűnik alátámasztják Tari et al. (1993) által javasolt szerkezeti modellt.

A fejezet második felében igazolom, hogy az Északmagyarországi-öbölben egy jelentős tengerszint csökkenés következtében alakult ki az árapály-formálta üledékképződés. Ez a jelenség pillanatnyilag egyedülálló az árapály üledékek tengerelöntéshez kapcsolódó számtalan példája között.

SAMENVATTING

Dit proefschrift behandelt de sedimentologie, de paleogeografie en de structureel-geologische evolutie van een binnenzee in Noord Hongarije (Noord Hongaarse Bekken) en in het bijzonder van de vroeg Miocene ***Pétervására Zandsteen***. In **hoofdstuk 1** worden de vraagstelling en de werkwijze uiteengezet.

Hoofdstuk 2 geeft een overzicht van het paleogeografische, stratigrafische en structureel-geologische kader, waarin de ontwikkeling van het Noord Hongaarse Bekken geplaatst wordt. Met name de vroeg Miocene ontwikkeling van het intra-Carpathische gebied heeft veel auteurs bezig gehouden. In het vroeg Mioceen vond een grootschalige tectonische reorganisatie van microplaten plaats. Dit ging gepaard aan een sterke tectonische activiteit van grotere en kleinere structurele elementen. Hierdoor werden de bekken-evolutie, de sedimentaire vulling en ook de preservatie van de sedimenten sterk beïnvloed. Tijdens het vroeg Mioceen maakte het bekken deel uit van de Paratethys, een grote binnenzee die zich uitstrekte tot voor-Azië. In de fossiele afzettingen van deze binnenzee kunnen over afstanden van duizenden kilometers biologische en sedimentologische gebeurtenissen worden vervolgd en gecorreleerd aan grootschalige tectonische bewegingen. Voorts worden in dit hoofdstuk de Paleogene afzettingen in het Noord Hongaarse Bekken beschreven en worden de structureel-geologische processen die het karakter van deze afzettingen hebben beïnvloed besproken.

In **hoofdstuk 3** wordt de ***Pétervására Zandsteen*** in detail beschreven en worden gegevens uit het veld en sedimentologische parameters gepresenteerd. Vier facies eenheden worden onderscheiden. Deze worden gekenmerkt door hun verbreiding in ruimte en tijd, door de aard van het sediment en door de hierin voorkomende sedimentaire structuren. Nabij de vroeg Miocene kustlijn bestaat de sedimentaire opeenvolging uit afzettingen van grootschalige zandgolven met daarin conglomeraatlobben. Zandgolven waren tot 12 meter hoog en werden gevormd door sterke kustparalelle getijdestromingen. Meer zeewaarts nam de hoogte van de zandgolven af tot decimeter-hoge megaribbels. Bekkenwaarts gaan deze geleidelijk over in siltieten die in een diep-neritisch milieu zijn afgezet.

Hoofdstuk 4 gaat in op de ruimtelijke en genetische relatie tussen de ***Pétervására Zandsteen*** en het ***Darnó Conglomeraat***. Het ***Darnó Conglomeraat*** komt voor langs de Darnó breukzône en is een grofklastische eenheid met veel ofioliet fragmenten. Beide eenheden werden min of meer in de zelfde periode gevormd, maar in de literatuur zijn ze altijd als aparte eenheden behandeld. Uit het onderzoek blijkt dat beide eenheden nauw aan elkaar zijn gerelateerd. De Darnó conglomeraten werden afgezet op kleine, deels onderzeese delta's, bepaald door een actieve breukzône. De conglomeraten gaan zeewaarts over in de Pétervására Zandsteen, hetgeen wordt bevestigd door de sediment petrografische samenstelling van de sedimenten. Laterale breukbewegingen in het vroeg Mioceen hebben ertoe geleid dat de brongebieden van het sediment en de afzettingen in het bekken zijdelings ten opzichte van elkaar zijn verschoven. De toename naar het noorden van de hoeveelheid aan ofiolieten gerelateerde zware mineralen en ofioliet-pebbles wijst op een links-laterale

verplaatsing langs de Darnó breukzône. Conglomeraat-lobben die daarbij werden losgekoppeld van de fan-delta, werden verspoeld door de sterke getijdestromingen en als aparte lagen afgezet in een opeenvolging van afzettingen van zandgolven.

Hoofdstuk 5 begint met een overzicht van de eigenschappen van het astronomische getij en van de daaraan gerelateerde periodiciteiten die in getijde-beïnvloede afzettingen kunnen worden herkend (eb-vloed bewegingen, doodtij-springtij cycli, halfjaarlijkse en jaarlijkse cycli). Cycli die zijn vastgelegd in de zanden van de Pétervására Zandsteen worden besproken. Deze laten zien dat in het vroeg Mioceen het Noord Hongaarse Bekken werd gedomineerd door een getijde-systeem met twee maal per dag eb en vloed. Doodtij-springtij cycli en halfjaarlijkse en jaarlijkse getijde-cycli worden eveneens herkend. Analyse van de verzamelde gegevens wijst op een paleo-getijde amplitude van ca. 4 meter en een stroomsterkte tijdens springtij van 0,7 m/sec tot 0,9 m/sec. Berekeningen van de snelheid waarmee de zandgolven migreerden, op basis van korte zowel als langere opeenvolgingen, wijzen op een snelheid van 0,6 m tot 0,8 meter per 14 dagen en 10 tot 20 meter per jaar.

Hoofdstuk 6. De getijde-beïnvloede afzettingen in het Noord Hongaarse Bekken wijzen op een verbinding met de open oceaan, die in het vroeg Mioceen op grote afstand lag. Tussen het Noord Hongaarse Bekken en de aangrenzende Paratethys bekkens zijn geen vroeg Miocene sedimenten bewaard gebleven. Daarom is niet zonder meer duidelijk welke route de getijdegolven hebben gevolgd. Afweging van de verschillende mogelijkheden leidt tot de conclusie dat het oostelijke Slowaakse Bekken de verbinding moet hebben gevormd tussen het Noord Hongaarse Bekken en de open oceaan in het oosten. De getijde-beïnvloede afzettingen in het Noord Hongaarse Bekken, als ook in bekkens van ongeveer dezelfde ouderdom in het Alpine gebied, bieden de mogelijkheid om de opening en sluiting van zeestraten in de Paratethys te reconstrueren. Voorts laten ze zien dat getijde-beïnvloede afzettingen een uitstekend hulpmiddel kunnen zijn bij paleogeografische reconstructies.

In **hoofdstuk 7** wordt ingegaan op de voorwaarden die nodig waren voor de vorming van de getijde-beïnvloede afzettingen in het Noord Hongaarse Bekken. Een verbinding met de oceaan is één van de voorwaarden voor het optreden van getijdebewegingen. Daarnaast moet het bekken eigenschappen hebben gehad om een vergroting van de getijde-amplitude tot 4 meter mogelijk te maken. Door een semi-kwantitatieve analyse van de morfologie en de hydrodynamische condities van het Noord Hongaarse Bekken is bepaald welke factoren tot de vergroting van de getijde-amplitude en de ontwikkeling van een asymmetrisch getijde-systeem hebben geleid. De vergroting van de getijdeamplitude kon plaats vinden doordat de diepte en de lengte van het Noord Hongaarse Bekken en de aansluitende zeestraat door Oost Slowakije voldeden aan de voorwaarden die nodig zijn voor resonantie van de getijdebeweging. Tegengestelde sediment-transportrichtingen aan de oostzijde en de westzijde van het bekken worden geïnterpreteerd als het gevolg van een asymmetrie van het getij als gevolg van interactie van de getijde-bewegingen met de bodem van het bekken. Vergelijkbare studies in andere getijdebekkens kunnen een belangrijk hulpmiddel zijn bij de reconstructie van de bekkenontwikkeling.

In **hoofdstuk 8** worden de sedimentologische gegevens gebruikt voor een structureel-geologische reconstructie van de evolutie van het Noord Hongaarse Bekken. De veranderende afzettingsomstandigheden tijdens de vorming van de Pétervására Zandsteen ondersteunen het tektonische model zoals recent voorgesteld door Tari et al. (1993). De ontwikkeling van het ondiep marine bekken en de invloed van de tektonische opheffing in de loop van het vroeg Mioceen worden besproken. Het tweede deel van dit hoofdstuk bespreekt de ontwikkeling van het getijde-beïnvloede systeem in het bekken in relatie tot de wereldwijde daling van de zeespiegel in die tijd. Deze combinatie is uitzonderlijk; getijde-beïnvloede afzettingen zoals die in het Noord Hongaarse Bekken worden in de regel vooral gevonden in afzettingen die zijn gevormd in periodes met een stijging van de zeespiegel.

CHAPTER 1
GENERAL INTRODUCTION

The Lower Miocene shallow marine Pétervására Sandstone was deposited during a particularly interesting, geodynamically active and palaeogeographically unsteady interval in the history of the intra-Carpathian region. While the Mesozoic development of this area was strongly related to the evolution of the Tethys Ocean and to Alpine-Dinaridic tectonic processes, the Neogene of the intra-Carpathian area corresponded to the extension of the Pannonian Basin (Horváth, 1986). In between, the Palaeogene - early Miocene, was a period of tectonic reorganization in connection with the evolution of the Carpathians, when the present day configuration of tectonic megaunits of various origin was formed via large-scale plate movements (Csontos et al., 1992).

The tectonic processes in the Alpine-Carpathian-Dinaridic belt had a profound effect on the development of the intra-Carpathian Palaeogene sedimentary basins, e.g. the Hungarian Palaeogene Basin (Báldi & Báldi-Beke, 1985). In the Hungarian literature "Palaeogene" is often used in a wider sense. Deposits and events up to Middle Burdigalian are commonly regarded as Palaeogene, because the related sedimentary cycle terminated in the Early Miocene. The study of the basin-fill deposits, e.g. the Pétervására Sandstone (Báldi, 1983), thus the understanding of the evolution of the Hungarian Palaeogene Basin and of the ongoing palaeogeographic changes can greatly contribute to the reconstruction of the large-scale tectonic puzzle.

Chronostratigraphical and palaeoecological studies were carried out for most of the Hungarian Palaeogene basin-fill deposits (Báldi, 1973; Báldi-Beke, 1977; Báldi-Beke, 1984; Báldi, 1986; Báldi & Báldi-Beke, 1985; Nagymarosy & Báldi-Beke, 1988; Báldi-Beke & Báldi, 1990; and many others). Sedimentological and/or structural investigations of the Palaeogene deposits, however, started only in the late 80's.

The poorly understood Pétervására Sandstone, which covers a large area in northern Hungary, is almost barren of fossils and therefore little attention has been paid to it by palaeontologists. Nevertheless, a variety of sedimentary structures occurs, appealing for a sedimentological approach. The chronostratigraphic position of this formation and its shallow marine depositional environment were discussed by Báldi (1983), but details of the evolution of the depositional system, as well as its relation to other lower Miocene deposits in the neighbourhood were not clear. The tide-influenced sedimentary environment of the Pétervására Sandstone was proposed by Tari et al. (1989). This raised a number of questions about the palaeogeographic position of the Hungarian Palaeogene Basin, and its marine connections during the early Miocene.

The objectives of this thesis are:

- to determine the main sedimentary units, the dominant depositional processes, and to reconstruct the depositional environment of the Lower Miocene Pétervására Sandstone (Chapter 3),
- to show the pattern of sediment dispersal in connection with the potential effects of small-scale syntectonic processes (Chapter 4),
- to document and interpret the tide-influenced system (Chapter 5),
- to understand its oceanographical background and the reasons for the development of the tidal influence (Chapter 7),
- to analyze the potential palaeogeographic connections implied by the tide-influenced deposits in the basin, with other Early Miocene basins in the Central Paratethys region (Chapter 6),
- to pinpoint stratigraphic markers and to establish chronostratigraphic correlations within the formation and also with coeval deposits (Chapter 4 and 8),
- to determine the main controlling factors in the development of the depositional environment of the Pétervására Sandstone, and to distinguish signals of tectonics and eustasy (Chapter 8).

The structure of the thesis

The thesis consists of four major units: chapter 2, 3-4, 5-7 and 8. In Chapter 2 a summary of the regional geology is given for readers not familiar with Paratethyan stratigraphy and the intra-Carpathian geology. Only those aspects of the regional geology are discussed, which are necessary to understand the problems and the approach presented in the following chapters. The subsequent chapters contain the new results. Most of these are separate publications (mainly in press or under review), which discuss various aspects of the development of the Pétervására Sandstone. Chapter 3 and 4 contain detailed descriptions of the studied sedimentary sequences. In Chapter 3 field data are presented and interpreted in terms of processes and the depositional environment is analyzed. Chapter 4 deals with the sedimentology and the potential source of the conglomerates found in the tide-influenced environment. In addition the relation between the conglomerates of the Pétervására Sandstone and the solitary Darnó Conglomerate is discussed. In the following three chapters (5 to 7) the main accent is put on the tide-influenced deposits. The various tidal cyclicities which can be recognized in the sedimentary record of the Pétervására Sandstone are discussed in Chapter 5. In Chapter 6 potential palaeogeographic connections, which must have existed in order to allow the transfer of tidal energy to northern Hungary are clarified. In addition, an outlook to different tide-influenced settings in the region is given. In Chapter 7 oceanographic conditions and ways of amplification of tidal motions are discussed, which must have occurred in order to produce the tide-influenced sediments in the North Hungarian Bay. Finally in Chapter 8 a dynamic picture of the early Miocene sedimentary system is given in relation with the evolution of the basin, large-scale tectonics and eustatic influences.

CHAPTER 2

GEOLOGICAL SETTING

Stratigraphy and palaeogeography: the Paratethys

From the Oligocene onwards a separate branch of the Tethys developed in the Alpine-Carpathian-Dinarid region (Báldi, 1980). A chain of basins of various tectonic origin was covered by the same mass of water, sharing the same aquatic biotas. This sea, which occupied the area of the present Rhone valley, the Alpine-Carpathian foredeep, the "intramontane" basins from Austria to the Ukraine and the Ponto-Caspian area, is called Paratethys (Fig. 2.1). Its central part, from Bavaria to the Eastern Carpathians, is the Central Paratethys.

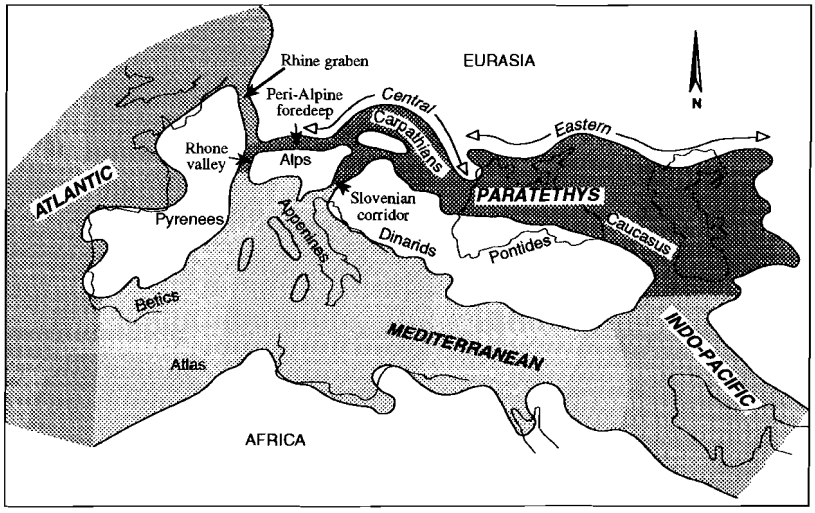

Fig. 2.1. The Paratethys as a bioprovince had connections towards the Atlantic (North Sea), the Mediterranean and the Indo-Pacific regions (from Rögl & Steininger, 1983). The map shows where Paratethys deposits are presently found. The actual shape of the Paratethys changed through time (see discussion below and in Chapter 6).

The Paratethys had various connections to open marine areas, but these seaways were blocked from time to time, giving way to the development of endemic fauna and flora. Endemism usually was characterised by booms of monospecific nanno-flora assemblages (Nagymarosy, 1992) and rapid expansions of specific mollusc communities. Isolation and reopening of the Paratethys occurred repeatedly. Because of the endemic events, local bio- and chronostratigraphical correlation schemes have been developed, which have been calibrated with the standard stratigraphic colums (Fig. 2.2).

For the regional chronostratigraphy in the Central Paratethys the reader is referred to Báldi & Senes (1975) and Steininger & Senes (1971). A recent compilation of Paratethyan bio- and chronostratigraphy is given by Steininger et al., (1990). Parts of his chart are under discussion, particularly the isochroneity of the Aquitanian/Burdigalian boundary with the Egerian/Eggenburgian boundary. It is agreed that the base of the Burdigalian is within the NN2 nannoplankton zone (cf. Steininger et al., 1990 with Haq, 1991, Fig. 2.2b). The boundary of the NN1/NN2 nannoplankton zones, however, is considered to be equivalent to the Egerian/Eggenburgian boundary (Fig. 2.2; Nagymarosy & Báldi-Beke, 1988; Nagymarosy & Müller, 1988). If these above statements are true, then the Egerian/Eggenburgian boundary must be older then the Aquitanian/Burdigalian boundary. The biochronological framework used in this study is based on Nagymarosy's data on calcareous nannofossils (Nagymarosy & Báldi-Beke, 1988; Nagymarosy, 1990b and many other unpublished data), therefore the division below (Fig. 2.2) has been used.

It should be kept in mind that different authors use different groups of fossils and different chronostratigraphic subdivisions when proposing an age for various deposits and events. This constrains precise correlations.

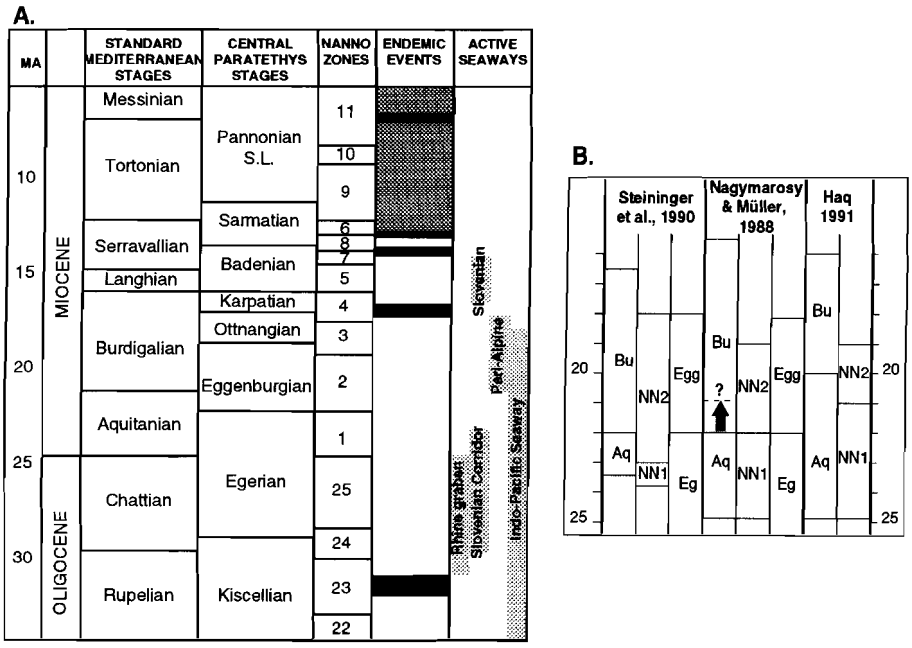

Fig. 2.2. A. Correlation of standard Mediterranean and Central Paratethys stages (sligthly modified after Nagymarosy & Müller, 1988). Periods of isolation, characterized by endemic faunal communities (black bars), and active seaways are shown (after Nagymarosy, 1990a). B. The insert shows the different opinions regarding early Miocene chronostratigraphic boundaries. Note that the difference between Haq's (1991) and the modified stratigraphic subdivision by Nagymarosy & Müller (1988) is only minor.

Evolution of the Central Paratethys

The first isolation of the Paratethys occurred during the Early Oligocene (Fig. 2.2; NP23 nannofossil biozone, "Cardium lipoldi event"; Báldi, 1980; Rögl & Steininger, 1984; Nagymarosy, 1990a), most probably because of a reorganization of the Alpine thrusts (Nagymarosy, 1990a). Seaways, towards the Atlantic via the North Sea (Rhine graben) and the Mediterranean (Rhone valley, Slovenian corridor) had been blocked (Fig. 2.1). Reopening through the Slovenian corridor and the Rhine-graben took place during the NP24 nannozone (Fig. 2.2). This reopening resulted in an influx of boreal biota (Rögl & Steininger, 1984). Afterwards, around the Oligocene-Miocene boundary, both seaways were closed again (Nagymarosy, 1990a). This has been inferred e.g. from the distribution of mammals (Steininger et al., 1985).

During the early Miocene (NN2 nannoplankton zone) radiations of cosmopolitan giant bivalves and Mediterranean fauna elements were observed in the Paratethyan aquatic biota. This change has been ascribed to a global sea-level rise and a warming of the climate (Báldi, 1980; Steininger et al., 1985; Haq, 1991). In the Burdigalian the sea flooded the peri-Alpine depression again from the Rhone valley and the Central Paratethys (Fig. 2.1; Rögl & Steininger, 1983; Rögl & Steininger, 1984; Nagymarosy, 1990a). This peri-Alpine seaway, however, was short-lived. The seaway become filled up to sea level, and a second separation occurred in the late Early Miocene (Fig. 2.2; NN4 nannoplankton zone, "Rzehakia beds"; Rögl & Steininger, 1984; Nagymarosy, 1990a). Around the mid-Early Miocene the connection between the Eastern Mediterranean and the Eastern Paratethys had been also closed by orogenic events (Nagymarosy, 1990a). Before the Middle Miocene the Mediterranean influence was restored by a connection through the Slovenian corridor again (Fig. 2.1). The endemic "Konka fauna" arrived from the enclosed Eastern Paratethys (Kókay, 1985). The third and final separation occurred in the late Middle Miocene (Sarmatian event; Fig. 2.2). Since then the Central Paratethys has been completely separated from other seas and gradually evolved into the Pannonian Lake (Rögl & Steininger, 1984).

Plate tectonic evolution of the intra-Carpathian area

Tectonic units within the intra-Carpathian area

In the intra-Carpathian area two different tectonic domains were recognized through paleogeographical studies of Mesozoic rocks (Géczy, 1973; Kázmér & Kovács, 1985). The Mid-Hungarian Line separates the North and South Pannonian Units (Fig. 2.3a; Balla, 1984). The South Pannonian unit consists of the Tisza and Dacide subunits, which are thought to have merged before the Tertiary, and thus behaved as one microplate during the Cenozoic (Fig. 2.3a).

The Mesozoic depositional history of the North-Pannonian-West Carpathian unit is firmly tied to that of the Tethys Ocean, as is shown by the close resemblance of sedimentary sequences in this unit with sequences in the Alpine and Dinarid realms. While the North Pannonian unit has

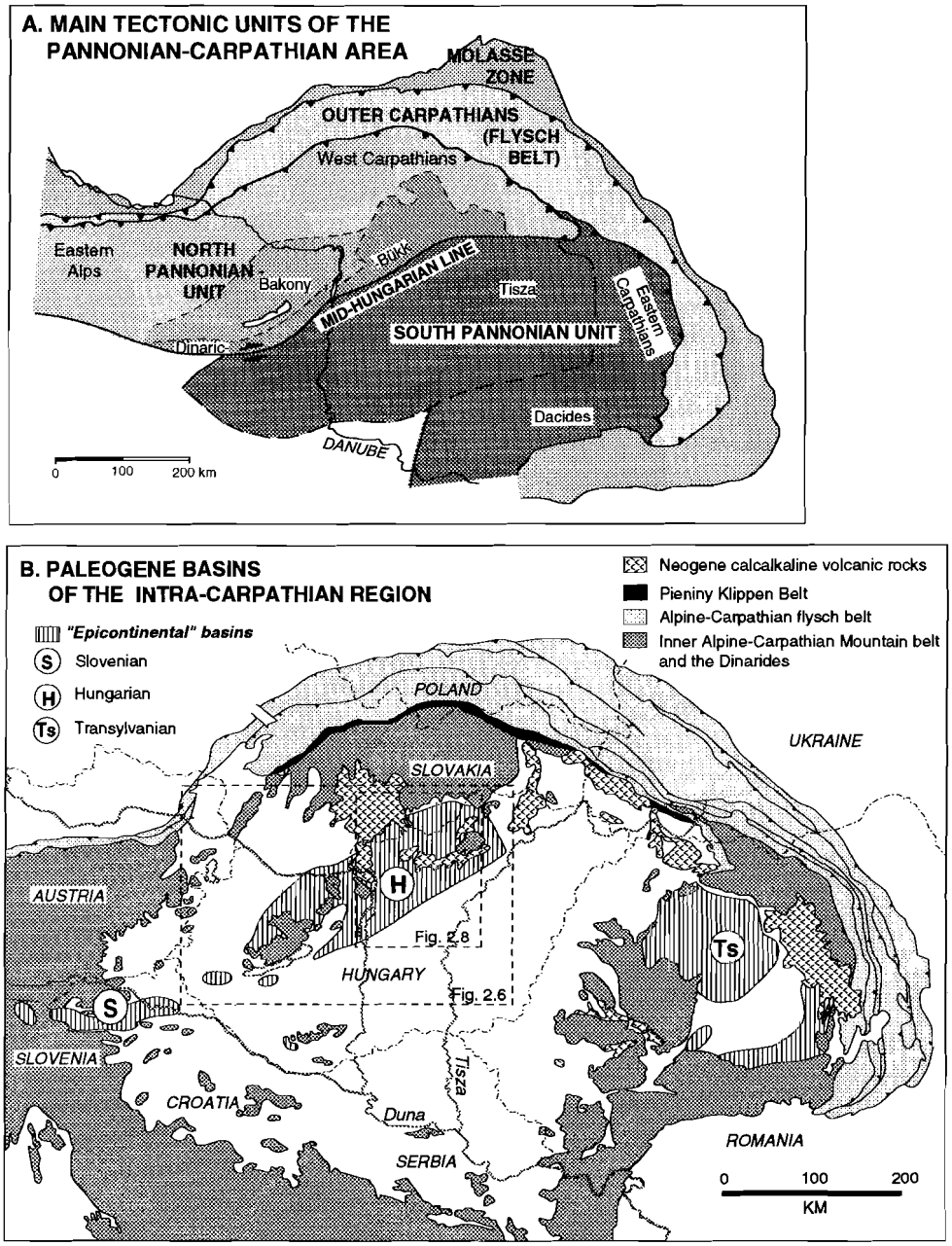

Fig. 2.3. A. The main Cenozoic tectonic units of the intra-Carpathian area (after Csontos et al., 1992) with B. the "epicontinental" Palaeogene basin fragments (by courtesy of Tari). The Pieniny Klippen belt separates the intra-Carpathian area from the Outer Carpathian flysch nappes. Subunits of the North-Pannonian block are also shown.

"African" affinity, the Tisza Unit, which is located south of the Mid-Hungarian Line at present, was formerly situated along the northern margin of the Tethys, close to stable Europe. During the Tertiary the North and South Pannonian Units were juxtaposed by large-scale dextral strike-slip displacement (Balla, 1984), and they provided a united basement for subsidence of the Neogene Pannonian Basin (Horváth & Royden, 1981) (Fig. 2.3). One of the main goals of recent tectonic studies is to determine the mechanism and timing of these plate movements.

Forces behind the movements of "microplates"

Kázmér (1984) and Kázmér & Kovács (1985) suggested that, due to collision of Europe and Africa, the "continental escape" of Bakony subunit, a central part of North Pannonian Unit, occurred during the middle Oligocene (Fig. 2.3a). Balla (1984) presumed that the whole North-Pannonian-West-Carpathian Unit (Dinaridic-Bükk, Bakony and Inner Western Carparthian subunits) escaped. Based on palaeomagnetic data (cf. Márton et al., 1992) rotation of large rigid blocks was supposed (Balla 1987a, 1988). In contrast, Tari (1991), Marko et al., (1991) and Csontos et al., (1992) suggested that, as a consequence of wrench tectonics, only rotation of minor blocks occurred. The model of "escape tectonics" was elaborated further by Csontos et al., (1992), who combined the effects of the two possible driving mechanism: 1./ compression between the colliding continents triggered the eastward escape of the North Pannonian Unit, while 2./ subduction of the Magura Ocean promoted the escape and provided the space needed (Fig. 2.4).

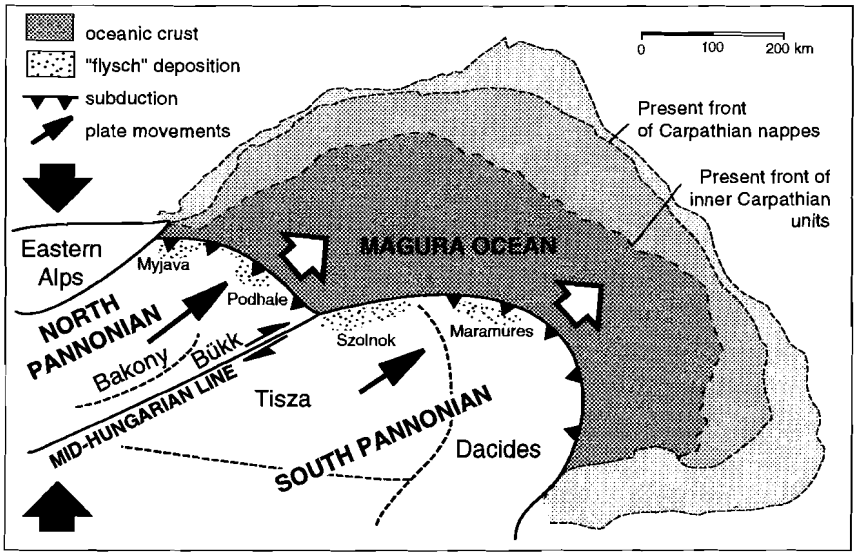

Fig. 2.4. Driving mechanism of large-scale block movements within the intra-Carpathian area during the late Eocene to Oligocene (after Csontos et al., 1992). A pair of big black arrows and big empty arrows indicate the "push" of collision and the "suck" of subduction respectively. The oceanic crust of thre Magura had mainly been consumed by the Miocene.

Late Eocene compressional tectonics in the Buda area was interpreted in connection with the early stage of escape movement of the North Pannonian Unit (cf. Fodor et al., 1992). Displacement of Palaeogene basins (Fig. 2.3b) along the Dinaric-Bükk shear zone reflects intense plate motions which may have started during the Late Oligocene (Csontos et al., 1992). The final event of the "escape" is dated by overthrusting of the North Pannonian Unit on top of Lower Miocene Wildflysch in the Maramures (Botiza) area (Fig. 2.4). Therefore Csontos et al., (1992) supposed that the escape of the North Pannonian unit occured mainly during Late Oligocene - Early Miocene.

The escape of the North Pannonian block led to an oblique collision with the European continent. Thus the consumption of the oceanic crust, and also the thrusting of the innermost flysch units (e.g., Magura flysch) took place gradually from west to east (Fig. 2.3b; Jiricek, 1979). A second phase of thrusting of the outer flysch nappes (Silesian, Moldavian; Fig. 2.3b) onto the Carpathian molasse foreland from the Middle Miocene onwards was associated with the extension of the Pannonian Basin in the intra-Carpathian area (Horváth, 1986).

A detailed timing and reconstruction of large-scale tectonic events cannot be carried out without a proper knowledge and understanding of the Palaeogene - Early Miocene sedimentary record, found mainly in the Hungarian Palaeogene basin.

The Hungarian Palaeogene Basin

Extension or compression? About the genesis of the basin

Subsidence and deposition in the Hungarian Palaeogene Basin (Figs. 2.3b & 2.6) started in the Middle Eocene and continued until the early Miocene (Báldi & Báldi-Beke, 1985; the most recent review of basin-fill deposits and various models for basin evolution is given by Tari et al., 1993). Meanwhile there was a clear shift of the depocenter from the southwest to the northeast (Fig. 2.5). Therefore the Hungarian Palaeogene basin is commonly regarded as a series of small elongated Basins, each produced by the same mechanism, discussed below. The history of each asymmetrical basin "segment" shows a two-phase stratigraphic fill: after initial shallow marine conditions a very rapid subsidence followed, that introduced a longer period of bathyal deposition (500 m - 1000 m). In the second phase gradual shallowing upward occurred and finally subaerial erosion of former basin-fill deposits took place (cf. Báldi & Báldi-Beke, 1985).

Báldi & Báldi-Beke (1985) and Royden & Báldi (1988) regarded the Hungarian Palaeogene Basin as a transtensional basin, resulting from the activity of large transform faults (e.g., the Mid-Hungarian Line, Fig. 2.6). Their interpretation emphasizes the role of minor faults, which bounded the individual basins. However, the presence of these faults is not always clear. They cannot be considered uniformly as extension-related synsedimentary faults (e.g., Buda Line first defined by Báldi & Nagymarosy, 1976, reinterpreted by Fodor et al., 1992).

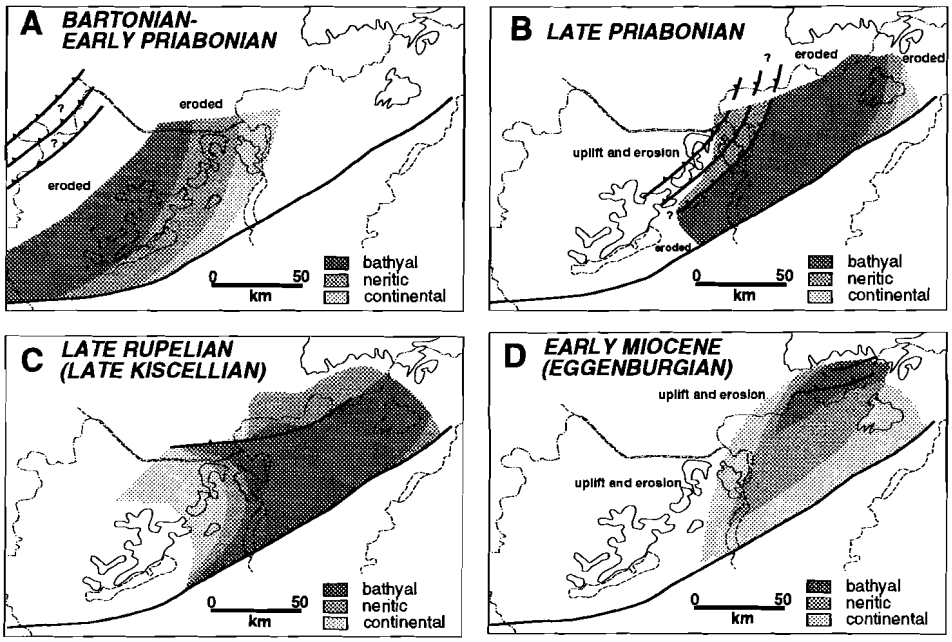

Fig. 2.5. Eastward migration of the depocenter in the Hungarian Palaeogene Basin during the Eocene - early Miocene period (after Tari et al., 1993).

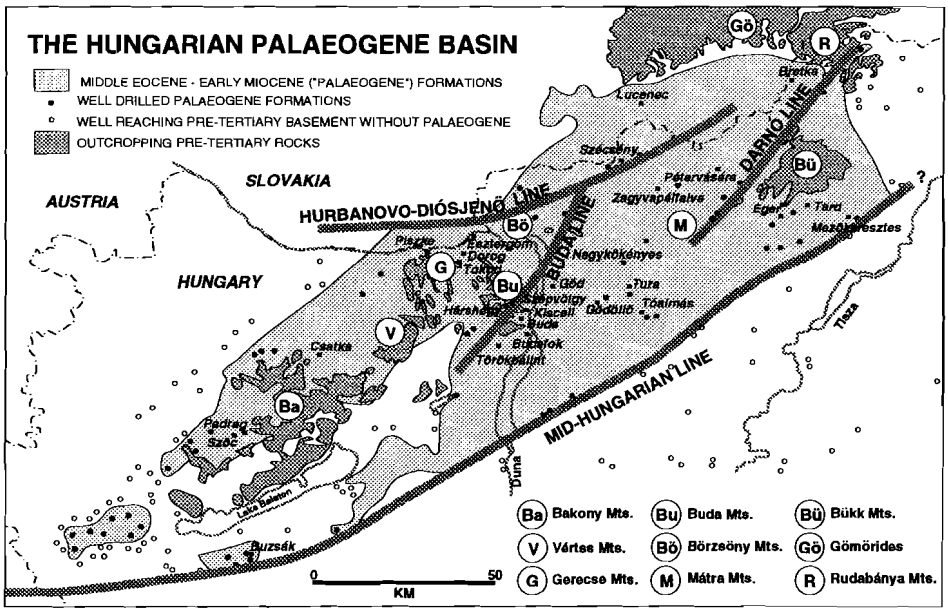

Fig. 2.6. Map of the Hungarian Palaeogene Basin showing the main tectonic lines and names of geographic locations mentioned in the text below (by courtesy of Tari).

Late Eocene - Early Oligocene synsedimentary phenomena, - slope-related mass flow deposits, slides, slumps, creeps and debrites -, in the Buda Mountains (Fig. 2.6) reveal the presence of thrust-folds and evidence of compressional tectonics (Fodor et al., 1992). This Eocene-Oligocene compressional tectonics produced the Buda Line (Figs. 2.5b & 2.6), which was interpreted as the highest of the anticlines, that formed above blind thrusts (Fodor et al., 1992). The Buda Line is an important structural line, which was inherited from the Oligocene to the early Miocene as a topographic height, and formed the western boundary of the youngest, northeasternmost Palaeogene basin "segment". The transpressional character of the Darnó line (Fig. 2.6), that bounded the Palaeogene basin in the east during the Early Miocene, is also well known (Schréter, 1942a).

The presence of major compressional structural elements during the Palaeogene in combination with indications of strike-slip movements led Fodor et al., (1992) to ascribe the evolution of the Hungarian Palaeogene Basin in terms of transpression, in contrast to models suggesting a transtensional origin (Báldi & Báldi-Beke, 1985).

Tari et al. (1993) interpreted the compression as a major, Eocene-Oligocene structural feature. It was considered separately from the strike-slip movements dated to be of late Early Miocene. The two-phase stratigraphic fill and the asymmetry of the Hungarian Palaeogene Basins described above was interpreted to be typical of a retroarc flexural basin (Tari, 1992; Tari et al., 1993). Such a depression is likely to develop, when backthrusts are formed behind the main thrust-fold belt. Although his model explains many, formerly poorly understood features, the presence of south-verging backthrusts north of the Hungarian Palaeogene basin and south of the main northward-verging Carpathian thrusts (cf. Fig. 2.3b) has not been identified as yet (for more discussion see Chapter 8).

The North Hungarian Palaeogene Basin

The sedimentary record of the youngest, northeasternmost Palaeogene basin "segment", between the Buda and Darnó lines (Fig. 2.6), is the most complete one. This part of the basin is traditionally called North Hungarian Palaeogene Basin, although deposition occurred from Late Eocene to early Miocene (Fig. 2.7). During this interval the main depocenter was between the Buda and Darnó lines. Some marginal shallow marine deposits are found further westward and eastward, but their thickness is not significant in comparison with the main basin fill, which exceeds 2500 m.

The southernmost part of the basin was displaced along the Dinaric-Bükk shear zone (Csontos et al., 1992) and now forms the Slovenian Palaeogene basin (Fig. 2.3b). The correspondence between the two basin segments is fairly good for the Oligocene deposits (Nagymarosy, 1990b). However, during the Early Miocene sedimentation was entirely confined to the north Hungarian area. The southern limit of marine deposition shifted northward, away from the Mid-Hungarian Line (Fig. 2.5d).

The activity of the Buda Line, as a synsedimentary anticline, was intensive in the Late Eocene - Early Oligocene, and it diminished during the Late Oligocene (Fodor et al., 1992). During the Oligocene-early Miocene the Buda Line functioned as a facies boundary (Báldi & Nagymarosy, 1976), because it formed a gentle topographic height.

Sedimentation from late Eocene to early Miocene (Egerian)

In the North Hungarian Palaeogene Basin marine sedimentation started with breccias and conglomerates in the Buda area (Magyari, 1991) and continued with shallow marine nummulitic limestones all over the basin (Fig. 2.7). These are overlain by the bathyal Buda Marl, rich in calcareous turbiditic intercalations (Fig. 2.7). It is followed by an anoxic shale, the Lower Oligocene Tard Clay, which represents the first isolation of the Paratethys (cf. Fig. 2.2, Báldi, 1980). Upward in the section sandy turbidites are intercalated in the Tard Clay, and the top of the formation is rich in fish remains.

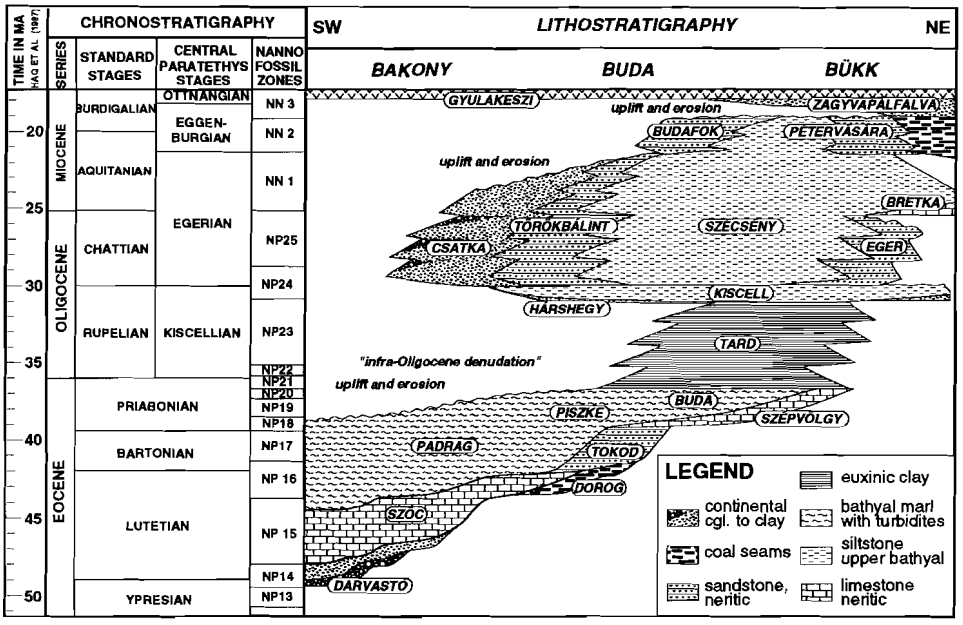

Fig. 2.7. Lithostratigraphy of the Hungarian Palaeogene Basin (after Tari et al., 1993). In addition to the main lithologies, depositional depths of the formations are also indicated. Lateral relationships are shown properly, but the vertical ones may differ between southern and northern localities (by courtesy of Tari).

The overlying Kiscell Clay is a widespread open-marine deposit, which was formed after the reopening of the Paratethys, when ventilated bottom conditions were restored. Contemporaneous shallow marine sandy sedimentation began along the basin margins (Hárshegy

Sandstone), which shortly later was covered by the Kiscell Clay (Fig. 2.7). During deposition of the Kiscell Clay, the basin rapidly subsided and became highly underfilled. Based on studies of foraminifera and molluscs the depositional depth was estimated to have been at minimum 400-800 m (Báldi, 1983), but more recent estimates give values of up to 1000 m (Báldi & Horváth, 1990 personal communication; Báldi & Nagy-Gellai, 1990).

Around the mid-Oligocene, the depositional depth in the North Hungarian Palaeogene basin decreased significantly. In central areas deposition of shallow bathyal siltstones started (Fig. 2.7, Szécsény Schlier; Báldi, 1983). Along the eastern side of the basin lobes of redeposited conglomerates and canyon fills were formed. These are overlain by the upward coarsening and shallowing Eger Clay and Sand (Fig. 2.7), representing various nearshore environments (Tari & Sztanó, 1992).

Along the western side of the basin the Kiscell Clay is overlain by the upwards shallowing Törökbálint Sandstone (Fig. 2.7). In the Buda area large load casts and turbidite beds are found at the base of this formation (Fodor, 1992 personal communication), which indicates intense redeposition around the mid-Oligocene. South-east of the Buda Line (Cinkota-Göd area, Fig. 2.6) the Törökbálint Sandstone exceeds 300 m in thickness. There it is represented by silts with sandy intercalations, indicating a trough at bathyal depths (Báldi, 1983).

At the start of the Miocene, a further northeastward shift of the depocenter occurred. Thus the sea flooded basement rocks, formerly exposed northeast of the basin, and a basal Miogypsina limestone (Bretka Fm) of Aquitanian age was deposited (Báldi, 1986). As the rise of relative sea level and the transgression continued, deposition of the Szécsény Schlier extended further towards the northeast (Fig. 2.8). This transgression also affected the western margin, but the presumably active Buda Anticline prevented inundation of large areas.

Eggenburgian in the northern part of the basin

The Szécsény Schlier Formation represents the deepest depositional environment in the North Hungarian Palaeogene basin from Egerian to Eggenburgian. Schlier is a local name for mollusc-bearing sandy, clayey siltstones, formed in the deep-neritic shallow-bathyal realm. Depositional depth of the Szécsény Schlier was estimated to have been at minimum 60 m, but in general between 130-300 m (cf. Báldi, 1986). The thickness of these bathyal siltstones exceeds 1000 m in the central part of the basin (Fig. 2.8). It extends further to north and is called Lucenec Schlier in South Slovakia (cf. Vass et al., 1988). Except for the deepest parts of the basin, deposition of the schlier ended abruptly when the deposition of sandy formations started in the Eggenburgian (Fig. 2.8). It is overlain by the Eggenburgian shallow marine Pétervására (Glauconitic) Sandstone and the littoral Budafok Sand Formations on the eastern and western sides of the basin, respectively (Figs. 2.7 & 2.8).

Sediments deposited during the Eggenburgian are entirely confined to the area defined by the Buda and the Darnó Lines (Fig. 2.8). The present distribution of Eggenburgian deposits at the

surface is well known. Towards the south only some sporadical occurrences are known from boreholes.

The littoral Budafok Sand (Fig. 2.8) overlying the Szécsény Schlier or disconformably the Törökbálint Sandstone (Fig. 2.7) was restricted to a narrow patch along the Buda Line. The Pétervására Sandstone was deposited along the eastern margin of the basin (Fig. 2.8). Both shallow water formations prograded towards the center of the basin and finally, before the end of the Eggenburgian, the basin had been filled up to sea level. The top of the sequence is formed by the terrestrial Zagyvapálfalva Formation, which is represented by coastal plain to alluvial plain facies. The termination of Palaeogene-early Miocene deposition was marked by volcanic eruptions producing a 40-60 m thick cover of rhyolite tuff. Culmination of volcanism was dated radiometrically at 19.5±1.6 million year (Balogh-Kadosa in Hámor, 1985).

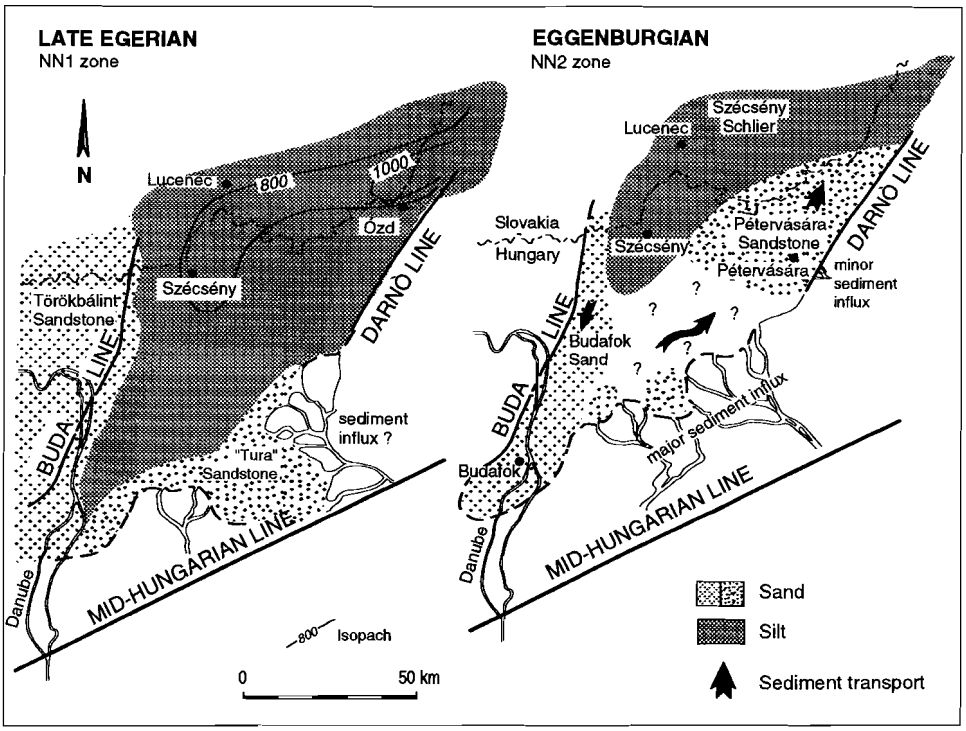

Fig. 2.8. The palaeogeography of the North Hungarian Bay during the Early Miocene. The reconstruction is based on subsurface data from the southern part of the bay (Nagymarosy unpublished, 1988; Lakatos et al., 1991) and on field data from the northern part of the bay (Báldi, 1986; Sztanó & Tari, 1993; Chapter 3). During the Late Egerian the bathyal Szécsény Schlier was deposited in a depression determined by the Buda and Darnó Lines. The bay became significantly shallower during the Eggenburgian and most of the area was filled by shallow neritic sandstones. Isopachs of the Szécsény Schlier clearly show the deepest, still subsiding, part of the bay during the Eggenburgian.

Egerian to Eggenburgian sedimentation in the southern part of the basin

In the southern part of the north Hungarian Palaeogene Basin a large number of boreholes has penetrated a succession of deposits, which differs from the above described sequence (Nagymarosy unpublished, 1988; Lakatos et al., 1991). In these boreholes significant disconformities are observed and some Palaeogene formations are completely missing.

Close to the southeastern margin of the basin - in the Gödöllő-Tóalmás-Tura-Jászberény area (Fig. 2.6 and "Tura" Sandstone on Fig. 2.8) - the bathyal Kiscell Clay (NP24) is disconformably overlain by sandstones of Egerien age (NP25/NN1; chronostratigraphy based on unpublished nannoplankton investigations by Nagymarosy and Báldi-Beke). The thickness of this sandstone is fairly variable, and ranges from 40 m to 400 m. In some of the boreholes the uppermost 100 m of the sandstone is already Eggenburgian (NN2/3) in age. These sandstones of Egerian-Eggenburgian age were identified as Pétervására Sandstone (Lakatos et al., 1991). This was based mainly on lithological analogy (Nagymarosy pers. comm.). No study of facies or depositional environment of the sandstone has been carried out, because the samples were small pieces of cores only. The Egerian-Eggenburgian sandstones bodies in the south and the Pétervására Sandstone in the north seem to be separate lithosomes. Unless there is enough evidence available to show any direct relationship, the Egerian-Eggenburgian sands along the southestern basin margin are treated as different deposits, and therefore not discussed in this thesis.

The unconformity between Kiscellien and Egerien deposits is obvious in many boreholes. Further to the north, around Nagykökényes, the Lower Miocene deposits (Upper Egerian to Ottnangian) are completely missing. The late Early Miocene (Karpatian) sediments disconformably overlay the Lower Egerian Szécsény Schlier. This indicates a significant period of erosion during Eggenburgian and/or Ottnangian times.

Summary of the Hungarian Palaeogene Basin

Nagymarosy (1990b) and Csontos et al., (1992) suggested that the Palaeogene basins of Slovenia and Hungary were part of the same basin, dissected by wrench-faults. Litho- and biostratigraphical evidence indicates that the main offset may have occurred in the Late Oligocene to Early Miocene interval. During the Palaeogene the overall tectonical style was compression (Fodor et al., 1992 Tari et al., 1993). The southern margin of the Hungarian Palaeogene basin must also have been strongly affected by compression, particularly during the Late Oligocene to Early Miocene. As a result blocks or most likely anticlines may have been uplifted and thus have produced a highly differentiated bottom topography.

Following the Kiscell Clay, deposition of early Egerian silts continued in local depressions at many localities with intercalations of redeposited sediments. At the same time a significant northward shift of the southern shoreline took place, followed by subaerial exposure of the former

basin-fill deposits of mainly Kiscellian age (Tari unpublished report, 1992). On the "new" margin (Gödöllő-Tóalmás area, Figs. 2.6 & 2.8) deposition of glauconitic sandstones started in the Late Oligocene (Egerian), most likely in shallow marine environments (cf. Nagymarosy unpublished, 1988; Lakatos et al., 1991).

The stratigraphic gap between the Kiscell Clay and the overlying sandy deposits, and the coeval redeposition reported from other localities are interpreted as the result of a major sea-level fall around the mid-Oligocene (Tari unpublished, 1992). The signal of this sea-level fall was most likely enhanced at areas, which were formerly and/or simultaneously uplifted by transpression along the southern basin margin.

Sedimentation continued during the Egerian, with the deposition of the Szécsény Schlier surrounded by sandy deposits along all of the margins (Fig. 2.8). The deepest part of the basin was a relatively narrow trough from SSW to NNE. Around the beginning of the Miocene the centre of subsidence shifted to the northeast, forming a WSW-ENE striking trough north of the previous trough. The "new" depression was continously filled by bathyal silts. This subsidence also resulted in a flooding of the former northeastern margin.

At the Egerian/Eggenburgian boundary another sea-level fall affected the evolution of the area (discussion in chapter 8). This eustatic sea-level fall was also recognized in other Central Paratethys basins (Rögl & Steininger, 1983; Baráth & Kovác, 1989; Rusu, 1989; Bachmann & Müller, 1992 and many others). As a result, deposition of the schlier was followed by the deposition of coarse-grained sandy formations on the eastern and western flanks of the basin (Fig. 2.8). The strongly compressed southern margin of the Palaeogene basin became subaerially exposed again. As a consequence the Egerian glauconitic sandstone became subject of subaerial erosion and recycling. Reworking of Egerian sands during the Eggenburgian provided a great bulk of debris for the Pétervására Formations. From south to north the basin gradually was filled to and above sea level.

It is concluded that the different character of the sequences in the north and the south indicates synsedimentary tectonic activity in the southern part of the basin. The ongoing transpressive tectonics in combination with major sea-level falls produced more unconformities in the southern sequences than in the northern ones. The top of the "Palaeogene sedimentary succession" - Late Eocene - to early Miocene, - is a major disconformity, indicating the inversion of the Palaeogene basin before extension-related subsidence of the Pannonian basin started in the Middle Miocene (Horváth, 1986).

CHAPTER 3
THE PÉTERVÁSÁRA SANDSTONE

General description

The Pétervására Sandstone is exposed in northern Hungary and southern Slovakia over an area of 1500 km^2 (Fig. 3.1). The northernmost occurrences are found east of Filakovo near the Slovakian/Hungarian boundary. The most southern exposures are found in the northern foothills of the Mátra Mountains. The bulk of the Pétervására Sandstone occupies the Heves-Borsod Hills between Salgótarján and Ózd. The westernmost outcrops occur east of Szécsény. Towards the west and the north the Pétervására Sandstone interfingers with coeval bathyal sediments (Fig. 2.8; Szécsény Schlier). In the east the formation is bounded by the Darnó fault zone (cf. Fodor et al., 1992). Deposits, assumed to be part of the Pétervására Formation, known from sporadic boreholes 50 km south of the study area, are not discussed in this thesis (see Chapter 2). A comprehensive review and summary about the Pétervására Sandstone is given by Báldi (in Hungarian 1983, in English 1986). In South Slovakia the formation is known as the Filakovo Sandstone (summarized by Vass et al., 1988).

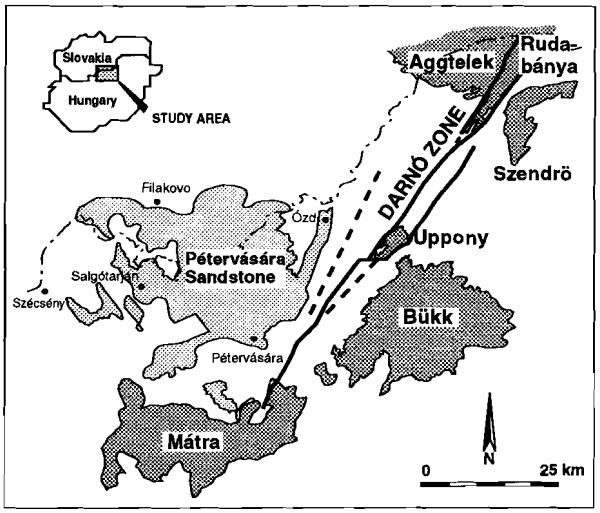

Fig. 3.1. The area with outcrops of the early Miocene Pétervására Sandstone in North Hungary and South Slovakia.

The thickness of the Pétervására Sandstone increases from the SE towards the NW, from 150 m to 600 m (Báldi, 1983; Hámor 1985). It comformably overlies bathyal siltstones (Fig. 2.7; Szécsény Schlier, Báldi, 1983), but the transition from bathyal deposits to the neritic sandstone is rapid and covers only a few metres (vertical changes in the depositional sequence are discussed in Chapter 8). The Pétervására Sandstone consists mainly of medium- to coarse-grained cross-stratified sandstone with conglomerate intercalations, which contain remnants of large molluscs

(Cs.Meznerics 1953, 1959 Báldi, 1983). The reconstructed faunal assemblage indicates normal saline, strongly agitated shallow-marine waters (Báldi, 1983, 1986). Shark teeth found throughout the formation also confirm its marine origin. The Pétervására Sandstone is overlain by terrestrial deposits (Zagyvapálfalva Formation). They form the top of the early Miocene sedimentary cycle (Fig. 2.9; and Chapter 8).

In general the various sandy lithologies are poor in fossils. Altough the exact chronostratigraphic position of the formation is subject of debate (cf. Báldi, 1983 and Hámor, 1985), the mollusc fauna in the conglomerates provides direct evidence for an Eggenburgian age. Studies of nannoplankton in the underlying silty deposits (Báldi-Beke, Nagymarosy in Báldi, 1983; .i.Nagymarosy & Báldi-Beke, 1988;), also confirmed that the Pétervására Sandstone cannot be older than Eggenburgian (early Miocene).

Petrographically the Pétervására Sandstone is a lithic arenite (Vass et al, 1988). The majority of the clastic material consists of fragments of neutral-acidic volcanics, plutonic rocks, some metamorphics and some sedimentary rock fragments (for details see Chapter 4). These constituents are not characteristic enough to determine potential source areas with sufficient accuracy. Nevertheless sediment transport pathways from S to N indicate that source areas should have been located south of the North Hungarian Palaeogene basin (Chapter 2.). Less than 10% of the total clastic material in the Pétervására Sandstone has been derived from an ophiolitic rock suit east of the Darnó fault (Chapter 4).

The Pétervására Sandstone has also been known as "glauconitic sandstone" for long. The amount of galuconite is highly variable throughout the formation. The coarser facies seems to contain more glauconite than the finer ones. The surface of the larger glauconite grains displays signs of longer transport (Bondor, 1960). According to radiometric age determinations origin of the glaucnite grains dates back to the Eocene, though its potential source rocks are unknown (Balogh Kadosa in Báldi, 1983). All these suggest that the glauconite has been intrabasinally reworked (Bondor, 1960; Báldi, 1983). Further research is needed to clarify the origin of the glauconite grains.

Some characteristic outcrop sections and major localities: field observations

Outcrop conditions are reasonably good, due to weak to moderate cementation of the sandstone. The Pétervására Sandstone crops out in isolated spots on gentle hillsides along valleys of streams. In addition, at some major cliffs bedrock has been exposed by erosion following intensive deforestation in the 17-18th century. The whole area is tilted with a few degrees towards the north due to very young tectonics (Fodor, pers. comm., 1992). Therefore the sandstone, exposed in the main south-north striking valleys, becomes younger towards the north.

Field observations will be summarized below (Fig. 3.2) and some characteristic outcrop sections are described (Fig. 3.3).

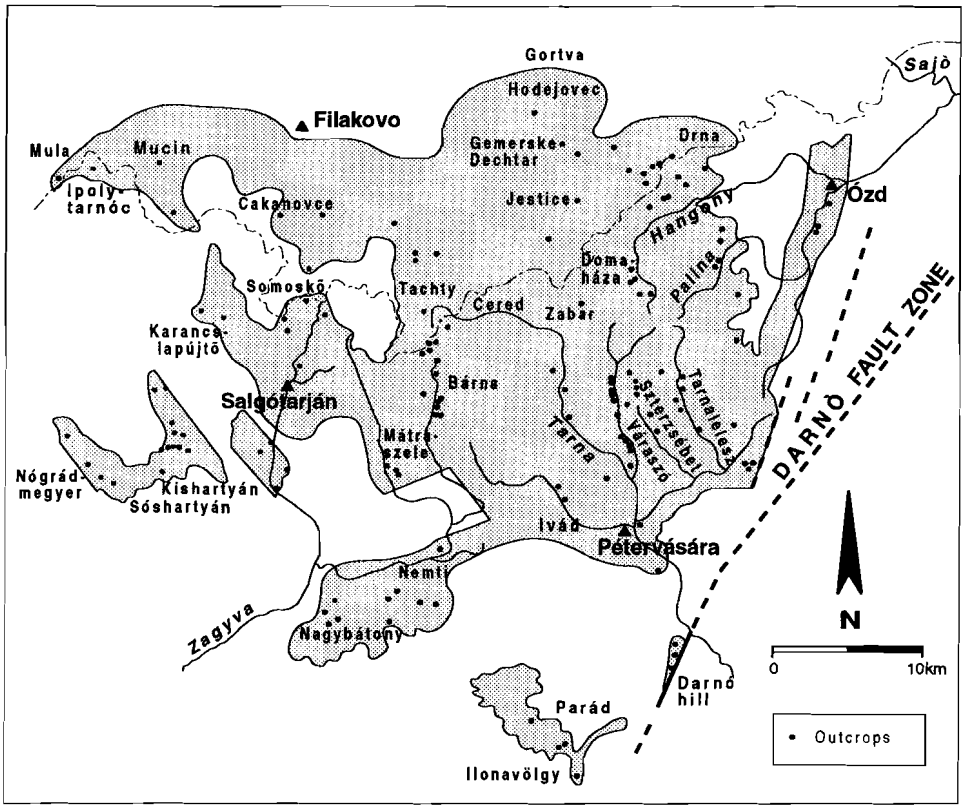

Fig. 3.2. Locations of Pétervására Sandstone and geographic names used in the text.

Pétervására

The type locality shows a 5 m high succession of upward thinning thick-bedded (10-50 cm), medium-grained glauconitic sandstone. Bedding planes are sharp, weakly undulating surfaces, covered by mm-thick, mica-rich silt laminae. There are small differences between the inclination of beds, but neither small- nor large-scale crossbedding could be recognized. Elsewhere in the Pétervására Sandstone such sedimentary structures are characteristic.

Part of the sandstone is exposed in a bentonite open-pit mine south of the village. The floor of the pit is a strongly cemented surface of coarse glauconitic sandstone, overlain by 1 m of bentonite. In the bentonite well developed current ripples have been preserved. Above, about 10 m of silty, very fine sandstone with undulating, discontinouos bedding surfaces occurs.

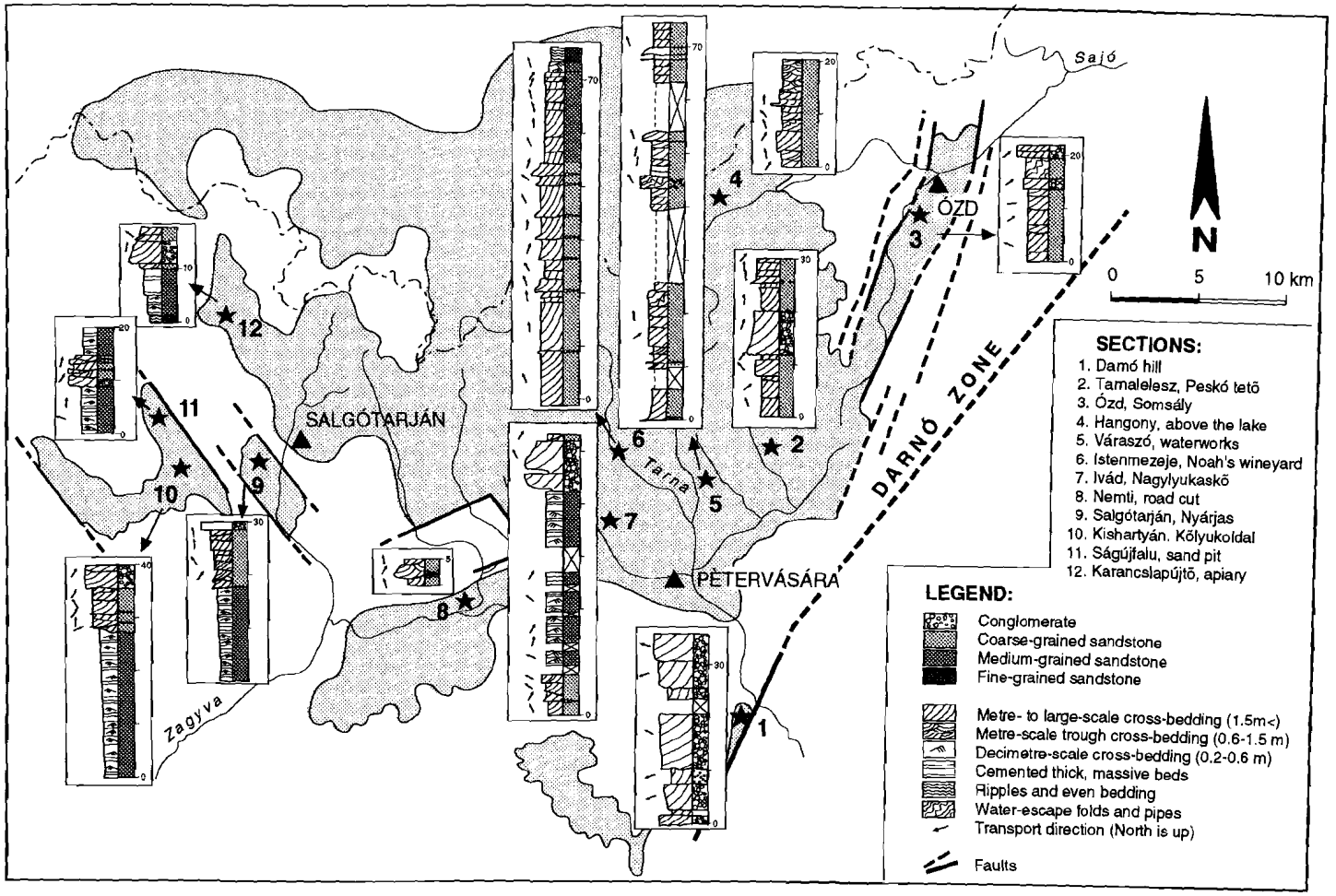

Fig. 3.3. The most important outcrop sections of the Pétervására Sandstone.

Váraszó and Szenterzsébet valleys

At the southern entrances of both valleys (Fig. 3.2) glauconite-rich fine- to medium-grained sand with intercalations of undulating silt occurs. Flaser bedding is common. Thickness of beds ranges from 5 to 20 cm. A direct contact with the Szécsény Schlier is not visible, but the topographic expression of both formations indicates that the schlier lies underneath. Above, an approximately 5-15 m thick succession of silty sands and a series of medium-grained weakly cemented, mica-rich sand occurs. Bedding planes are only shown by rows of moderately cemented nodules.

North of Váraszó an approximately 70 m thick section of metre-scale cross-bedded sandstone is located (Fig. 3.3, section 5). Grain size changes between medium-grained to granular sand. Dip directions and angles of foresets vary between N-NNE in the subsequent sets. Normally height of cross-stratified beds does not exceed 1-2 metres. At two horizons synsedimentary deformations, gently folded foresets and water-escape pipes appear. Further to the north trains of m-scale cross bedding can be followed over 6 km in the valley. Some sets exceed 3.5 m in height. Foresets are usually steep and tabular (Fig. 3.4). Foreset laminae (1.5-8 cm thick) are displayed by slight variations of the grain size.

In the parallel Szenterzsébet valley (Fig. 3.2) similar m-scale cross-bedding can be studied both vertically in cliffs (e.g. Nagykő) and longitudinally along the valley. At the head of the valley a set of 1.8 m height with tangential foreset laminae in medium- to coarse-grained sandstone is present. These foresets are reactivation surfaces, followed by slightly steeper foresets, which downlap on the reactivation surfaces (Fig. 3.4). Dip of successive foresets decreases again until a next bunch of tangential foresets. Parallel with changes of dip angle, the thickness of foreset laminae also increases and decreases. At another exposure in the same valley tangential foresets of medium-grained sandstone are covered by 10-30 cm long silt drapes continuing in finer grained bottomsets.

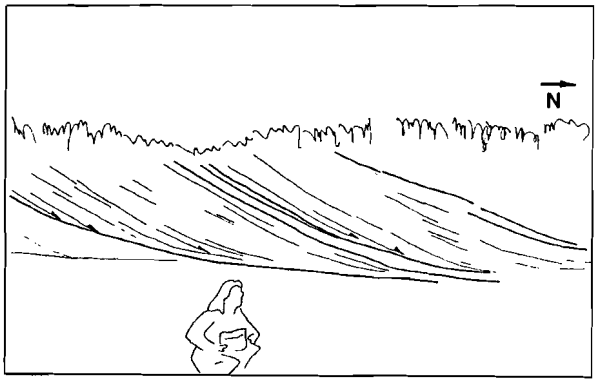

Fig. 3.4. The shape of foresets changes cyclically from tangential to steep planar and to tangential again in a 1.8 m high cross-set (Szenterzsébet valley). The steep planar foresets downlap onto the tangential reactivation surfaces . This is interpreted to be the result of spring-neap tidal cyclicity (see later this chapter and chapter 5). Thickness of single foresets is hardly visible in this outcrop. Drawn after photograph and field sketch.

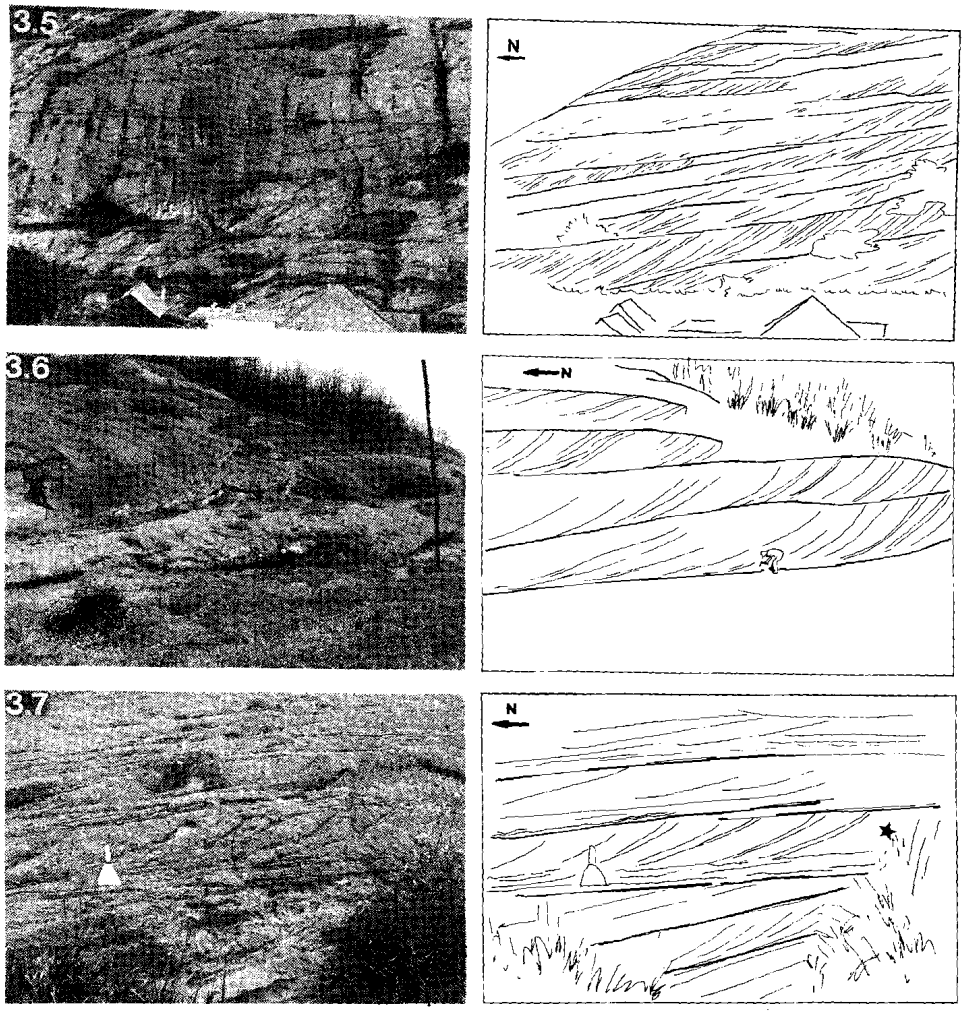

Fig. 3.5. "Noah's vineyard" at Istenmezeje. Large- to giant-scale crossbedding is stacked up to a total thickness of over 70 m. The steep set at the lower part of the picture is 10 m high. Note the about 10^{0} teconic tilt towards the north (left), the change in thickness of cross-sets and the dip of foresets. The name of the cliff comes from the appearance of cemented nodules along the joints. (Houses are built just below the cliff)

Fig. 3.6. Other cross-sets consisting of very well sorted medium-grained sanstone with tangential foresets and poorly developed bottomsets at "Noah's vineyard".

Fig. 3.7. Dm-scale cross-bedding with alternating drape-spacing (cm and dm) near to the top of the Istenmezeje section. These are interpreted as products of subsequent stronger and weaker dominant tidal currents, and thus reflect the diurnal inequality of the early Miocene tides. Length of the hand shovel is 20 cm.

Tarna valley and Istenmezeje

"Noah's vineyard", at Istenmezeje in the Tarna valley, is a 500 m long, more than 60 m high outcrop (Fig. 3.2, 3.3, section 6). It is built up of moderately cemented, glauconitic sandstone. At cross-points of bedding planes and tectonic joints extremely strongly cemented nodules have formed, that explain the strange name of the locality (Fig. 3.5). The lower 50 m consists of medium- to coarse-grained sand, and the upper 20 m of very well-sorted medium-grained sand (Fig. 3.3). Stacked large- and giant-scale cross-sets are particularly characteristic for this exposure (Figs. 3.5, 3.6). Average height of sets is 3.5 m, but 8-12 m high sets also occur. Two upwards thickening-thinning series of sets occur.

Erosional pits at the base of the sets are not significant. Dip angle of foresets is 25° on the average, but occasionally it exceeds 30°. The large foresets are not covered by mud drapes. In another outcrop, 4 km to the north, rippled, double muddrapes occur in bottomsets of large-scale trough-crossbedding. At the top of the section the height of sets decreases to 20-40 cm, parallel with a decrease of grain size to fine sand. Well developed mud drapes on foresets and bottomsets, and alternations of cm- and dm- thick foresets occur (Fig. 3.7). The cross-bedded fine sandstone is overlain by thin-bedded micaceous fine-grained sand with silty laminations. The relatively abrupt appearance of fine-grained sediments above the thick succession of coarse sandstone marks a surface of regional importance (see Chapter 8).

North of Istenmezeje along the main road and also in side-valleys the thin-bedded, fine silty sand crops out in small spots. The low and very gentle topography of the area also indicates that the bedrock is soft fine sand, which continues along both sides of the Slovakian/Hungarian border.

Cered road

In the cut of the sinuous mountain road along the state boundary, north of Bárna valley (Fig. 3.2), 2-3 m thick successions of thin-bedded, moderately cemented massive very fine- to fine-grained sandstone occurs. There is some variation in grain size and degree of cementation (Fig. 3.8). The latter accentuates the bedding. In one exposure, a few metres wide and very shallow trough with rippled mud drapes occurs. Nearby dm-scale cross-stratification in coarse sand is shown by variations in glauconite content of foresets.

Between Tachty and Gortva (Slovakia)

On the Slovakian side, near Tachty (Fig. 3.2) the massive, well sorted, thin-bedded fine- to very fine-grained sandstone, found along Cered road, continues, with a generally low glauconite and high mica content. The gentle hills between Tachty and Gortva, around Hodejovec and Jestice consist of the same rock variety (Fig. 3.2, 3.8). Rarely fine, dm-scale cross-bedding is obviously present, with burrows up to one cm in diameter. At Gemerske Dechtar (Fig 3.2) a great variety of burrows occurs, and rippled sand with well developed mud drapes is common.

From Zabar to Domaháza

A very thick series (over 100 m) of horizontally bedded silty fine-grained sand is exposed along the mountain pass between Zabar and Domaháza (Fig. 3.2). Ripples, lenses and cemented nodules are common in the sand, which is associated with flasers and laminations of silt. A solitary set of m-scale cross-bedding in coarse, pebbly sand, with sigmoidal foresets occurs within the silty sand (Fig. 3. 9). Some foreset laminae have been weakly folded due to dewatering processes.

Around Domaháza horizontally bedded, fine- to medium-grained mica-rich sandstone occurs. Occasionally silty laminations and small trace fossils are present in these sandstones.

Hangony valley

Along the northern side of the Hangony valley (Fig. 3.2), the Pétervására Sandstone is exposed with large-scale crossbedding with well-developed foreset laminations as the dominant sedimentary structure. The variation of thickness of foresets was measured at two exposures (Fig. 5.2; analysis of data in Chapter 5). Grain size of laminae varies between medium- and coarse-grained sand, with occasional horizons of small pebbles. Cross-sets are 2-4 m thick and often show erosional lower boundaries. Occasionally the sigmoidal bend of the upper part of the foresets has been preserved.

In some sections the almost full succession from the Szécsény Schlier to the Pétervására Sandstone is exposed. The fine-grained transitional part of the sandstone is covered in these cases, but 20 m above the schlier, metre-scale cross-bedding appears, with thin foreset lamination. Above some few large sets (3 m high), and smaller but still m-scale crossbedding follows with mud drapes and tiny trace fossils (Fig. 3.3, section 4). In an adjacent valley the section is similar: giant sets up to 7 m thick compose a 20 m high section topped by m-scale cross-sets. In this upper part the presence of finer bottomsets is obvious.

The series of superb outcrops continues on the Slovakian side, south of Drna (Fig. 3.2). Dominantly m-scale sets of cross-stratified medium- to very coarse-grained glauconitic sandstone are stacked into 10-40 m thick successions. Occasionally sigmoidal bundles were found. Grain-size variation in the foresets is well developed. Pebbly horizons contain fine bioclastic debris. Low angle differences between series of foresets may have been formed due to reactivation instead of the superposition of a next set.

Around Ózd

At the southern part of the city, thin-bedded, very well sorted, medium-grained mica-rich sandstone occurs. Bedding is revealed by a few cm thick layers of slightly finer sand, and mm-thick silt drapes, as well as by horizons of strongly cemented nodules. It mostly looks massive. Wide and shallow troughs withouth internal cross-lamination occur only locally. Mud-filled burrows with a length of a few mm are abundant. The sandstone is exposed at maximum 20 m of thickness.

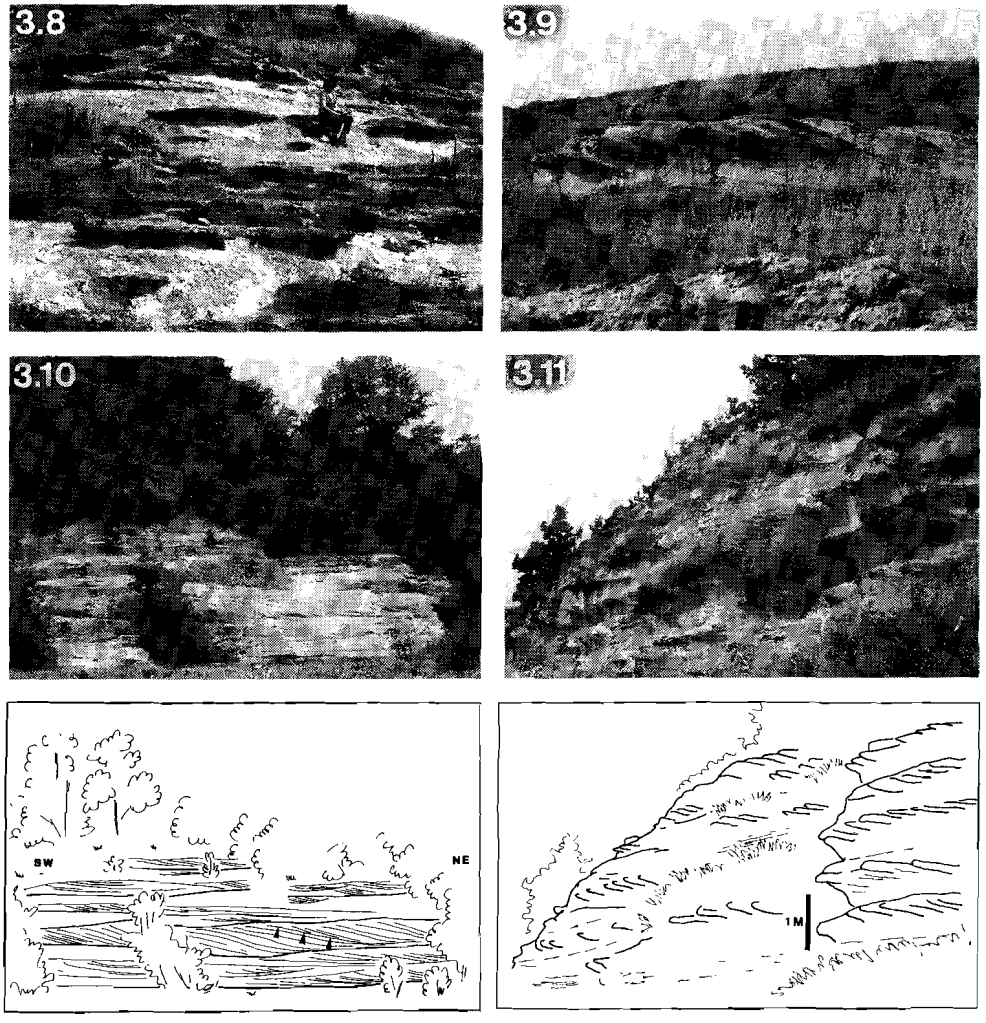

Fig. 3.8. Thin-bedded fine-grained friable sandstone, with strongly cemented horizons near Hodejovec (Slovakia).

Fig. 3.9. Cross-bedded set in medium- to coarse-grained sandstone with sigmoidal foresets along the Zabar-Domaháza road.

Fig. 3.10. Cross-sets and mud-draped foresets in medium-grained sandstone in the Palina valley. Transport is to NNE (right). Note the gentle dipping reactivation surface (marked by arrows) ovarlain by steeper downlapping foresets .

Fig. 3.11. Cementation accentuates the difference between fine-grained bottomsets and coarse-grained foresets in m-scale dunes near Palina valley.

Upwards in the section coarse- to very coarse-grained, pebbly sandstone with a significant amount of mollusc debris appears abruptly. The dominant sedimentary structure is large-scale (5-10 m high) trough cross-bedding. Pebbles are not confined to the bases of troughs, but form foresets as well.

South of Ózd (Fig. 3.2, 3.3, section 3), only large-scale cross-sets are exposed. The lower 10 m of the section consists of medium- and coarse-grained glauconitic sandstone, with undisturbed, parallel foresets. Above very coarse, granular sandstone and conglomerate follows. Foreset laminae have been folded and twisted into large dishes and pipes (Fig. 4.5), which are interpreted as water-escape deformations (see discussion in chapter 4).

Palina valley

In this valley and its side valleys 15-30 m high sections consisting of medium-to coarse grained sandstone occur. Stacked, 1-2 m high cross-sets with tangential foresets, occassional reactivation surfaces and mud drapes (Fig. 3.10) alternate with large-scale (more than 3 m high) cross-beds with dominantly steep planar foresets. Alike at Hangony the sections are topped by dm- to m-scale crossbeds with well developed fine-grained bottomsets (Fig. 3.11).

Along the forestry road, through the hills between the Palina valley and the adjacent Gyepes valley, horizontally bedded fine-grained, silty sand crops out. Bedding is undulatory. Flasers and continuous laminae of silt alternate with sand.

Tarnalelesz valley and Peskótető

East of the southern entrance of the Tarnalelesz valley (Fig. 3.2) a continuous, upward coarsening and thickening section represents the transition between the underlying schlier and the Pétervására Sandstone (Kő-hegy). The lower 2 m of the section consists of mica-rich, fine sand with silty laminations and flasers and is overlain by a 10 m thick series of friable, medium-grained glauconitic sandstone with mud drapes separating beds of 10-30 cm thick. Pieces of ripped-up mud clasts, flakes and burrows up to 1 cm in diameter are abundant. Towards the top dm-scale crossbedding appears, which grades upwards into m-scale cross-bedding building up the upper 5 m of the section. The bottomsets of the stacked cross-sets contain mud-drapes and a high amount of rip-up mud clasts. The rate of cementation also increases upwards, parallel with the increasing grain size.

Along the Tarnalelesz valley large-scale cross-sets crop out. They consist dominantly of coarse, glauconitic sandstone. Pebble strings in bottomsets and pebbly foresets are common. Occasionally mud drapes are present on the lower parts of the tangential foresets .

The most important outcrop is the section at Peskótető (Fig. 3.3, section 2). Large-scale cross-bedding with sets of 2.5-5 m high, consisting of medium-grained, mixed bio- and siliciclastic sandstone forms the lower part of the section. Above, an 8 m high (Fig. 4.6) cross-bedded conglomerate bed occurs, with a large amount of mollusc debris. The size of the gravel decreases

upwards from small cobbles at the base to granular sand at the top. Foresets are defined by 5-10 cm thick graded beds and elongate cemented nodules (Fig. 4.6). Dip angle of foresets is steep. Similarly to that of normal sandy bedforms it reaches values as high as 30°. Upon the conglomerate again medium- to coarse- grained sandstone occurs with large-scale (5 m high) cross-bedded sets. The topmost 5 m of the section consists of medium- to fine-grained sand with sets of decreasing height. Here finer grained bottomsets are common.

Darnó hill

For a detailed description see underheading "Darnó Conglomerate" in Chapter 4.

Around Parád (foothills of the Mátra)

Only minor outcrops were found in this area (Fig. 3.2). They show mainly silt-rich, micaceous fine sand. Admixtures of greenish-grey tuffaceous material are also characteristic here. In Ilonavölgy, near Parád (Fig. 3.2), a bed of pebbly sand contains molluscs which confirm the Eggenburgian age of the formation. Sedimentary structures as found elsewhere in the Pétervására Sandstone are not present there.

Around Nagybátony

Near Nemti (Fig. 3.2) the southward facing cut of road no. 23 exposes a typical example of the pebbly large-scale trough cross-stratification (Fig. 3.3, section 8). Grain size is variable from medium-grained sandstone to large pebbles. The bottom of the large troughs is paved by the coarsest material. Pebbly foresets are rare.

The village of Nagybátony (Fig. 3.2) sits on an anticline (Schréter, 1940). Therefore sequences exposed in the southern wing of the anticline are younger towards the south. In the village a 15 m thick section of dominantly medium-grained glauconitic sandstone occur. Large-scale (3-4 m high) and m-scale, mainly trough cross-sets are stacked on each other. Large rip-up mud-clasts and cm-thick burrows are common. The adjacent "Szoros-patak" section - stratigraphically sligthly higher - consists of 10 m thick fine-grained, micaceous, glauconitic sand with silty interlaminations and cemented nodules. Above a thickening upwards series from dm- to m-scale crossbedding in medium- to coarse-grained sandstone appears. Foresets are tangential, with a pebble lag in m-scale beds, and mud drapes and fine-grained bottomsets in dm-scale beds.

In small outcrops a few km to the east (e.g. Mátramindszent, Szuha), m- and dm-scale cross-bedding in fine- to medium-grained glauconitic sandstone occurs. Often a variation of glauconite content stains foresets in well-sorted sand.

Ivád valley

In the valley thick successions of thin-bedded virtually massive fine sandstone with cemented nodules occur. The main section (Nagylyukaskő) is a 50 m high sequence (Fig. 3.3,

section 7). At the bottom a few stacked sets of m-scale crossbedding occur in medium to coarse pebbly sand with cemented nodules and dewatering structures. It is overlain by a 35 m thick series of thin-bedded medium-grained glauconitic sandstone. Mud drapes on dm-scale foresets and on bottomsets, rip-up mud clasts and burrowed bedding planes are common. Above, two conglomerate beds (3 and 6 m thick) appear, with faintly oblique internal bedding planes. At the top of the sequence cross-bedded fine-grained sandstone occurs.

Bárna valley and vicinity

In Mátraszele (Fig. 3.2) thin-bedded fine-grained sandstone appears with a thickness of about 50 m. In the lower part of the succession mud drapes on dm-scale tangential foresets of medium-grained glauconitic sandstone are abundant. Upwards, dm-scale crossbedding is shown by a fluctuation of the glauconite content, and mud drapes become rare. Higher up the amount of mica increases, internal sedimentary structures are obliterated, and massive fine-grained sandstone with cemented nodules occurs.

The hills around Bárna (Fig. 3.2) are built up of massive, thin-bedded fine-grained sandstone up to a thickness over 100 m. Recognition of bedding depends on subtle grain size-variations, which have determined the rate of cementation. Internally no sedimentary structure is apparent in the 10-30 cm thick beds. Some silt laminantions and small mud-filled burrows appear occasionally. North of the village the protected Szárkő hill consists of medium-grained sandstone, but distinct bedding planes cannot be seen. The thickness of the faint beds is 0.5-1 m.

Around Salgótarján

Two main outcrops are found in the city of Salgótarján (Fig. 3.2). At Nyárjas (Fig. 3.3, section 9) an upward coarsening section occurs. The lowermost 20 m consist of weakly cemented fine sandstone with dm-scale tangential crossbedding covered by mud drapes. Rip-up mud-clasts are very common. Above, bed thickness increases and gradually m-scale crossbedding appears in well-sorted medium grained-sandstone with silty bottomsets and large mud-filled burrows. At the top a 1 m thick massive pebbly granule bed occurs.

Medium- to coarse-grained glauconitic sandstone with carbonate concretions, biotite flakes and bentonite fragments is exposed in a cut of road no.21. Large-scale trough crossbedding, mud-draped tangential foresets with noteworthy fluctuations of grain size and muddy bottomsets are present (Fig. 3.12). Packages of closely and widely spaced mud drapes in combination with shallower and deeper erosional scours indicate fluctuations in current energy and in migration velocity of the bedforms (Fig. 3.12).

Near to the state boundary, at Somoskőújfalu railroad cut (Fig. 3.2), an approximately 15-20 m high succession of very fine, micaceous silty sand occurs. Primary bedding was destroyed most likely by intense bioturbation. Rarely lenses with ripples are present.

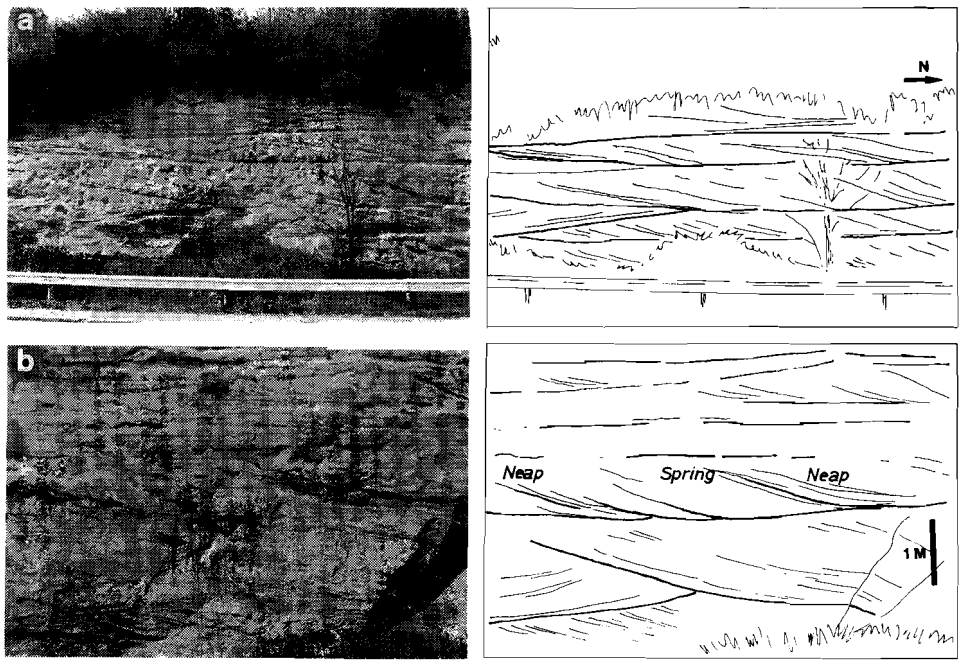

Fig. 3.12. a. Metre-scale crossbeds with tangential foresets and very wide troughs at Salgótarján, roadcut. Mud drapes on the toes of foresets and in bottomsets are common. b. Closely and widely spaced foresets with mud drapes as the result of spring - neap tidal cyclicity.

Another pebbly section occurs at Karancslapújtő (Fig. 3.3, section 12), above a few metres of micaceous, glauconitic fine sand with silty laminations. Large pebbles in a very coarse sandy matrix form foresets. There are muddy bottomsets with large vertical, sand-filled burrows. It is overlain by a cross-bedded set consisting of very coarse, pebbly sandstone.

Around Cakanovce (Slovakia)

Stacked sets of m-scale crossbedding occur in medium- and coarse-grained glauconitic sandstone near Cakanovce. Gently dipping tangential and sigmoidal foresets with rip-up mud-clasts at their bases are common. Dm-scale sets of trough crossbedding with mud clasts and tiny burrows are also present. Further to the west, at Mucin, the dm-scale cross-bedding is very well developed in very coarse sand, which shows a great contrast with the very common mud drapes (Fig. 3.13). In the westernmost outcrop of the Pétervására Sandstone, near Mula (Slovakia), thin-bedded fine sandstone with dm-scale sets, silty bottomsets and cemented nodules occurs.

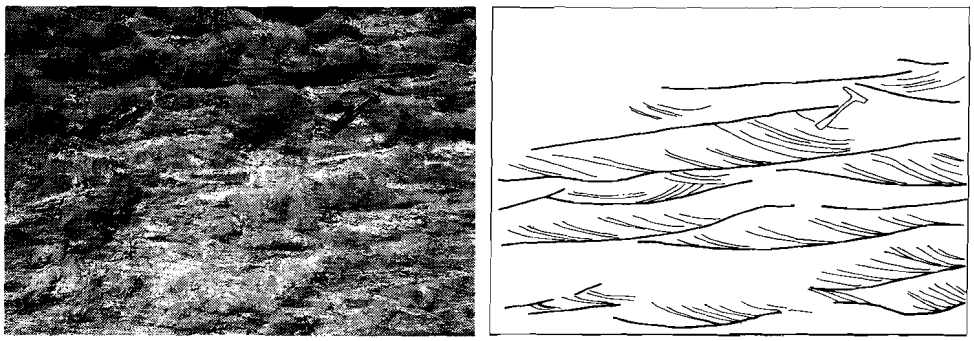

Fig. 3.13. Dm-scale trough cross-beds with thick calcareous mud drapes near Mucin, Slovakia.

Around Kishartyán

The Kőlyukoldal sequence south of Kishartyán (Fig. 3.2) is exposed in two different sections, separated by a minor fault. East of this fault a 17 m thick section is built up of stacked cross-stratified sets. Grain size decreases upwards from very coarse- to medium-grained sandstone, parallel with a decrease of height of sets. The height ranges from 0.6 to 3 m, and the average is 1.3 m. The presence of large troughs is obvious. The variation of grain size in foreset laminae, particularly in the large, lowermost set is spectacular (Fig. 5.5).

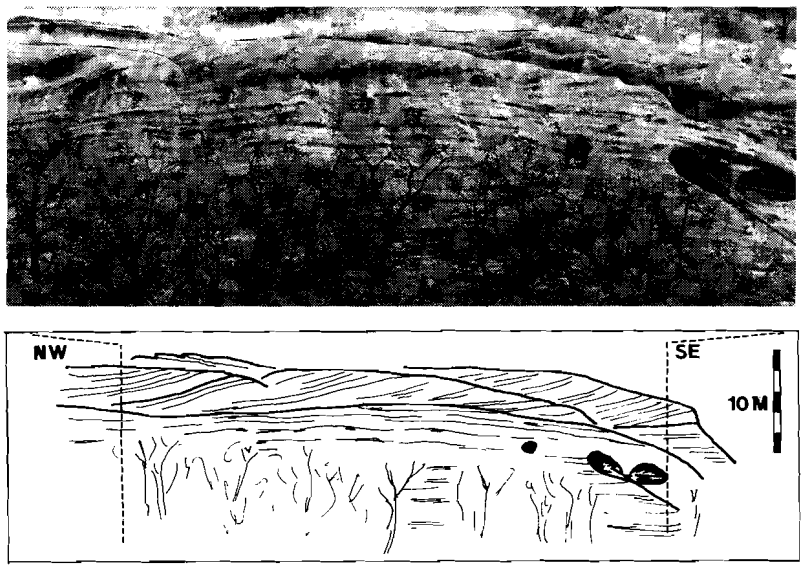

Fig. 3.14. Horizonatal strata of thick-bedded medium-grained sandstone are overlain by large-scale cross-bedded coarse-grained sandstone at Kishartyán Kőlyukoldal. Note the weakly erosional boundary between the two facies.

West of the fault an upward thickening and coarsening section occurs. The lower 15 m of the section (Fig. 3.3, section 10) consists of medium-grained glauconitic sandstone with dm-scale bedding. Bedding is indicated by rows of strongly cemented nodules and lenses, bedding planes are often marked by rippled silt drapes. Occasionally dm-scale cross-bedding is clearly present. Above, with a weakly erosional boundary, large-scale (up to 3 m high) crossbedded sets of very coarse - granular sandstone follow (Fig. 3.14). At the top of the section a large trough with pebbly foresets occurs.

At the nearby Sóshartyán (Fig. 3.2) very well-sorted fine sand with dm-scale delicate glauconitic sand appears a few metres above the Szécsény Schlier. Further to the west, at Nógrádmegyer, medium- to very coarse-grained sandstone with stacked m-scale cross-bedding occurs. The shape of foresets is tangential or occasionally sigmoidal. The top of the sequence is formed by large pebbly, granular foresets.

Facies units: description and interpretation

Summary of field observations

The Pétervására Sandstone consists of a series of fine- to coarse-grained sandstones with conglomeratic horizons. The dominant sedimentary structure is cross-bedding at various scales. Lateral changes and trends in grain size become obvious when the distribution of the average coarsest fraction, which reflects the flow competence, is plotted on the map of outcrops (Fig. 3.15).. Grain size decreases towards the NW - to the area of Salgótarján and with increasing distance to the Darnó fault - from dominantly coarse to very fine sand. Only conglomerate lobes form finger-like intrusions and interrupt the westward decrease of grain size. Roughly shore-parallel zones exist, characterized by the same grade of material. The same outline appears if size (thickness of cross-sets) or type of bedforms (laminated, thin-bedded, dm-scale, m-scale or large-scale cross-stratification) is plotted on the same map. There is a fairly good correlation between dominant grain sizes, types of bedforms and thickness of beds: the coarser the material, the larger the bedform. Conglomerate intervals exceed 6-8 m in thickness. In coarse-grained sandstones large- (2-4 m high) and giant-scale (6-12 m high) cross-sets occur. In medium-grained sandstone usually m-scale (0.6-1.5 m) cross-stratification is found, whereas dm-scale (10-50 cm) cross-beds occur in fine- to medium-grained sand. A similar zonation is found between Salgótarján and Szécsény (Fig. 3.4). The repetition of zones, which was also observed in vertical sections, was interpreted as the result of shifts of facies belts in time (discussion on vertical facies changes in Chapter 8).

Sediment transport directions, indicated by foreset-dip directions of various-scale cross-stratification, point uniformly towards the NNE. The average for 99 outcrops (over 700 measurements) is north (18° NE), with a circular standard deviation of 30° (Fig. 3.15). On the average the dip angle of foresets is 25°.

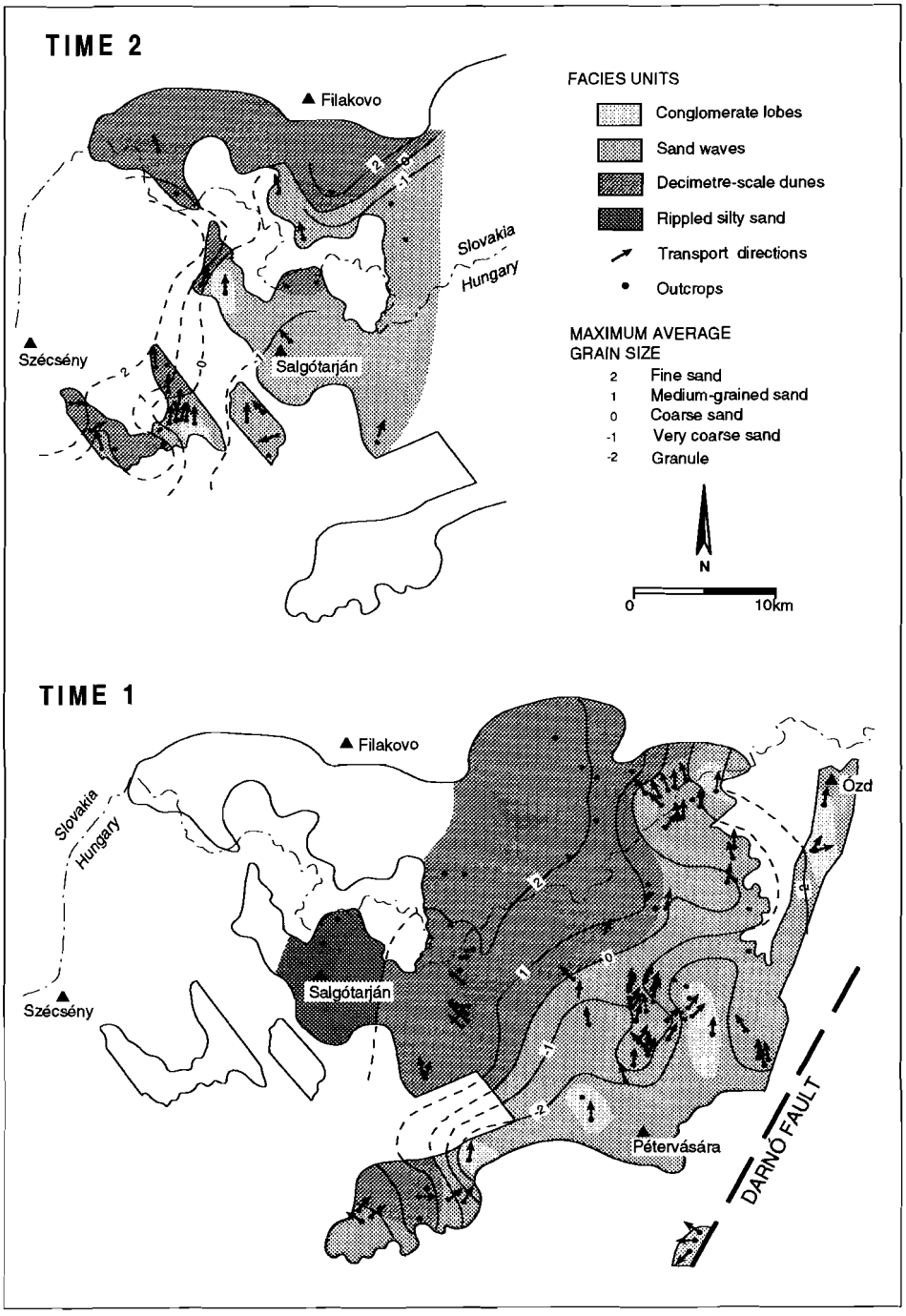

Fig. 3.15. Distribution of grain size and transport directions of the Pétervására Sandstone.

Based on the field observations, the distribution of grain size, bed thickness and bedforms, the following four facies units are distinguished in the Pétervására Sandstone:

- 1. Laminated silty fine to very fine sand;
- 2. Thin-bedded fine- to medium-grained sandstone with dm-scale crossbedding;
- 3. Medium- to coarse-grained sandstone with m-scale and larger scale cross-bedding;
- 4. Conglomerates with large-scale crossbedding.

Laminated silty fine to very fine sand

The thin-bedded silty facies has a thickness over 100 m. Horizontal bedding planes are uneven, undulatory with thin beds, which are a few cm thick and consist of fine, micaceous sand. They alternate with flasers of silt. In the sand faint cross-lamination is infrequently present. Commonly sedimentary structures within beds were destroyed by bioturbation, but it did not penetrate deeply into underlying beds. Occasionally dm-thick glauconitic sand with cross-bedding is intercalated in the silty-fine-sandy succession. Laterally continuous sandy beds with hummocky and swaley geometry were not observed.

Only the absence of molluscs distinguishes this facies unit from the atypical sandy variety of the underlying and/or laterally adjacent Szécsény Schlier Formation. In vertical sections the laminated silty, fine sand is usually overlain by thin-bedded fine- to medium-grained sandstone.

Interpretation

In general ripples, revealed occasionally by cross-lamination or indirectly by flasers of silt (cf. Reineck & Wunderlicht, 1968) indicate that currents were active during deposition. The intercalations of dm-scale cross-bedding point to temporary influences of strong currents, but no channels or erosional scours were encountered. The silty laminations reflect periods of very low current velocity during which suspended matter could settle. The moderate activity of bioturbating organisms indicates that this environment was not ideal for benthic organisms, most likely because of the almost continuous movement of silt and sand (cf. Bromley, 1990).

It is difficult to interpret this facies unit without considering the adjacent facies belts. Altough certain structures remind of deposits found in the intertidal flats of e.g. the Wadden Sea (cf. Oost & de Boer, 1994), deposition must have occurred at a water depth of several tens of metres, because this facies unit directly interfingers with the Szécsény Schlier formed at a water depth far below 60 m (Báldi, 1983). The laminated silty sand represents a transitional facies between the schlier and parts of the Pétervására Sandstone deposited in shallower water.

Thin-bedded fine- to medium-grained sandstone with dm-scale crossbedding

The fine- to medium-grained sand and sandstone contain a large amount of mica or intrabasinally reworked glauconite. Bedding is accentuated by weak alternations of grain size,

although the sediment is very well sorted in general. Bedding planes are usually covered by thin mud drapes and very fine sand. Beds are very often structureless and characteristically contain cemented nodules and horizons (Fig. 3.8). The clay content is about 40 % in the nodules and 10 % in the host rocks (Vass et al., 1988). The thickness of beds ranges from 0.2 m to 0.6 m. Dm-scale crossbedding is present in many places. Foresets are covered by mud drapes (Fig. 3.7, 3.13) or are displayed by an alternation of glauconite-rich and glauconite-poor laminae without any silty cover. Trace fossils, horizontal and vertical burrows (simple, curved and branching forms), with muddy walls and sandy fills, occur and vary in length from several mm to cm.

The thickness of this facies unit exceeds 100 m. Laterally it interfingers with the laminated, silty fine sands towards the west and with medium- to coarse-grained sandstone with large-scale crossbedding towards the east. In vertical sections it usually covers the silty facies unit and is overlain by coarse sandstone or conglomerates.

Interpretation

The fine- to medium-grained sand and the dm-scale cross-bedding infers a relatively high-energy environment, with stronger currents than those which goverened the deposition of the laminated silty sand. The lack of ripples, which well could have been formed in this fine- to medium-grained sand, clearly indicates that currents exceeded 0.3 m/s^2 and were strong enough to build dunes (cf. Southard & Boguchwal, 1990). The currents dominantly flowing towards the north have built small dunes, which were dominantly straight-crested (no erosional scour; Fig. 3.7) or ocassionally sinuous-crested (trough cross-bedding; Fig. 3.13).

Mud drapes reflect slackwater periods, when current velocities were reduced to zero. They thus point to a tide-influenced depositional environment (cf. Visser, 1980). The occasional occurrence of double mud drapes infers a subtidal regime. The dip directions of foresets in combination with the local palaeogeography demonstrate that the dominant tidal current was the ebb current. Bedforms generated by the subordinate current were not found.

One excellently preserved series of mud-draped foresets in dm-scale dunes shows an alternation of drape spacing from cm to dm (Fig. 3.7). The successive thick and thin foresets are interpreted as the products of subsequently stronger and weaker ebb currents, and thus indicate a semidiurnal system with a well developed diurnal inequality (cf. de Boer et al., 1989).

The virtually massive appearance of the majority of the beds can be the result of several factors. 1. Sorting is very good, therefore bedding is not apparent. 2. Burrowing organisms completely reworked the uppermost 10-20 cm of the sediments (cf. Oost, 1994). The activity of burrowers is obvious at many localities. (e.g. Ivád, Fig. 3.2). 3. During cementation original structures were blurred. The latter is suggested by the presence of dm-scale cross-bedding in strata just below cemented nodules (e.g. Hodejovec, Fig. 3.2).

Erosional scours or channels were not found in the field of dm-scale dunes, which was formed below low-tide level. Dm-scale dunes covered a large area (see Fig. 3.15), inshore of the zone where the fine silty sediments were deposited, and offshore of the patch of larger bedforms.

Medium- to coarse-grained sandstone with m- and larger-scale cross-bedding

Cross-bedding at various scales developed in the medium- to coarse-grained glauconitic sandstones. The most common is m-scale cross-bedding in well-sorted medium-grained sandstone with set thicknesses of about 0.6-2 m. Foresets are usually tangential, sometimes sigmoidal, and close to their base they are often covered by few-dm-long mud drapes. Finer grained, silty bottomsets, occasionally with ripples are also abundant. Drape spacing and depth of scours at the base of sets (scallops, cf. Rubin, 1987) often show lateral variation (Fig. 3.12).

Large-scale cross-bedding with sets of 2-3 m height commonly consist of planar or tangential foresets (Fig. 3.4, 3.5 and 3.6) with laminae of alternating medium- and coarse-grained sandstone. Mud drapes do not occur in this type of cross-bedding. Thickening and thinning series, and/or increasing and decreasing dip angles of foresets and cyclic variations of the shape of foresets are commonly present (Fig. 3.4 and 3.12). Even larger sets, up to a height of 12 m (giant-scale cross-bedding; Fig. 3.5) consisting of coarse to very coarse sand were encountered. Usually foresets in coarse sand with a large set height are steep (over 30°) and tabular. In medium-grained sand foresets show a lower dip (20°) and they are tangential or sigmoidal and have a smaller set height. The majority of these cross-beds indicate a 2D-geometry of bedforms. Occasionally trough cross-bedding, mainly on m-scale indicate a more complex, 3D geometry (Fig. 3.12). The foreset dip directions are exclusively towards the north.

In some outcrops (see Fig. 3.3 for locations) 50-70 m thick successions of stacked series of various scale cross-stratified sets occur. Set boundaries are partly erosive, though the depth of erosion cannot be estimated. This facies unit interfingers with thin-bedded sandstones towards the west. Intercalations of conglomeratic horizons are common towards the east.

Interpretation

The cross-bedding clearly indicates that relatively large, straight- or sinous-crested subaquaous dunes were present in the "Pétervására Sea". The variability of shape and steepness of foresets, in combination with differences in grain size are obviously the consequence of differences in current velocity (cf. Allen, 1984; Collinson and Thompson, 1989). Steep, planar foresets in coarse sand, formed by avalanching, reflect the smallest relative flow strength and imply straight-crested dunes. The tangential foresets in the same coarse sand point to higher current velocities (Terwindt, 1988). Rippled bottomsets could not form, because the grain size was too coarse to form ripples (cf. Allen, 1984). Tangential foresets in medium-, or even in fine-grained sand indicate that currents were capable to transport sand in suspension.

No erosion was likely to occur in front of the large straight-crested dunes with steep-faced sets in coarse-grained sand. Therefore the set thicknesses of the underlying beds give a good figure for the dune height. The height of the largest sets (several metres) thus indicates a depositional depth between 15-30 m (cf. Allen, 1984, Figs 8-20 and 11-25; Davies & Flemming, 1991).

The occasional presence of mud drapes on foresets of medium-grained sand confirm the idea that sequences were formed in a tide-influenced environment (cf. Visser, 1980). Such

interpretation is, moreover, verified by the lateral variations of inclination and/or shape of the foresets. The most important observation, however is the cyclical variation in thickness of foresets (Tari et al., 1989; and Chapter 5). Similar features were observed in many tide-influenced deposits (Visser, 1980; Boersma & Terwindt, 1981; Siegenthaler, 1982; Allen & Homewood, 1984; Houthuys & Gullentops, 1988a,b Kreisa & Moiola, 1986; Davies & Flemming, 1991) indicating spring-neap tidal periodicity. Thick-thin couplets reflect the diurnal inequality of tides (cf. de Boer et al., 1989), and imply a semidiurnal tidal system.

Foresets of dunes monotonously dip to the north. This indicates a highly asymmetrical tidal system (cf. Allen, 1980), with a dominant current to the north. The subordinate, southward directed current must have been too weak to erode even the upper lee faces of dunes, and thus to contribute to the generation of master bedding (Allen, 1980; or reactivation surfaces, Visser, 1980). The palaeogeographic situation shows that the dominant current must have been the ebb current (Chapter 2, and discussions in Chapter 6). The large-scale straight- or sinuous-crested dunes formed under the influence of asymmetrical, oscillating tidal currents thus can be named sand waves, which fall into class IIIA (*sensu* Allen, 1980). Neither the sedimentary structures nor their distribution indicates that channels have been active, rather a sheet-like morphology is inferred.

Conglomerates with large-scale crossbedding

(A more detailed description and interpretation of this facies unit is given in Chapter 4 in combination with the discussion of the basin-margin fan-deltas.)

Two types of conglomerate are present in the Pétervására Sandstone: 1. large-scale trough crossbedding in coarse to very coarse-grained sandstones with granular and pebbly foresets and 2. thick conglomerates consisting of large pebbles to granule with steep megaforesets. Within the field of sand waves both types are interbedded laterally as well as vertically. These conglomerate bodies attain a height of 3-8 m. The lower set boundaries of the pebbly trough cross-bedding are obviously erosive. However, in the thick conglomerates they are not erosive. Topsets or bottomsets are not present in either type. The coarse pebbly material always contains a great amount of bioclasts, i.e., coarse debris of shells which have lived near to the coast (cf. Báldi, 1983).

In the large-scale trough cross-bedding a high amount of pebbles is usually found at the base of large troughs. Strings of pebbles are aligned parallel with the coarse sandy foresets. Occasionally foresets up to a thickness of 45 cm are present, consisting of large pebbles within a coarse sandy matrix. In the large-scale trough cross-bedding deformation of coarse, pebbly foresets into folds, pipes and dishes (Fig. 4.5) also has occurred.

The thick conglomerates are composed of large pebbles-cobbles with a massive, coarse sandy, granular matrix at their base and they consist of granules at their top. This gives an overall graded appearance to the conglomerate bodies. Some megaforesets up to 10 cm in thickness also

show graded bedding. In other cases megaforesets are not characterized by apparent grain-size variations, but by differences in rate of cementation. Dip angles of foresets reach 30°, and dip directions are also dominantly towards the north.

Interpretation

Megaforesets in large conglomerate beds were partly formed by grain avalanching, as suggested by the presence of the coarsest grade at the bottom of the beds (cf. Nemec, 1990). Other foresets, with a well-developed sandy matrix, may have been formed by high-concentration mass-flows or minor slides of water-logged sediment (cf. Lowe, 1982). The megaforeset dip directions of the thick conglomerate lobes indicate that they were driven by the same tidal currents as the adjacent sand waves. The inertia of these large conglomeratic lobes must have been too large for average tidal currents and they most likely moved and accreted only under the most vigorous tides. The mixed-bioclastic composition points to a different origin for the conglomerates, than for the sand waves. This is also indicated by their position relatively close to the supposed palaeo-shoreline (Fig. 3.15; see also chapter 4).

The pebbles and the associated bioclastic material in the large-scale trough cross-beds must have been derived from erosion of the thick conglomeratic lobes described above. The foresets of the pebbly dunes were formed as loose grain avalanches (cf. Nemec, 1990). The hydrodynamic behaviour of the dominantly straight-crested sand waves must have been changed by the admixing of a significant amount of coarse, pebbly components (cf. Allen, 1984). Bedforms progressed towards a new equlibrium with tidal currents, and thus pebbly sinous-crested dunes developed.

Folds, pipes and dishes have been formed during synsedimentary water escape (cf. Lowe, 1975). Forceful dewatering occurred as the result of very rapid sedimentation and triggering by seismic shocks.

General facies model

Four facies units are distinguished in the Pétervására Sandstone. They appear in shore-parallel zones in successively deeper water. In the direction of the shore no shallow intertidal deposits were found. Instead a tectonically preformed, rocky shore with a steep and narrow pebbly beach is supposed to have been present, along which small fan-deltas provided some clastic material into the basin (Fig. 3.16; Chapter 4).

Relatively close to the shore (Chapter 4) at a depth of about 20 m an extended field of large, sligthly sinuous-crested sand waves was built by strong northward-directed tidal currents. Not too many modern shallow marine equivalents can be found for the Pétervására sand waves. The tidal sand ridges (or sand banks) off the coasts of England, Belgium and Holland show a different internal architecture. Sand waves there are just minor bedforms superimposed on the

ridges (cf. McCave & Langhorne, 1982; Belderson, 1986). Similarly the models based on fossil examples of sand ridges also cannot be applied (cf. Bridges, 1982). Instead the Pétervására sand-wave field is interpreted as a tidal sand sheet (cf. Stride, 1982; Belderson, 1986). The best modern analogues are the fields of sand waves in the Torres Straight, Australia (Harris , 1988, 1991) and those in the English Channel described by Berné et al. (1988).

Studies of modern systems suggest that in relatively shallow offshore water a currrent velocity of 30-55 cm/s at 1 m above the bed, which is equivalent to 50-90 cm/s peak surface velocity, is neccessary to build sand waves (Belderson, 1986). Indeed, in the Pétervására Sandstone small sand waves (dm-scale dunes) occur in the next facies zone offshore of the large sand waves (Fig. 3.16). The same strong tidal currents, acting in deeper water and thus corresponding to weaker currents near to the sediment surface, generated dm-scale sand waves in the fine to medium sand there. This deeper, somewhat lower energy environment allowed the deposition of more suspended matter from the sedimentary column and incidentally allowed the activity of burrowing organisms.

Still further offshore deposition of silty sediments has been dominant.

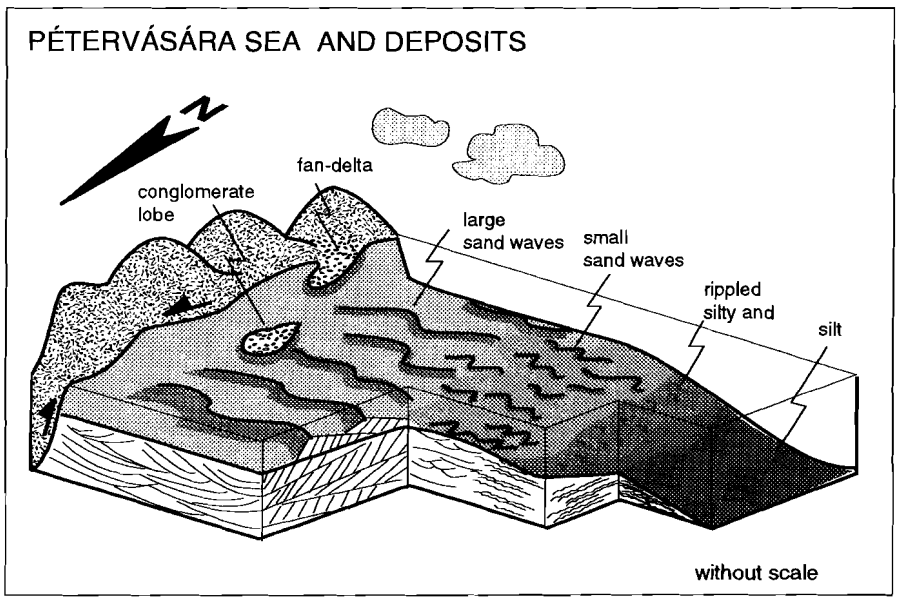

Fig. 3.16. Birds-eye view of the facies belts and the depositional environment of the Pétervására Sandstone.

CHAPTER 4.
SEDIMENTOLOGY AND PROVENANCE OF COARSE-GRAINED DEPOSITS ALONG A TECTONICALLY ACTIVE MARGIN OF A TIDALLY INFLUENCED EMBAYMENT, EARLY MIOCENE, NORTHERN HUNGARY

Abstract

The early Miocene Darnó Conglomerate and the Pétervására Sandstone with conglomeratic intercalations, both cropping out in northern Hungary, were formed along the eastern margin of the shallow tide-influenced Eggenburgian sea. Their petrographic composition indicates a close relationship, as does the Loibersdorf-type mollusc assemblage found in both formations.

The Darnó Conglomerate was a small fault-controlled fan-delta (underwater alluvial delta-cone) supplying coarse clastics into the "Pétervására Sea". The coarse-grained clastics, which were admixed to the material of the Pétervására Sandstone, have been derived from the Meliata nappes (Triassic-Jurassic ophiolite series) east of the Darnó Fault zone. The roughly northward shore-parallel transport is revealed by the orientation of the large bedforms of the Pétervására Sandstone, and also reflected by the distribution of certain ophiolite-derived heavy minerals and pebbles. Their increasing accumulation along the northern part of the coast indirectly infers a contemporanuous left-lateral displacement along the Darnó Fault. A reduction of the amount of less resistant pebble components of ophiolite-related origin and a relative enrichment of the resistant components towards the west indicates that part of the pebbles was transported in the offshore direction.

The conglomerates in the Pétervására Sandstone occur as small lobes interbedded with units consisting of tidally-driven sand waves. The composition of the lobes indicates that they were formed as small spit-platforms attached to the fans of the Darnó Conglomerate. As base level rised they became drowned and their sediment was washed into the basin and reworked by the strong northward-directed tidal currents into elongated lobes.

Introduction

Two lower Miocene conglomeratic successions are exposed in northern Hungary adjacent to the Darnó fault zone (Fig. 4.1). The Darnó Conglomerate (Báldi, 1983, 1986) is found only in a small area along the Darnó Fault (Fig. 4.1, definition of this structural element after Fodor et al., 1992). The tide-influenced Pétervására Sandstone is known from a relatively large area west of the Darnó Fault. A sedimentological field study (Chapter 3) revealed that mollusc-bearing pebbly

horizons within the Pétervására Sandstone show a close resemblance to the Darnó Conglomerate in many aspects. It was also noticed that coarse-grained intercalations are not confined exclusively to the top of the Pétervására Formation (Fig. 4.2, Chapter 3 and 8), as was assumed previously (Báldi, 1983, 1986). Pebbly intervals are equally abundant in the lower parts. The large-scale sedimentary facies distribution within the Pétervására Sandstone suggests that the Darnó Fault defined the paleo-shoreline of the shallow tide-influenced sea during the early Miocene. This implies synsedimentary tectonic activity during deposition of the Pétervására Sandstone and the Darnó Conglomerate.

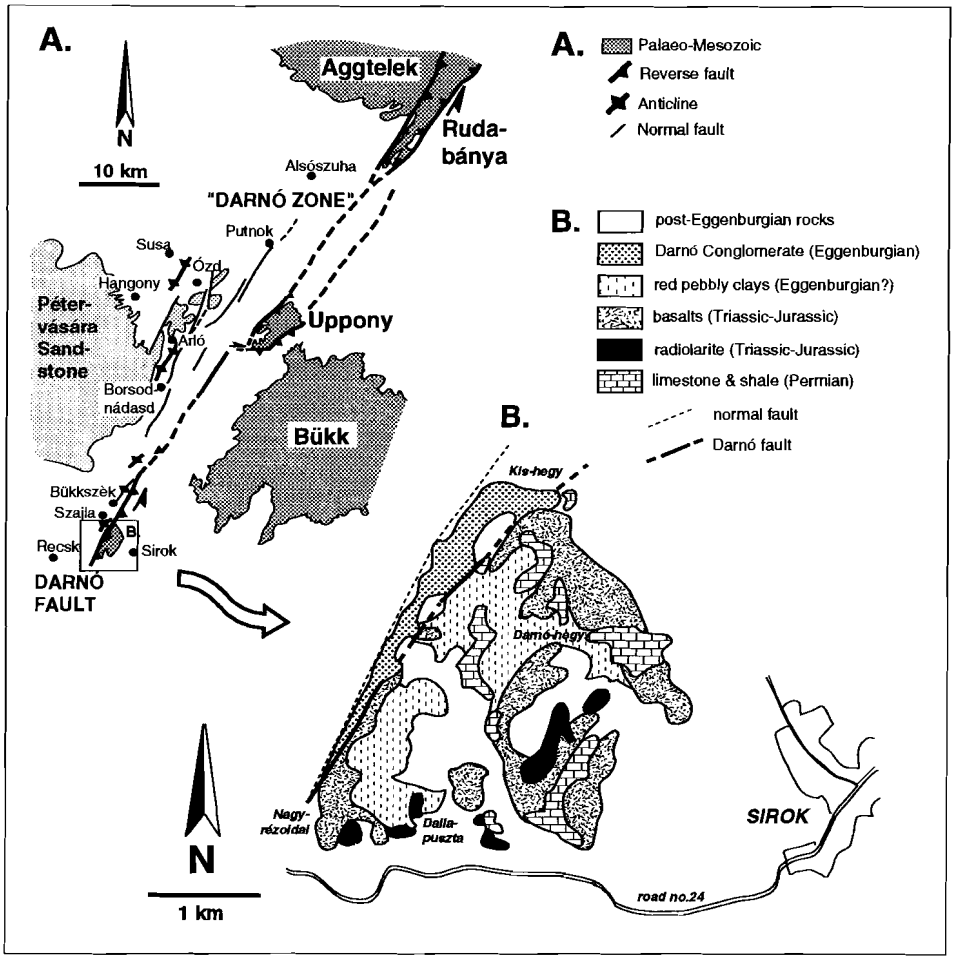

Fig. 4.1. A. Spatial relationship of the Pétervására Sandstone and the Darnó Conglomerate along the Darnó zone in northern Hungary (after Fodor et al. unpublished report, 1992). Only the southern segment of the fault zone is called Darnó Fault. B. Geological map of the Darnó hill, the only place where the Darnó Conglomerate is exposed (modified after Kiss, 1958b and Félegyházi & Vecsernyés, 1969).

The aim of this paper is to analyse and compare facies, depositional environments, spatial relationships and petrographical characteristics of the pebbly parts of the Pétervására Sandstone and the Darnó Conglomerate, as well as the influences of tidal currents and tectonic processes during their formation.

Geological setting in a historical perspective

The Pétervására Sandstone and the coeval Darnó Conglomerate were deposited in the "Hungarian Palaeogene Basin" (Báldi & Báldi-Beke, 1985), which existed in the intra-Carpathain area from the Eocene to the early Miocene. It was part of a series of epicontinental marine basins behind the Carpathian orogenic belt (Chapter 2). The Pétervására Sandstone, bounded by the Darnó Zone (cf. Fodor et al. unpublished report, 1992) in the east, conformably overlies bathyal siltstones (Szécsény Schlier; Báldi, 1986). The Darnó Conglomerate disconformably overlies basement rocks (Kiss, 1958a) uplifted along the Darnó Fault at the eastern margin of the basin (Fig. 4.1). During the early Miocene (Eggenburgian), when the studied sequence was formed, deposition occurred at shallow depth and the basin gradually was filled up to sea level (Sztanó & Tari, 1993, Chapter 8). Terrestrial beds form the top of the Eocene - early Miocene sedimentary sequence (Fig. 2.9; Báldi, 1986).

The Pétervására Sandstone

The Pétervására Sandstone is exposed in northern Hungary and South Slovakia (Fig. 4.1). It was mapped first as Chattian (Late Oligocene) glauconitic sandstone (Schréter 1929, 1940, 1942a, 1942b, 1948, 1953; Majzon, 1942; Tomor, 1948a, 1948b; Bartkó, 1952; Hegedüs, 1952; Jaskó, 1952; Radnóty 1952; Varga, 1975). Three different lithotypes were distinguished, as well as an areal repetition of such units. This repetition, however was explained as the result of post-depositional tectonic displacements (Szentes, 1943). The glauconitic sandstone is also known from South Slovakia, where Vass et al. (1988) distinguished six members.

Based on sedimentological features four laterally connected shore-parallel facies units with both vertical and lateral repetition were recognized in the sedimentary architecture of the Pétervására Sandstone recently (Figs. 3.15 & 4.2; Sztanó, 1992a,b). During the first phase the depositional system was aggradational, and in the second phase it was progradational. This is thought to be the result of changes in rates of sedimentation and eustasy (Chapter 8). Deposition during the early phase occurred in the direct vicinity of the Darnó Zone (Fig. 4.1).

Studies of nannoplankton (Báldi-Beke, Nagymarosy in Báldi, 1983; Nagymarosy & Báldi-Beke, 1988) and of large molluscs in the conglomeratic horizons (Cs-Mezneric 1953, 1959; Báldi, 1983) indicate an Eggenburgian (early Miocene) age for the Pétervására Sandstone.

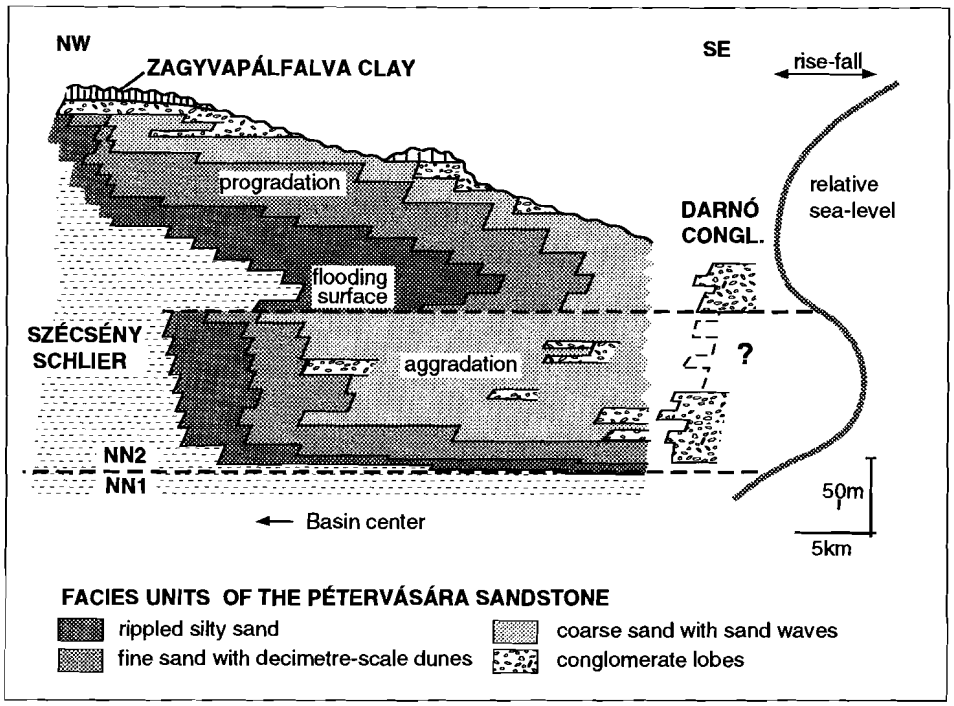

Fig. 4.2. Sedimentary architecture of the Pétervására Sandstone based on correlation of field sections and the deduced changes in relative sea-level.

The Darnó Conglomerate

The Darnó Conglomerate was mapped by Schréter (1940). Coarse conglomerates occurring between a post-Miocene normal and a reverse fault bounding the Darnó hill in the NW (Fig. 4.1) were described as littoral in origin and Burdigalian in age (Schréter, 1952).

Kiss (1958a) noticed that at the western side of the Darnó Zone a stratigraphic gap extents from the Triassic to the Eocene. In the fault zone, where the conglomerates disconformably overlie a diabase of Triassic-Jurassic age (Dosztály & Józsa, 1992), the gap continues up to the Early Miocene. The Darnó Conglomerate contains characteristic components, such as "diabase" clasts of various size and sharp, broken pieces of radiolarite and chert (Kiss, 1958a). All clastic sediments have been derived from the area east of the Darnó Fault, where these rock varieties occur *in situ* (Fig. 4.1). It was suggested that pebbles of the Darnó Conglomerate have been rounded locally by wave action and were not transported by fluvial processes (Félegyházi & Vecsernyés, 1969). Metre-scale cyclicity in the grain-size distribution was recognized, but not explained (Félegyházi & Vecsernyés, 1969). On the eastern side of the Darnó Fault red clays were found, which contain pebbles of siliceous schists and radiolarite, as well as pieces of subtropical silicified trunks (Greguss, 1956). These are presumably also Eggenburgian in age and terrestrial in origin, but none of these assumptions have been proved as yet.

Báldi (1983) focused on the faunal assemblage in the Darnó Conglomerate. The observed *Chlamys-Ostrea-Anomia* community clearly indicates an Eggenburgian age (Loibersdorf type fauna). *Balanus-Ostrea*-bearing strata reflect deposition in the littoral zone with highly agitated water, while *Ditrupa*-bearing horizons indicate depths around 20-30 m and/or more quiet conditions. Pebbles were regarded dominantly abrasive in origin, but fluvial transport of part of them was supposed (Báldi, 1983, 1986). Redistribution of the pebbly material by longshore currents was hypothesized. The occurrence of coarse-grained intercalations in the top of the Pétervására Sandstone in the adjacent Recsk area (Fig. 3.2, "Ilonavölgy beds" and ore-prospecting boreholes) were regarded to be the product of the above longshore currents. A palaeontological comparison of the Darnó Conglomerate and the "Ilonavölgy beds" indicated the same age for both formations. However no common mollusc species were found and the pebble composition in both deposits is different (Főzy & Leél-Őssy, 1985).

The Darnó Fault Zone

The Darnó Fault was first defined by Telegdi-Roth (1937) as the fault bounding the Darnó hill in the west (Fig. 4.1). Schréter (1942a) found small anticlines near Bükkszék, with lower Oligocene clay in the core surrounded by upper Oligocene deposits. These anticlines were produced by an ESE-WNW oriented compressional stress field. Minor folds were also described from the Lower Miocene glauconitic sandstone. In harmony with this compressional stress field, the Darnó Fault was interpreted as a Late Oligocene-Early Miocene reverse fault (Schréter, 1952).

Jaskó (1946) recognized a northward widening zone of long tectonic lines, with vertical and/or lateral displacement, which was named Darnó Zone (Fig. 4.1). Along this zone, small anticlines occur (Fig. 4.1, at Susa-Ózd Jaskó, 1952; at Arló Schréter, 1948; at Borsodnádasd Tomor, 1948a; Ózd-Tornaalja Schréter, 1953). Overthrusting of Palaeozoic and Mesozoic rocks over Oligocene deposits was recognized around Sirok (Schréter, 1952), but also Lower Oligocene clay was found above the early Miocene glauconitic sandstone (Tomor, 1948a), indicating that the compression continued after the early Miocene.

Geophysical surveys in the 70's and ore prospecting around Recsk revealed no significant difference in the structure of the basement at opposite sides of the Darnó Fault (Fig. 4.3). Zelenka et al., (1983) suggested a 20-30 km left-lateral displacement with a reverse component of the order of some kilometres. The activity of the line was dated to have been between the Late Eocene and the middle Miocene (Badenian) periods of andesitic volcanism, which produced the Mátra Mountains. The late Middle Miocene (Late Badenian-Sarmatian) left lateral wrench faulting of the Mátra Mountains was also attributed to the activity of the Darnó Fault (Balla & Havas, 1982). It was supposed that in the 8-10 km wide Darnó zone different lines were active at different times. The Darnó zone was assumed to continue through the Mátra to the south, where it joins a major tectonic element, the Balaton line (Fig. 4.3; Zelenka et al, 1983; Nagymarosy, 1990b). The Balaton

line, however, separates basement units of different origin (Transdanubian and Bükk), and was active as a right-lateral fault during the Oligocene and as a left-lateral compressive element during the Miocene. Therefore it cannot be the continuation of the Darnó Zone (Balla, 1989; Csontos et al., 1992). Recent field studies of microtectonical features in the Darnó Zone revealed left-lateral and reverse displacement during the early Miocene, left-lateral strike-slip around the Early/Middle Miocene boundary (Ottnangian - Early Badenian) and extension related normal faulting from the Middle Miocene (middle Badenian) onward (Fodor et al., 1992).

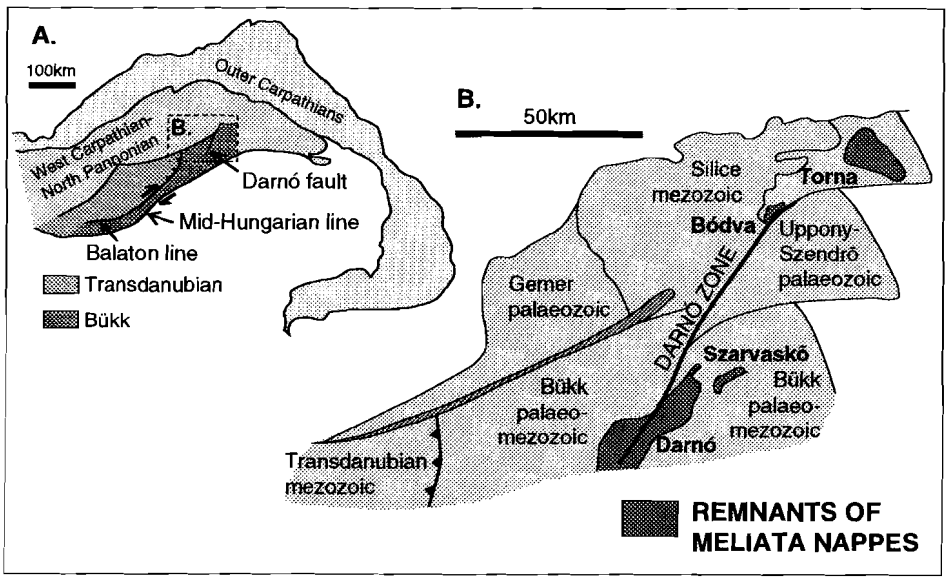

Fig. 4.3. A. Map of the main tectonic units and some important lines discussed in the text (after Csontos et al., 1992). B. The present distribution of Meliata oceanic nappe remnants above different basement units in northern Hungary and south Slovakia (after Csontos, in press).

Petrographic facies

Samples for the petrographic study were collected from conglomerates along the Darnó Fault, and from the conglomeratic intercalations and the adjacent sandstones of the Pétervására Sandstone (Fig. 4.4). The composition of rock fragments was studied by petrographic microscope and additionally heavy mineral assemblages were determined.

In the conglomerates and sandstones a characteristic index group of clastics (both rocks and minerals) was searched on the base of the following criteria:

- easy to determine with a simple petrographic microscope,
- not a common type of rock or assemblage (e.g. quartzite),
- sufficiently abundant in the samples,

- the index group should contain rock types of different hardness,
- the elements of the index group must be known from an area relatively close to the sedimentary basin. The potential source rocks should, morover, be characteristic and petrographically well known for comparative purposes.

The Darnó Conglomerate and the Pétervására Sandstone contain rock fragments and minerals, - radiolarites, hyaloclastites, (meta)basalts, (meta)dolerites, (meta)gabbros and serpentinites, - which fill the above requirements. In addition some pumpellyite-bearing rock types occur. In the heavy mineral assemblage actinolites with a narrow blue-amphibole zone are present. These rock types were identified in almost every studied sample. This suite of rocks and minerals represents ophiolite-derived material. The closest possible source of the ophiolites are the Meliata nappes, which occur at four localities in the neighbourhood: in the Darnó hill, in the Bükk mountains (Szarvaskő nappes), in the Bódva valley and in the Tarna unit (south Slovakia) (Csontos in press; Fig. 4.3).

Beside these ophiolitic rock fragments, many other rock types and minerals were determined, particularly in the Pétervására Sandstone. There are neutral-acidic volcanics (andesite, dacite, etc.), plutonic varieties (sienite, monzonite), rather special metamorphic rocks (chloritoid-bearing micaschist, clinozoisite-rich micaschist, rutile-bearing gneiss) and some sedimentary rock fragments (crystalline limestone, sandstone, siltstone, siliceous schists). This paper does not discuss the orgin of these rocks, because it is not possible to define the potential source areas with sufficient reliability.

The detritus of ophiolit related origin composes more than 80% of the Darnó Conglomerate, but less than 10% of the total clastic material in the Pétervására Sandstone. Nevertheless the similar ophiolitic detritus is present in both deposits, implying the same source. Significant trends in the amounts of ophiolitic fragments within the Pétervására Sandstone were detected in two directions. 1. The number of ophiolitic fragments decreases towards the northwest, with increasing distance from the Darnó Zone (Fig. 4.4). 2. Except for the Darnó Conglomerate, the highest amount of ophiolitic rock fragments was detected in the northernmost locality (3) and it decreases to the south along the Darnó Zone (Fig. 4.4).

A good correlation exists between transport distance and hardness of different rock types. Serpentinite appears only in those conglomerates (locality 1), which occur in the very Darnó zone. The most distal occurrence of metagabbro is 30 km NW of the Darnó Fault (locality 9 on Fig. 4.4). The most distal basalt grains (locality 10) were found ca. 35 km from the supposed source area, while the most persistent rock type of ophiolite-related origin - radiolarite - persists to the westernmost localities (locality 11, Fig. 4.4).

Two samples (from localities 3 and 9, Fig. 4.4) contain pumpellyite-bearing rock fragments. This type of pumpellyite is uniquely known from the Darnó-Szarvaskő area (Józsa unpublished data).

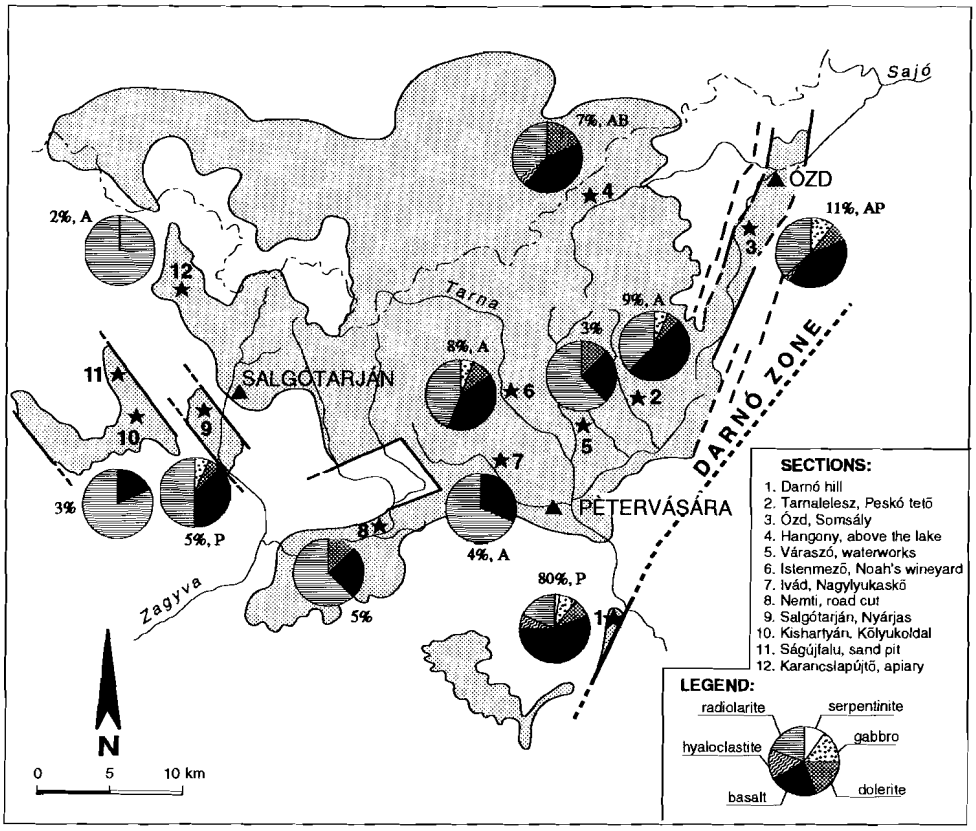

Fig. 4.4. Pie diagrams showing differences in composition of ophiolite-derived detritus at different localities. The percentages indicate the share of ophiolite-related clastics (volcanites and radiolarite together) in the total amount of clastic material. Between the heavy minerals some source-specific minerals were also found: pumpellyite (P), aktinolite (A) and blueamphibole (B). The share of volcanics decreases towards the younger units (9, 10, 12) in the west. In the Pétervására Sandstone, along the Darnó Zone, the amount of ophiolite-derived clastics increases towards the northeast. The Darnó Conglomerate (1) is mainly built up of ophiolite-derived clastics.

In the heavy mineral assemblage a few grains of actinolite were found (locality 4). In narrow zones of the actinolite grains transformation to blue amphibole has taken place. This form of blue amphibole in association with actinolites is known exclusively from the Bódva valley (Fig. 4.3), where only a very small amount of blue amphibole is present (Józsa unpublished data). Minor amounts of blue amphibole grains were also described from the Slovakian localities of the Pétervására Sandstone (Vass et al, 1988). The presence of blue amphibole is uniquely confined to the northern occurrences of the Pétervására Sandstone.

Potential source rocks: the Meliata nappes

In northern Hungary close to the Darnó Fault Triassic-Jurassic rocks of ophiolite-related origin form part of the Meliata nappes (Fig. 4.3; Csontos, in press and references therein). It is supposed that originally the same ophiolitic series was present over a broad area. In the Darnó hill (Fig. 4.3) mainly radiolarite, hyaloclastite, (meta)dolerite, (meta)gabbro and (meta)basalt appear (Földessy, 1975). Here the top of the ophiolitic series had been eroded. A similar ophiolitic suite is present in the Szarvaskő area in the Bükk mountains (Fig 4.3; Balla, 1984). A characteristic secondary mineral assemblage - chlorite, albite, pumpellyite, prehnite, actinolite, tremolite, brown hornblende and epidote - has been determined from these localities (Földessy, 1975; Balla, 1984). Only the pumpellyite is exclusively characteristic for the Szarvaskő locality, and thus is suitable to verify the source area.

Further to the NE, in the Bódva valley (Fig. 4.3) similar rock types are found (Réti, 1985). Here a distinction can be drawn because of the occurrence of serpentinite and a different type of metagabbro. Among the secondary minerals no prehnite and pumpellyite are present, but instead zoisite and blue amphibol are common.

In South Slovakia the Darnó-Bódva ophiolitic belt continues (Torna unit, Fig. 4.3; cf. Csontos in press;). Different members of the ancient oceanic series crop out locally, mostly in teectonic windows (Hovorka, 1985). In the northeastern part of the area a large amount of blue-amphibole bearing rocks occurs (Kameniczky, 1957).

The presence and composition of ophiolite-derived debris in the Darnó Conglomerate and the coarse-grained constituents of the Pétervására Sandstone provide evidence that only the Meliata nappes - either the Darnó-Szarvaskő or the Bódva-Torna series - may have served as source areas. The Bódva ophiolites cannot have been the source of the Darnó Conglomerate, because of the different type of metagabbro and the lack of pumpellyite. Serpentinite fragments, found only in the Darnó Conglomerate (i.e., on the Darnó Hill), suggest that the eroded upper part of the adjacent Darnó series was similar to the still existing upper part of the Bódva series. It is concluded that the clastic material was shed mainly from the Darnó-Szarvaskő nappes.

Sedimentary facies

The Pétervására Sandstone

The dominant sedimentary structures are various-scale cross-bedding, both in sandstones and in conglomerates (Fig. 3.3, Chapter 3). There is a fairly good correlation between grain size and dominant bedforms (Fig. 3.15). The most voluminous conglomerate bodies, with height up to 8 m, are situated 5-10 km west of the Darnó zone, within the field of sand waves (Fig. 3.3 and 3.15). They form finger-like "intrusions" and interrupt the trend of westwardly decreasing grainsize. Foresets of large-scale crossbedding both in sandstones and in conglomerates dip towards the north (Fig. 3.15).

The sedimentary features of the Pétervására Sandstone are inferred to be the result of a deepening of the basin towards the NW. Grainsize and dimensions of bedforms decrease in the offshore direction. Large-scale cross-bedded sets in sandstones are interpreted as tidal sand waves (cf. Allen, 1980), driven northward by strong ebb currents (Chapter 3). Close to the supposed palaeo-shoreline a fair amount of gravel-sized material occurs in the bottomsets and in the foresets of sand waves, and also sets consisting purely of conglomerates appear. Coarse-grained material always contains a great amount of bioclastics, mainly shell debris.

Large dunes with pebbly foresets

Some sand-wave-like bedforms consist of coarse to very coarse-grained sandstones with granular and pebbly foresets. These are regarded as pebbly dunes. The dip angle of foresets is around 20°. Most of the pebbles are aligned roughly parallel to foresets. High concentrations of pebbles are found in the troughs at the toe of the foresets. Some foresets, up to a thickness of 35-45 cm, are completely made up of large pebbles with a coarse sandy matrix. For the depositional mechanism avalanching is supposed. Large front-runner clasts accumulated at the base of beds (cf. Nemec, 1990).

Although the sand waves are dominantly straight-crested, the pebbly dunes are linguoid, as appears from the characteristic large-scale trough cross-bedding. The best examples of this facies are found in a roadcut near Nemti (Fig. 3.3, section 8), at Ózd and Kishartyán (Fig. 3.3, section 10). Another remarkable difference between sand waves and pebbly dunes is that foresets are not arranged cyclically in the pebbly dunes, although tidal cyclicity was observed in many sand waves. This implies either a pulsating discharge of coarse pebbles from nearby sources, or it indicates that the transport capacity of tidal currents was commonly too small for the coarse material. It seems that the sand waves were moved regularly by oscillating tidal currents, but that pebbly dunes moved only during the most vigorous spring-tide currents.

Rapid and large influxes of coarse-debris are also indicated by spectacular water escape structures. This kind of deformation is found at several localities, but the largest (up to 4 m high) occur in a pebbly dune south of Ózd (Fig. 3.3, section 3). Both vertically and laterally, zones of massive gravelly sand with irregular lower and upper boundaries alternate with zones of strongly deformed foresets, in which escape pathways can roughly be followed (Fig. 4.5). Structures resemble dishes and pipes as described by Lowe (1975) of extraordinarily large size and irregular spacing. Nevertheless the original bedding of these sediments can still be recognized. The deformational structures indicate that pore-water migrated along the foresets before bursting out. Rearrangement of grains was confined only to areas where the semi-consolidated sediment completely lost its strenght and mixed with the liberated pore-fluid.

Grainsize of foresets ranges from coarse-grained sand to granules, and up to large pebbles. Sorting is fairly good within individual foreset beds, which have a relatively large intergranular space. No clay-sized material is present, which could have blocked the free movement of the pore-

water (cf. Allen, 1984) and contributed to the generation of overpressure. In the range of grain sizes which are present, intergranular water moves freely. The forceful dewatering therefore is explained as the result of very rapid sedimentation and tectonically triggered liquefaction.

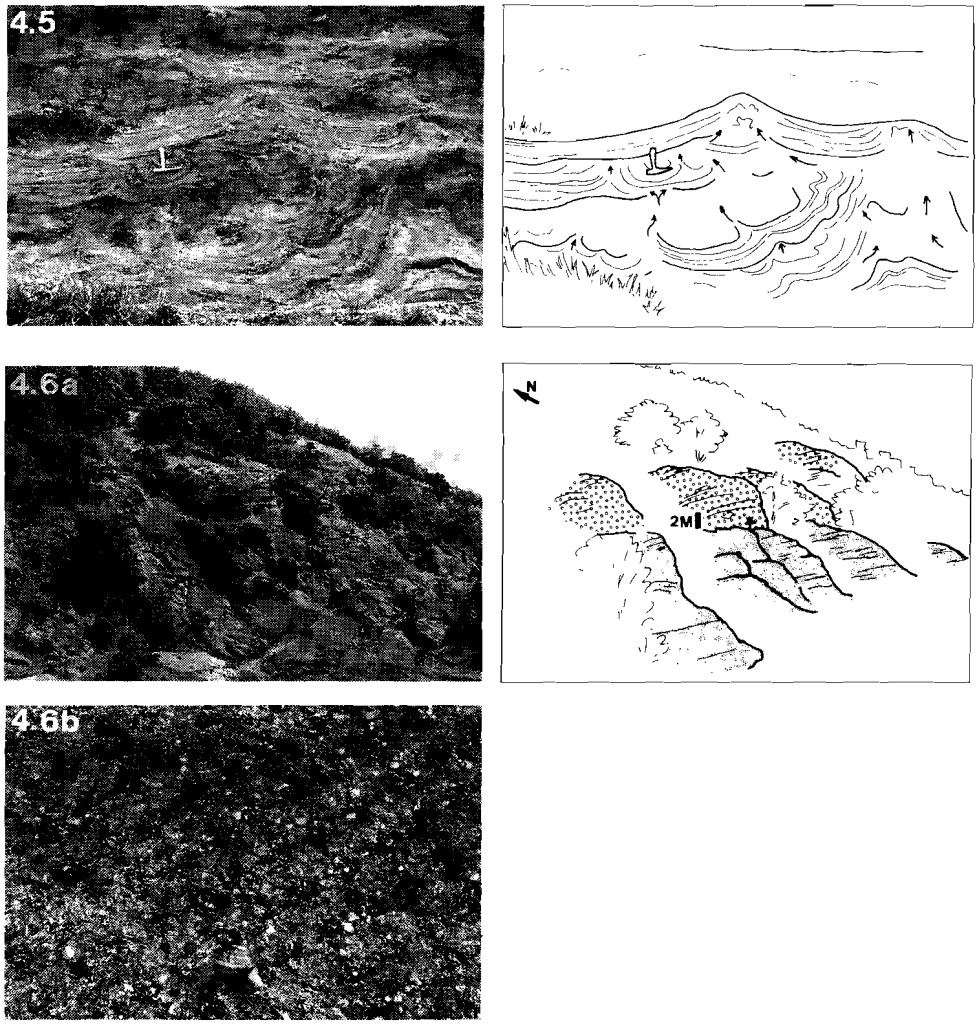

Fig. 4.5. Foresets in coarse to granular sand are strongly deformed by large-scale water-escape movements near Ózd. Arrows indicate escape pathways of the pore-fluid.

Fig. 4.6. A. Large conglomeratic dune at Peskó-tető north of Tarnalelesz in the Pétervására Sandstone. Megaforesets are featured by different rates of cementation. B. Close-up of the base of the conglomeratic dune, marked by an asterix on the line-drawing. Pebbles float in a very coarse sandy matrix. For scale see the 5 cm long snail.

Thick conglomerates with megaforesets

Within the sand-wave field some thick conglomeratic bodies appear (Fig. 4.6). From a distance most of them resembles sand waves, having megaforesets with the same dip to the north as do the sand waves. Dip angle of foresets reaches 30°. The thickness of the sets however is much greater than that of the under- and overlying sand waves, and goes up to 8 m (Fig. 4.6). The lower and the upper set boundaries do not show signs of erosion. The size of gravel generally decreases upwards in the sets and ranges from small cobbles at the base (Fig. 4.6) to granules at the top of the sets.

Bottomsets made up of fine sand or mud are rare, and their thickness does not exceed 20 cm (Fig. 3.3, section 7 and 12). At other localities (Fig. 3.3 section 2 and Drna in Slovakia) fine-grained bottomsets are completely lacking. The fabric of the basal conglomerates is usually massive with a very coarse sandy matrix (Fig. 4.6). The foresets are similar: coarse to very coarse sand with floating pebbles, which decrease upwards in concentration and size. Sometimes it is difficult to recognize the individual foresets, because of the rather homogeneous grain population. The long axes of pebbles occasionally show weak foreset-parallel orientation. There is no evidence for the presence of topsets. Upon the conglomeratic dunes ordinary sand waves occur, with presumed minor erosional truncations.

Foresets of the conglomeratic dunes in the Pétervására Sandstone either were formed by loose grain avalanches (cohesionless grain falls) as is shown by the large front-runner clasts or as high-concentration grain flows (cf. Nemec, 1990), in which larger pebbles were transported in the sandy matrix. The slope-parallel internal shear is reflected by the bed-parallel pebble orientation.

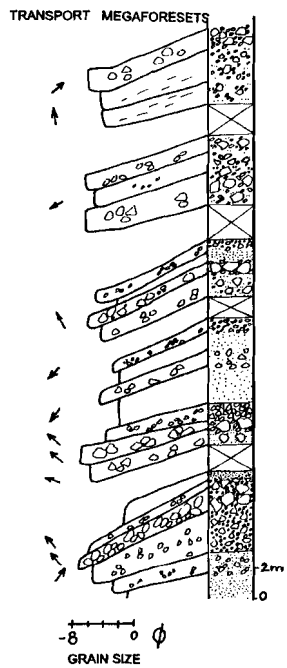

Fig. 4.7. Sedimentary log of the Darnó Conglomerate in an small valley south of Kis-hegy. Arrows show transport directions as indicated by dip direction of megaforesets.

The Darnó Conglomerate

Steeply-dipping conglomerate beds crop out only in a narrow zone along the Darnó Fault (Fig. 4.1). The thickest vertical section observed was 20 m, and occasionally beds could be followed in the dip direction over 70 m. Two profiles and some small outcrops (Fig. 4.7) allowed the reconstruction of the architecture of the conglomerate.

In general the conglomerates are badly sorted with grain sizes ranging from very coarse sand to cobbles. Some of the beds are polymodal clast-supported conglomerates. The majority, however, is bimodal with clusters of gravel floating in a very

coarse sandy matrix (Fig. 4.8). Fabric ranges from pebbly sand to an almost closed, clast-supported gravelly fabric with minor amounts of sand in between. Normal grading was not observed, but faint inverse grading was recognized in several beds. Coarsening and fining upward successions are difficult to determine (Fig. 4.7).

The shape of the particles is determined by their parent rock. Many of them are flat (red slates, radiolarite, various kinds of schists and even diabase). Roundness is highly variable. It is mainly fairly poor. Angular pieces of parent rocks are common (Fig. 4.8). Well rounded pebbles and cobbles were found only in a few beds (Fig. 4.7). Clusters of flat pebbles usually show a weak imbrication, dipping upslope. Many pebbles are aligned subparallel to the bedding and dip downslope with a 10° greater angle than the dip of the beds (Fig. 4.8). Some well rounded pebbles have a smooth polished surface of wind-blown origin, which indicates a semi-arid climate in the source area and only very short subaquaous transport.

Beds of conglomerates dip dominantly towards the northwest with a wide deviation. Dip angles vary between 10-40°, but usually they are around 30°. Low-angle differences between individual strata of 0.2-1.5 m thick are common. A solitary package of oppositely (SE) dipping beds was observed in the Kis hegy section (Fig. 4.9).

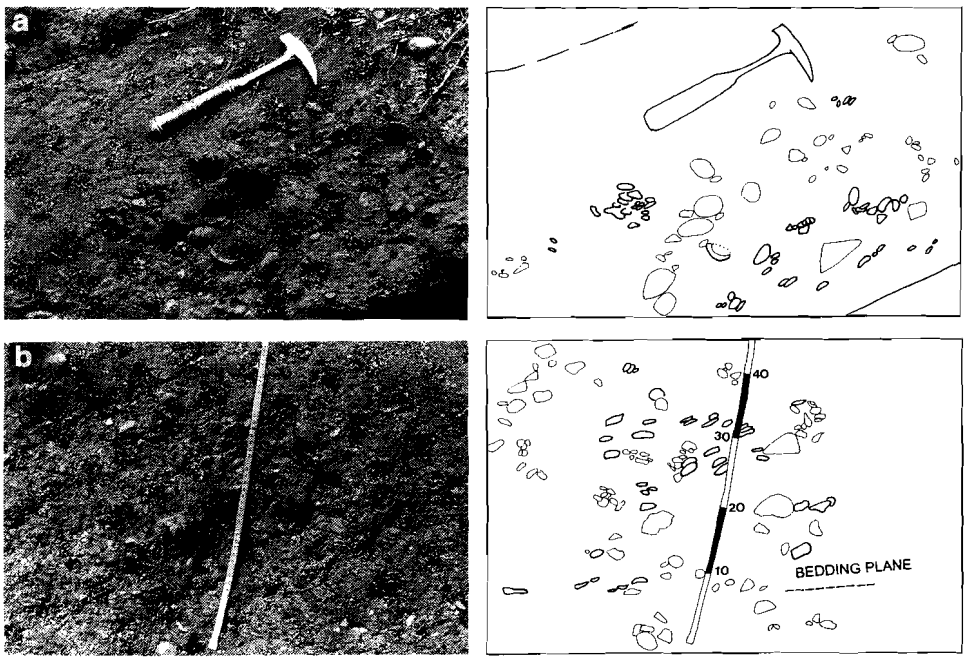

Fig. 4.8. Fabric of the Darnó Conglomerate. A. Megaforesets consisting of matrix-supported coarse conglomerates with subrounded pebbles. Note imbrications in pebble clusters and subparallel arrangement of other pebbles. Transport from right to left, i.e., from E to W. B. Disorganized polymodal conglomerates with angular pebbles. Transport to the left.

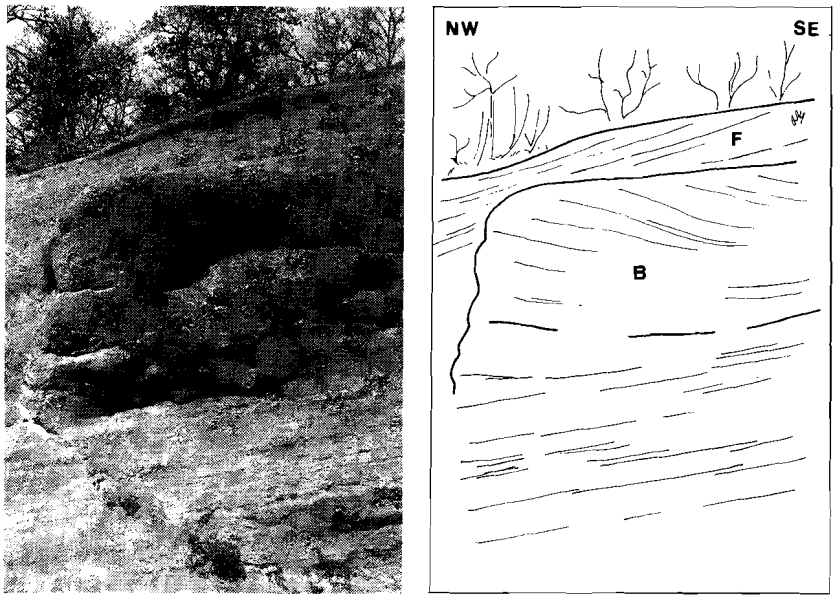

Fig. 4.9. Between the westward dipping megaforesets (F) some beds dip in the opposite direction at Kis hegy (Darnó hill). These beds are interpreted as backsets (B), (cf. Postma, 1984).

Internal erosional surfaces between steeply dipping beds were recognized only once. This erosional surface was found in a small outcrop on the Darnó hill, and cuts asymmetrically into beds of extremely badly sorted sands and matrix-supported conglomerates. Above this erosional surface grain size increases and dip angle decreases downslope, indicating the gradual infilling of a scour.

The amount of mollusc shells, either as broken pieces or fairly completely preserved, is fluctuating throughout the Darnó Conglomerate. Their presence forms good evidence of a marine origin of the majority of the conglomerates.

Considering the above features, the pebbly sands and conglomerates must have been deposited by subaquaeous gravity-flows on the steep slopes of a delta. The great dip angle, the lack of normal grading and the varying amount of sandy matrix infers high-concentration debris flows or grain flows (cf. Lowe, 1982; Nemec, 1990). Downrolling of a loose assemblage of solitary particles is not likely, because down-dip segregation or front runner clasts were not found. Clusters of large clasts, concentrated in the upper parts of the sandy matrix, and the subparallel arrangment of flat pebbles infers a certain strenght and internal shear of the transporting medium. The lack of mud-sized material increased slope stability (cf. Colella et al., 1987). Slumps were not found. The erosional surface described above, however, is interpreted as a slump-scar filled up by progressively coarser conglomerates. The upslope dipping package of strata (Fig. 4.9) represent backsets (cf. Postma, 1984) and also points to slope instability (cf. Colella et al., 1987; Nemec, 1990).

The depositional system of the conglomerates: a discussion

The Darnó Conglomerate

The primary sedimentary structures of the Darnó Conglomerate (e.g. steep megaforesets, mass-flow deposits, backsets, etc.) indicate that they were formed on steep slopes of a small coarse-grained delta. The exposures do not allow to determine the complete thickness of a delta cycle. Palaeo-waterdepth, however, can be estimated by means of palaeoecology. The greatest depositional depth indicated by molluscs, is between 20-30 m (Báldi, 1983, 1986). Bedforms in the offshore direction (e.g. sand waves in the Pétervására Sandstone) support this estimate (Chapter 3). Basin depth thus must have been of the order of tens of metres. So the Darnó Conglomerate represents a shallow-water delta, with a steep Gilbert-type profile fed most likely by a steep, gravelly alluvial cone or by a line source of unstable bedload channels of alluvial fans (cf. Postma, 1990).

Outcrop conditions prohibit the reconstruction of the delta morphology. The "distal" part of the delta (bottomsets) was cut by a mid-Miocene fault (Fig. 4.1). Moreover, neither topsets nor the alluvial parts of the delta are exposed (unless the red clays of presumably the same age represent topsets, Fig 4.1). These may have been removed as the result of later fault activity and uplift (cf. Fig. 4.10). Conical deltas, however, usually precede the development of Gilbert-type deltas, which commonly have a subaerial delta plain and thus also topset beds (Nemec, 1990).

The Darnó fan-delta relied on basin-margin fault scarps of the Darnó Fault, which, as is indicated by many features, was active during deposition. Altough this kind of deltaic setting is sensitive to tectonically induced base-level changes, which normally cause the stacking of delta cycles (cf. Gawthorpe & Colella, 1990), downlaps of megaforesets were not found in the Darnó Conglomerate. Instead the tectonic activity is most obviously shown by the distribution of specific assemblages of pebbles and minerals in the basin (to be discussed below). The changes in megaforeset dip directions and angles was caused by lobe switching. If a linear source had been present along the margin, subsequent beds with 90° difference in dip direction (Fig. 4.7) would have received their material from adjacent distributaries along the Darnó Fault.

It was supposed that the Darnó Conglomerate, which overlies a Triassic-Jurassic ophiolitic series, was produced by "the Burdigalian/Eggenburgian sea-level rise", flooding the former shore (Kiss, 1958a; Félegyházi & Vecsernyés, 1969; Báldi, 1983, 1986). Signs of intense relative sea-level rise, however, are present only in the middle part of the depositional sequence of the coeval Pétervására Sandstone (Fig. 4.2 and Chapter 8). Instead, the occurrence of the coarse clastics, i.e., the Pétervására Sandstone and the Darnó Conglomerate, is explained to have resulted from a relative sea-level fall just preceeding the Eggenburgian sea-level rise (Chapter 8). Active deltaic deposition should have prompted at the same time, i.e., following the marked base-level fall. The conglomerates in the lower, aggradational part of the Pétervására Sandstone (Fig. 4.2) indicate that sources of coarse clastics were indeed present along the Darnó Fault already at that time.

In concert with the above data the following hypothetical model can be set for the development of the Darnó Conglomerate (Fig. 4.10). The relatively drastic base-level fall (prior to the Eggenburgian, Chapter 8) induced stream incision into the surroundig dry land east of the Darnó zone (Fig. 4.10). Incision could continue until an equilibrium-graded stream-profile was reached (cf. Posamentier & Vail, 1988). During the subsequent slow base-level rise (Chapter 8) the alluvial valleys may have been partly flooded, and thus some conglomerate could be deposited as valley fill. However every step of tectonic uplift (greater than that of relative sea-level rise) promoted further incision and cannibalization of valley-fill deposits (Fig. 4.10). This maintained erosion and transport of gravel-sized material into the basin. These processes explain the relatively coarse debris, derived almost exclusively from sources in the Darnó zone.

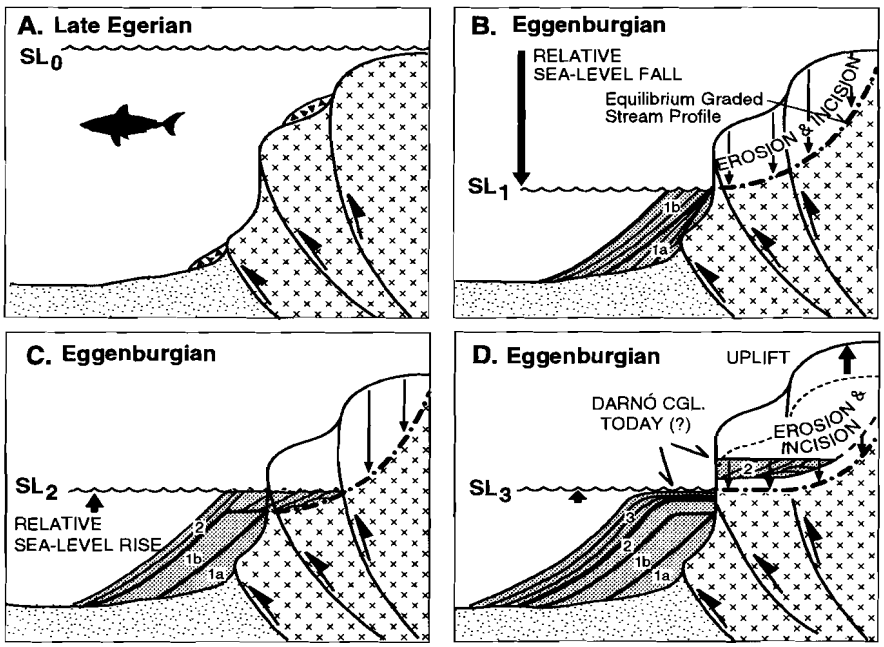

Fig. 4.10. Hypothetical model for the development of the Darnó Conglomerate. B. Development of a conical and a Gilbert delta as a result of a significant base-level fall and valley incision. C. Flooding and partial fill of valleys during base-level rise. D. Erosion and new incision during tectonic uplift of the hinterland, cannibalizing sediments formerly deposited in the alluvial valleys. The relative rates of sea-level rise and tectonic uplift determined whether erosion or deposition dominated.

The presence of a voluminous fluvial or alluvial feeder system along the Darnó Fault can be excluded. If a major feeder system had been present, it would have filled the adjacent shallow basin rapidly. However, the amount of coarse clastics, which arrived from the Darnó region to the Pétervására Sandstone in the basin, is subordinate to the amount of somewhat finer clastics of different origin, which composes the majority of the Pétervására Sandstone (see petrographic facies

in this chatper and Fig 4.4.). This latter type of material, which is not present in the Darnó Conglomerate, was derived from other source areas which must have occurred in the south (cf. Fig. 2.8).

The Pétervására Sandstone

The conglomerates in the Pétervására Sandstone are lobate bodies with steep slip faces interbedded within the field of sand waves (Fig. 3.3, 3.15 and 4.2). They built out from the SE to the NW, but their fronts faced northward, parallel with the main transport pathways determined by the strong tidal (ebbs) currents (Chapter 3). Petrografically the conglomerates within the Pétervására Sandstone closely correspond to the Darnó Conglomerate, and they must have received part of their material from the same area. Therefore it is inferred that originally the conglomerate lobes have been attached to the shore and/or the fan deltas of the Darnó Conglomerate (Fig. 4.11).

It is highly unlikely that these small lobate conglomerates were active delta lobes, because subaerial delta plains were found nowhere, although looked for. In addition such delta plains should have been relatively long and narrow to be continuously connected with the main delta along the shore. Being relatively small and elongate the lobate conglomerate bodies also resemble certain coarse-grained spit-platform deposits as described by e.g. Nielsen et al., (1988). The spit-platforms, however, are also connected to beach deposits and occur below mean low tide level. They have steep subaquaous depositional slopes and topsets with subaerially exposed spit ridges. The formation of the spit platforms is determined by wave-action-related longshore sediment transport. It should, however, be noted that neither the formation of small deltas nor of spit-platforms is likely to happen when the tidal range is relatively large (Nielsen et al., 1988), like the value of about 4 m, which was estimated for the "Pétervására Sea" (Chapter 5).

Based on the size of the adjacent sand waves, depositional depth is estimated to have been around 15-20 m (cf. Allen, 1984 Chapter 3). Therefore the tops of the 6-8 m high lobes must always have been covered by a few metres of waters. Tide-induced sediment transport in the "Pétervására bay" was subparallel to the shoreline along the Darnó Fault and was directed towards the north. Therefore the conglomeratic lobes must have slowly migrated in the same direction driven by the strong ebb currents. Such longshore propagating tidal currents thus would have reworked parts of the fan-deltas of the Darnó Conglomerate, particulary during low-stands of the sea-level, and built spit-like lobes attached to the Darnó fan-deltas (Fig. 4.11). Indeed deltaic bodies in environments with strong tidal influence often are highly elongated and asymmetrical (e.g. Copper river delta, Galloway, 1976). Later, when relative sea-level rose, these spits must have been drowned, and become large subaquaous conglomerate bodies. As separate bedforms, these must have become subject to the strong, roughly shore-parallel currents. Thus they started to migrate obliquely offshore due to the drag of the tidal currents. Gradually their material was mixed with the sand brought in by the currents from the south, and they were incorporated into the sand-

wave field (Fig 4.11). Part of the pebbly material may have been eroded from the lobes and been transported into the inter-sand wave troughs, where it was distributed farther offshore. These "escaped" pebbles either remained trapped between the sand waves and got burried gradually or were transported by peak tidal currents up to stoss slopes of the sand waves and then deposited as pebbly foresets.

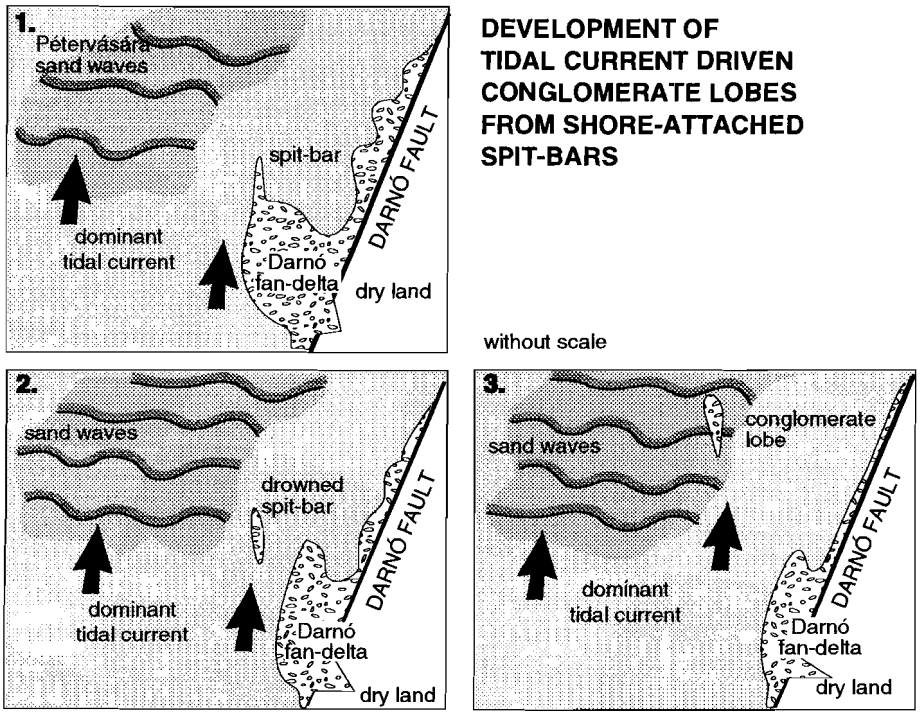

Fig. 4.11. Model of the development of conglomerate lobes in the Pétervására Sandstone. See text for explanation.

The role of the Darnó Fault

The pebbly material shed into the basin was roughly the same all along the Darnó Fault. At present these rocks are known only from a relatively narrow region east of the tectonic zone. However, it is supposed that during the early Miocene the Bükk area was covered with nappes consisting of the above rock varieties of ocean-floor origin (Fig. 4.3; Meliata-Szarvaskő nappes, Csontos in press).

The above described sedimentary and petrographic observations indicate that the Darnó Fault was active during the Early Miocene. The Darnó Fault determined the position of the palaeo-shoreline. Grain size of the sediment and the amount of volcanites decrease with increasing

distance from the Darnó Fault. The most important observation, however, is the northward accumulation of the ophiolite-relatad detritus along the paleo-shoreline (Fig. 4.4). This indicates that the strong shore-parallel tidal currents, which forced the lobes to migrate from south to north, spread their sediments in a northward direction (Fig. 4.12). Therefore all sediment which arrived into the basin with these conglomerate lobes must also have been transported to the north by the tidal currents. When the source area and the marine basin shifted laterally with respect to each other, due to the lateral strike-slip of the Darnó Fault, every "new" influx of sediment arrived into the basin north of the previous lobe (Fig. 4.12). In this way bed-load accumulation in the northern part of the basin was further promoted by the activity of the basin-margin faults. This interpretation of the northward accumulation of the ophiolite-derived clastics, implies a left-lateral strike-slip activity of the Darnó zone during the Eggenburgian (Fig. 4.12).

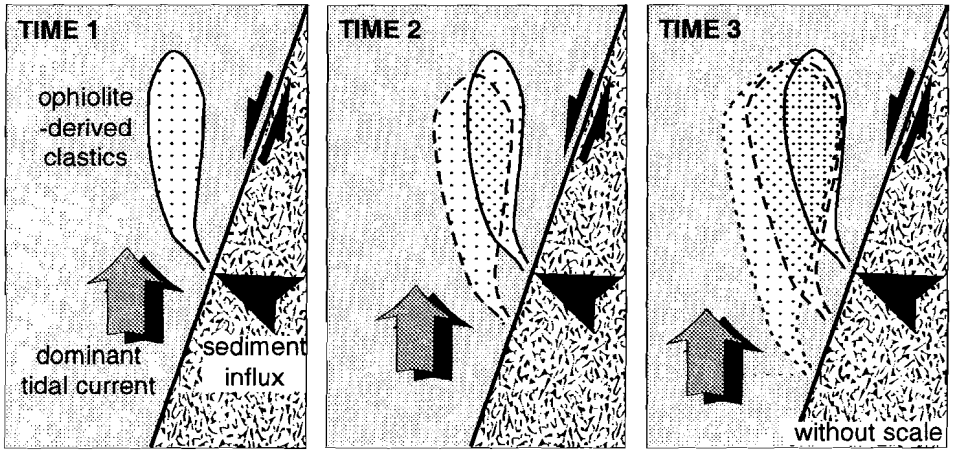

Fig. 4.12. Model of accumulation of ophiolite-derived clastics in combination with the left-lateral strike-slip along the Darnó Fault. All sediment was transported from south to north by the tidal currents. Moreover, due to the strike-slip activity, every younger lobe is situated northward of the previous one, thus resulting in a northward increase of the amount of ophiolite-derived detritus.

Conclusions

The Darnó Conglomerate of Eggenburgian age was deposited as a small coarse-grained (Gilbert-type) fan-delta system along the Darnó Fault in response to a marked relative sea-level fall and to tectonic activity of the fault during the Eggenburgian. The bulk of the pebbly material was derived from the subaerially exposed Meliata-Szarvaskő nappe, east of the Darnó Fault, consisting mainly of Triassic-Jurassic ocean-floor deposits, like basalts and radiolarites.

The coeval Pétervására Sandstone contains the same faunal assemblage and was deposited offshore, west from the Darnó Conglomerate and the Darnó Fault. The Pétervására Sandstone

primarily consists of sand, derived from the south. The sandy material was mixed with coarse-grained components. The petrographical composition of this coarse material is basically the same as that of the Darnó Conglomerate, inferring the same source area. A reduction of soft components (volcanites) and an enrichment of resistant ones (radiolarites) towards the NW indicate a transport route of pebbles away from the Darnó Fault into the basin of the Pétervására Sandstone. The pebbly material was transported slightly offshore by strong, almost shore-parallel tidal currents, and deposited as lobate conglomerate bodies with steep subaquaous megaforesets. These lobes originally must have been attached to the basin-margin fan-deltas, but after drowning they were moved northward by the strong tidal currents. An increase of the amount of ophiolite-derived material to the north provides indirect evidence for the left-lateral strike-slip activity of the Darnó zone during the early Miocene (Eggenburgian).

CHAPTER 5
TIDAL CYCLICITY IN THE SAND WAVES OF THE EARLY MIOCENE PÉTERVÁSÁRA SANDSTONE

Introduction

During the last two decades studies of modern and ancient tide-influenced sediments revealed a number of criteria for the recognition of such systems. The daily, monthly and even yearly changes in the relative position of the Earth, Moon and Sun govern variations in height of the tides through changes in gravitational attraction. Thus the velocity of tidal currents, the rate of sediment transport and sediment accumulation are influenced at these time-scales. Sedimentary features reflecting the periodical changes in strength and direction of currents include sand-mud rhythmites, mud drapes and mud couplets (Visser, 1980; Allen, 1981), reactivation surfaces or pause planes in cross-beds (Visser, 1980; Boersma & Terwindt, 1981), sigmoidal cross-bedding (Kreisa & Moiola, 1986), and systematic changes in shape or inclination of foresets (Boersma & Terwindt, 1981; Siegenthaler, 1982; Allen & Homewood, 1984; Terwindt, 1988; Houthuys & Gullentops, 1988a,b; Davies and Flemming, 1991).

The most convincing evidence of tidal influence during deposition is given by laterally or vertically accreted sediments, in which cyclic thickness variations correspond to semi-daily, fortnightly or even yearly tidal periodicities. Such deposits are known from Precambrian to Recent times. Well developed cyclicity of tidal origin was found by measuring thicknesses of foresets in large dunes (Visser, 1980; Allen, 1981; Homewood & Allen, 1981; Allen & Homewood, 1984; Kreisa et al., 1986; Santisteban & Taberner, 1988; Lapido, 1988; Uhlir et al., 1988 Kvale & Archer, 1991). Usually series covering periods only up to some few months are revealed in this way, due to the great size of the bedforms and the limited length of the exposures. More complete records up to several years are obtained by measuring thickness variations of sand-mud couplets of tidal flat or abandoned channel fill origin (Tessier & Gigot, 1989; Kvale et al., 1989; Dalrymple et al., 1991; Williams, 1991; Kvale & Archer, 1991). It is worth mentioning that tidal cyclicity patterns can be preserved even in growth lines of shells living in tide-influenced environments (Richardson, 1990; Murakoshi & Nakayama, 1992).

In the lower Miocene Pétervására Sandstone mud drapes on foresets of dm- to m-high dunes, and lateral changes in inclination and shape of foresets indicate the influence of tides (Chapter 3). Variations in foreset thickness were also recognized and interpreted as the result of diurnal tidal cyclicity (Tari et al., 1989). Recently acquired data by the author, documenting the migration of dunes over a year, however indicate that the system was semidiurnal. The aim of this chapter is to analyze the data of Tari et al. (1989), as well as the newly collected records of thickness variations and to discuss various tidal cyclicity patterns in the Pétervására Sandstone.

These data also allow to estimate the tidal range, the celerity of bedforms and the time involved in the development of dunes.

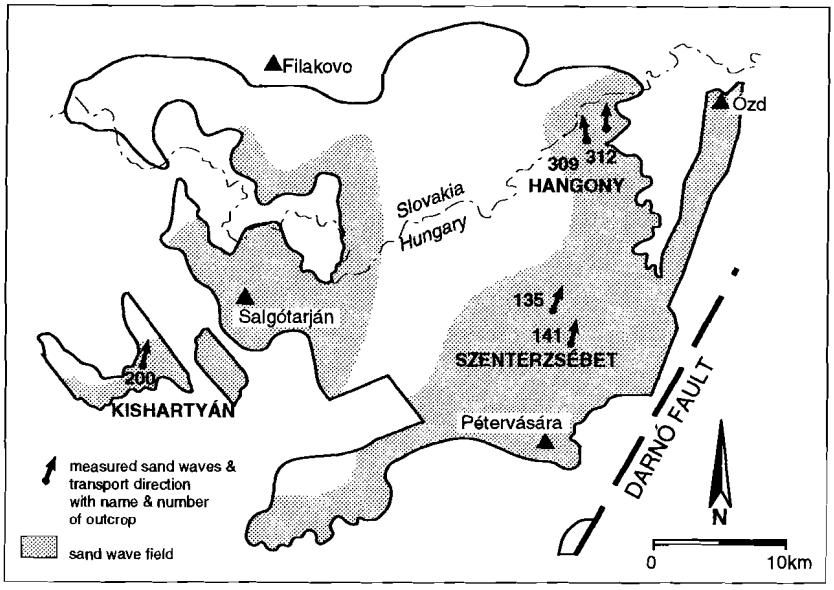

Fig. 5.1. Sand-wave fields within the Pétervására Sandstone. Locations where measurements were made are indicated.

Study area

A significant portion of the Pétervására Sandstone consists of sand waves formed by tidal currents (Fig. 3.3, Chapter 3). A large, almost continuous sheet of sand waves is exposed in the valleys NNE of Pétervására, up to the Slovakian boundary at Hangony (Fig. 5.1). Another more or less continuous sand-wave field crops out around Salgótarján (Fig. 5.1).

These sand waves were relatively large, as is indicated by the height of cross-bedded sets, up to 12 m. Cross-stratified sets of 1.5-3 m high are, however, the most common (Fig. 5.2). Foresets of sand waves are often shown by well developed variations in grain size, ranging generally from medium- to very coarse-grained sand. A single foreset usually consists of a lower coarse part grading into an upper medium-grained part (Fig. 5.2). The thicknesses of foresets were defined as the distance between the bases of subsequent coarse-grained laminae (Fig. 5.2), and were measured at several localities (Fig. 5.1). Foresets are relatively thin, averaging 37 mm with a range between 10-190 mm. In many cases the real thickness of foresets is even less than the measured values, because the strike of the outcrops is usually not parallel with the dip direction of foresets (e.g. Fig. 5.2). This offers no problems, because the *relative* differences within series of adjacent foresets are analysed below.

The thickness measurements usually resulted in short data sets (40-70 successive foresets). At two localities (Hangony 312 and 309, Fig. 5.1), however, two long and continuous series, up to 160 and 300 succesive foreset laminae in large, presumably straight-crested sand waves were measured.

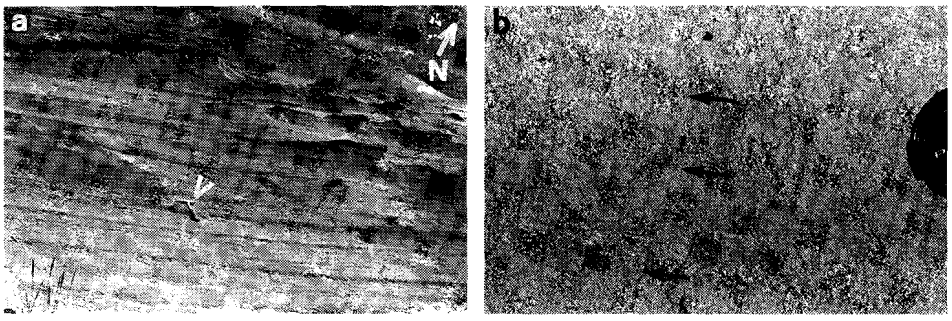

Fig. 5.2. A. Large, 3 m high sand wave with well developed foreset laminae at Hangony (locality no. 309). Note the long, straight foresets dipping towards the north, which is 30° inwards the hill. Cyclicity - corresponding to neap-spring-neap tides - is reflected by the darker laminae. B. Close-up of foresets consisting of coarse- and medium-grained parts, with a weak gradation in between. Arrows mark the bases of single foresets, which represent one tidal (always ebb) event.

Sedimentation and tidal cyclicity

Foresets deposited during one tidal cycle by the dominant tidal current are called bundles (Boersma, 1969). Bundle thickness can be regarded to be proportional to the rate of sediment transport, which depends on current velocity. This particularly holds if the subordinate current does not erode the foreset deposited by the dominant current. Small variations in current velocity are exagerated by the foreset record, because the rate of sediment transport is a third to fifth power function of the current velocity (Allen, 1984 p.96).

In a tide-influenced environment the velocity of flood and ebb currents depends largely on the tidal range, i.e., the difference of subsequent low and high water levels. These in turn are goverened by the gravitational attraction by the Moon and the Sun. These forces in combination with centripetal forces generate two bulges, which travel, as two tidal waves, over the Earth from east to west, due to the rotation of the Earth (Fig. 5.3). The magnitude of the gravitational forces varies to a different extent on various time-scales due to the revolution of the Moon and the Earth, and periodical changes in inclination of planes of rotation. Here the most prominent tidal periods spaning up to a year will be reviewed (for a detailed discussion see Pugh, 1987). These periods, considering the number of adjacent foresets which was measured, may be recognizable in the Pétervására Sandstone.

Half-daily cycle

The most obvious changes in sea level are shown every day, due to the migration of the two tidal bulges over the Earth. Normally one high and one low water (one tidal cycle) occur during half a day (12.41 h; therefore these are called semidiurnal tides), thus producing two tidal cycles daily (Fig. 5.3). In some areas only one tidal cycle occurs daily (diurnal tides). These, however, are relatively rare in modern seas (Lisitzin, 1974). The development of the end-members (pure semidiurnal and diurnal tides), as well as that of mixed systems, depends on a number of factors, the discussion of which is beyond the scope of this chapter.

Daily cycle (diurnal inequality)

A characteristic daily periodicity also occurs. The maximum of the tidal bulges, facing towards and away from the Moon, commonly does not coincide with the equator, due to the declination of the Moon (Fig. 5.3). This results in a higher and a subsequent lower high-tide event. This is called diurnal inequality. Its effect is most pronounced at higher latitudes and is not significant at the Equator. The diurnal inequality is a good indicator of semidiurnal systems (cf. de Boer et al., 1989).

Lunar semimonthly cycles

Not only the Moon, but also the Sun plays an important role in the development of tides. The Sun-Earth distance, however, is much greater than the Moon-Earth distance. Thus the magnitude of solar tides is approximately half of the magnitude of lunar tides, despite the Sun's much larger mass. At full and new moon, when the Moon, the Earth and the Sun are located on a straight line, the solar and the lunar tides are added, resulting in spring tides (highest tidal range). At the quadratures of the Moon, the Sun and the Moon are at right angles with respect to the Earth, resulting in neap tides (lowest tidal range; Fig. 5.3). The time between two spring tides is 14.8 days (Pugh, 1987). During this two-week period 28 tidal cycles occur in a semidiurnal tidal system and only 14 in a diurnal system.

The inclination of the Moon, which causes the diurnal inequality of the tides, changes over a slightly shorter period. The Moon passes over the Equator in every 13.6 days. At that time no diurnal inequality occurs (Fig. 5.3). This change is called the tropical tide (cf. Archer et al., 1991). On curves recording changes of the water level, these are the cross-over points (C), where the regular alternation of higher and lower high-tides shows a step-over to the alternation of lower and higher high-tides (Fig. 5.3). Because the duration of spring-neap cycles exceeds that of the tropical tides, cross-over occurs roughly two days earlier in each subsequent spring-neap cycle.

Lunar monthly cycle (monthly inequality)

Subsequent spring tides show a difference in height as well (Fig. 5.3 and 5.4). The tide-generating force of the Moon is greatest when it is closest to Earth in its elliptical orbit (perigee)

and smallest when it is furthest away (apogee; ±15% of average spring tide; Pugh, 1987). If a spring tide (either at full or new moon) occurs around perigee, than the next spring (new or full moon approximately two weeks later) will happen close to apogee (Fig. 5.4). This will yield in an alternation of higher and lower spring tides, which is the "monthly inequality".

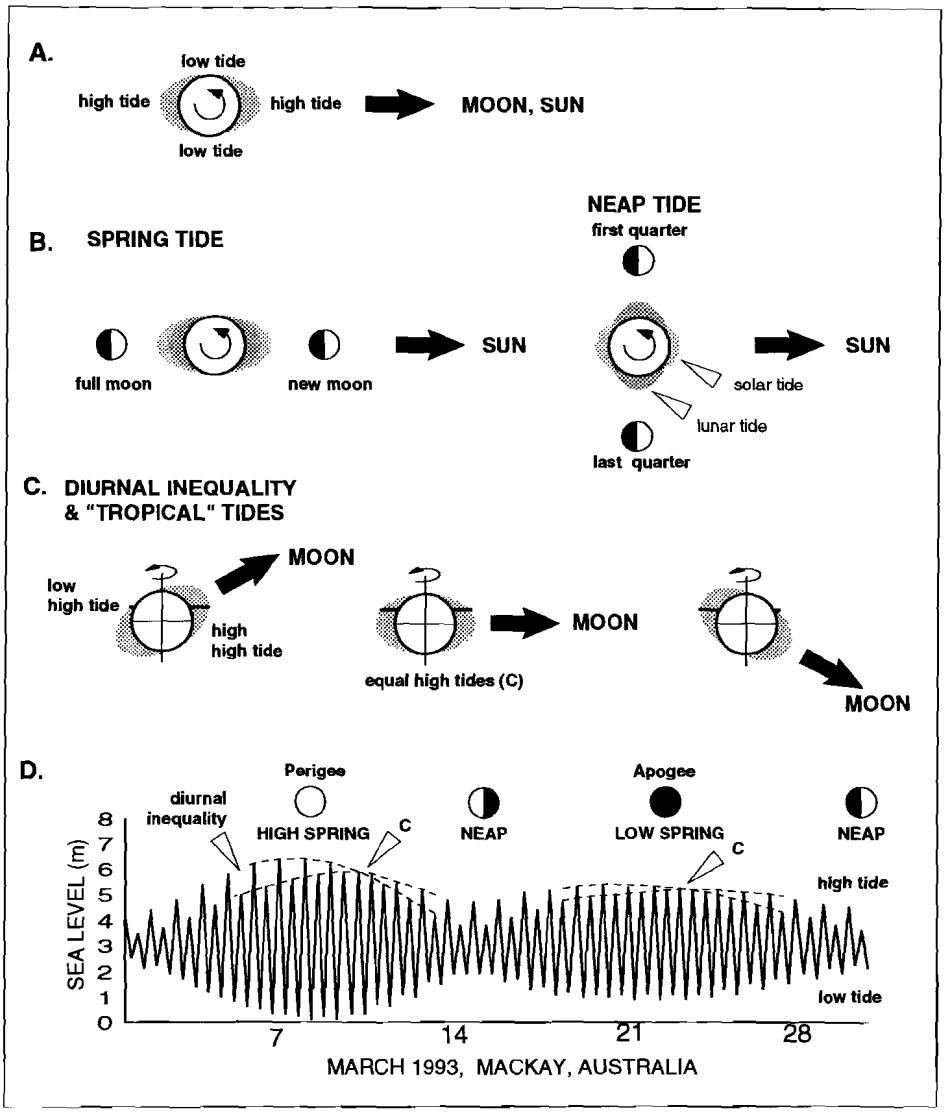

Fig. 5.3. Causes of semidiurnal (A), diurnal (C), spring-neap (B) and "tropical" (C) tidal cycles. See detailed explanation in text. D. Curve of daily sea-level changes at Mackay, NE Australia (hydrographic chart of 1993). The phases and the distance of the Moon, as well as its passage over the Equator (cross-overs) are indicated.

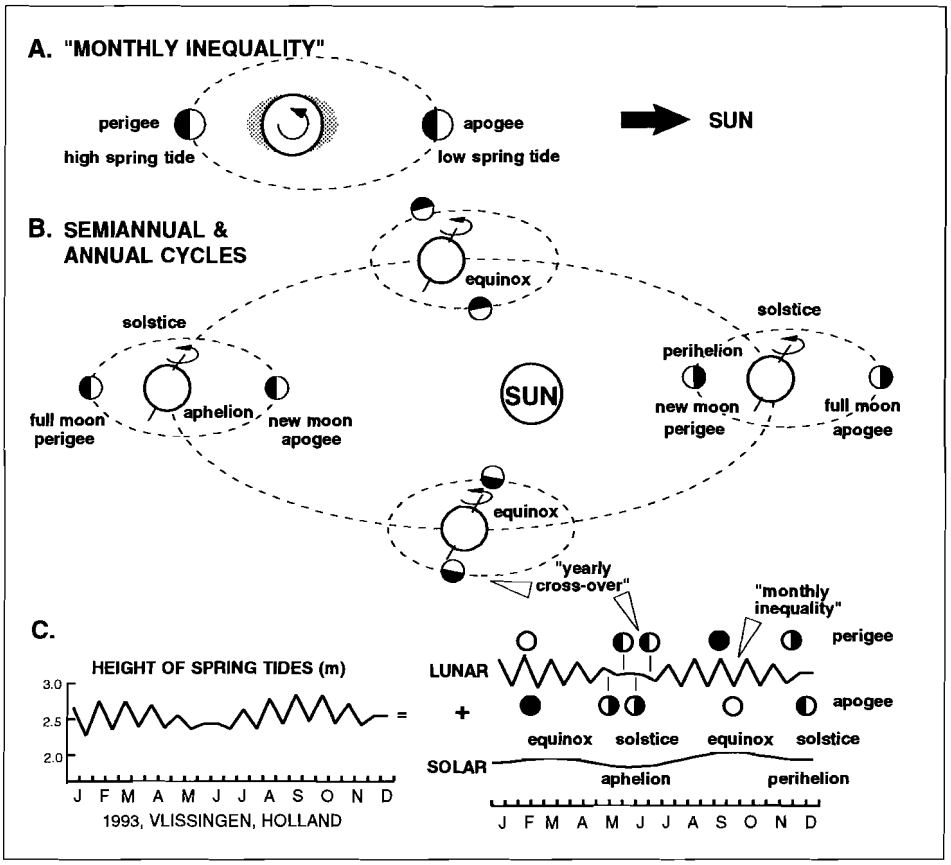

Fig. 5.4. Causes of fortnightly (A) and roughly half-yearly (B) changes of spring tides. See detailed explanation in text. C. Curve shows the height of subsequent spring tides at Vlissingen, The Netherlands (hydrographic chart of 1993). The "monthly inequality" and the "yearly cross-over" related to the position of the Moon are shown. The same data show the solar semiannual (solstice-equinox) and annual (perihelion-aphelion) effect as well.

Lunar "semiannual" cycle

There is a longer-term variation in height of spring tides as well (Fig. 5.4). This occurs because the time-span from perigee to the next perigee (27.6 day) differs from the time-span from new moon to the next new moon (29.5 day; Pugh, 1987). When spring tide coincides with the perigee of the Moon, the next spring - as described above - will be at about apogee. This results in the highest and the lowest spring tides, and thus in the largest monthly inequality. As the rotation continues there will be an occasion when both full and new moon are equally close to the Earth. At that time there is no difference in height between subsequent spring tides (cf. Williams, 1991) and the perigee and apogee of the Moon coincides with the neap tides. This interval can be called as the "yearly cross-over" of springs (Fig. 5.4). This also implies that during about six months highest springs occur at full moon, and in the next six months they occur at new moon (Fig. 5.4).

Solar semiannual cycle

This is due to tilt of the Earth axis, resulting in yearly variation in declination of the Sun. During solstice (winter and summer) when the declination is at a maximum, solar tides are small. During vernal and autumnal equinoxes, when declination is zero, the solar semidiurnal forces are at maximum at middle latitudes, resulting in high equinoctical spring tides (Fig. 5.4; cf. Pugh, 1987).

Solar annual cycle

The solar tide also varies yearly, with the distance between the Sun and the Earth. When the Earth is nearest (perihelion), the solar component is greatest, when the Earth is furthest (aphelion) it is smallest.

Both of these solar effects are small, and usually they are strongly blurred by seasonal climatic effects. Nevertheless they may be apparent. Annually 25-27 spring tides occur. This can help to distingush between semiannual and annual cycles.

The above described periods are well known and statistically significant in short-term sea-level records (cf. Pugh, 1987). The geological record - the successive thicknesses of cyclically accumulated strata - depends on many other factors. Suprisingly complete vertically accreted sand-mud couplets have been measured and the application of statistical methods (e.g. Fourier analysis) yields remarkable results. The sand-mud couplets reflect the various order tidal periods almost perfectly (cf. Tessier & Gigot, 1989; Kvale et al., 1989; Williams, 1991; Kvale & Archer, 1991; Oost et al., 1993).

However, in case of laterally accreting dunes, which develop in medium-grained or coarser sand, currents are often intermittently below the threshold of sediment transport (cf. Allen, 1981 de Boer et al., 1989). No foresets are deposited during such phases and therefore the record can be rather incomplete. Quite often migration of dunes even stops for longer times (Bartholdy pers. comm., 1992; Harris et al., 1992 and many recent studies). The opposite effect can occur e.g. during storms, which produce extremely thick laminae or erode part of the record. In addition, due to the limited size of exposures, the length of the record is usually small with respect to the length of cycles to be tested. Therefore the causual relationship of thickness variation of foresets of dunes with the periodic changes in the strength of tidal currents can rarely be confirmed by time-series analysis.

Methods

In order to see whether the various tidal cycles have been preserved in the foresets of the Pétervására sand waves, data were processed visually and by simple mathematical methods. In some of the data sets the presence of some long-term trend is obvious (Figs. 5.6d and 5.7d). These were analysed by calculating 6-7 order polinomial best fit curves.

The noisy pattern, both in raw data (Fig. 5.5a) and in data without the long-term trend (Figs. 5.6a & 5.7a) show a clear alternation of relatively thick and thin laminae. A laminae is thick, if it is thicker, and thin, if it is thinner than the average of its neighbours (cf. de Boer et al., 1989). Thick-thin alternations over a length of 5 lamina are shown on Figs. 5.5b-5.7b.

Data were also smoothed, i.e., the short-term trend was removed by a moving average procedure. The width of window was 3 and the averaging was applied three times. The result shows a gradual thickening and thinning of the successive laminae. The corresponding minimum points were projected back upon the "noisy" dataset (see vertical lines on Figs. 5.5-5.7), and compared both with local minima and thick-thin series. In this way neaps, where sedimentation had ceased, and bundle sequences from neap-to neap were determined.

Results

Diurnal inequality

Alternations of thick-thin laminae are interpreted as the result of the diurnal inequality of the tide. It is well recognizable in the raw data sets (Fig. 5.5b-5.7b). Little "irregularities" in the longer series - in the form of two successive thick or thin lamina are either crossover points (C on Fig 5.5-7) or reflect temporary influences (e.g. storms or relatively low current velocities). The shift in "cross-overs" - from after spring, to around spring and to before spring - is well developed at Szenterzsébet (Fig. 5.5. 141.b). A similar trend is present in the first two bundle sequences at Kishartyán (Fig.5.5 200.b). This shift is due to the difference in period between the lunar fortnightly (neap-to-neap, 14.8 days) and the tropical tide (13,6 days). Some cross-overs may have occurred around neap tide when no record has been preserved. Some bundle sequences do not contain thick-thin series or the series are too short to be significant. Such absence of thick-thin alternations occurs only incidentally and there is no reason to ascribe such absence to the occurrence of a mixed tidal regime, in which diurnal and semidiurnal periods alternate. Instead series of at least 8-10 alternating thick-thin lamina are common. Diurnal inequality is often present over the full length of bundle sequences, thus it is significant, and indicates a semidiurnal tidal system (cf. de Boer et al., 1989).

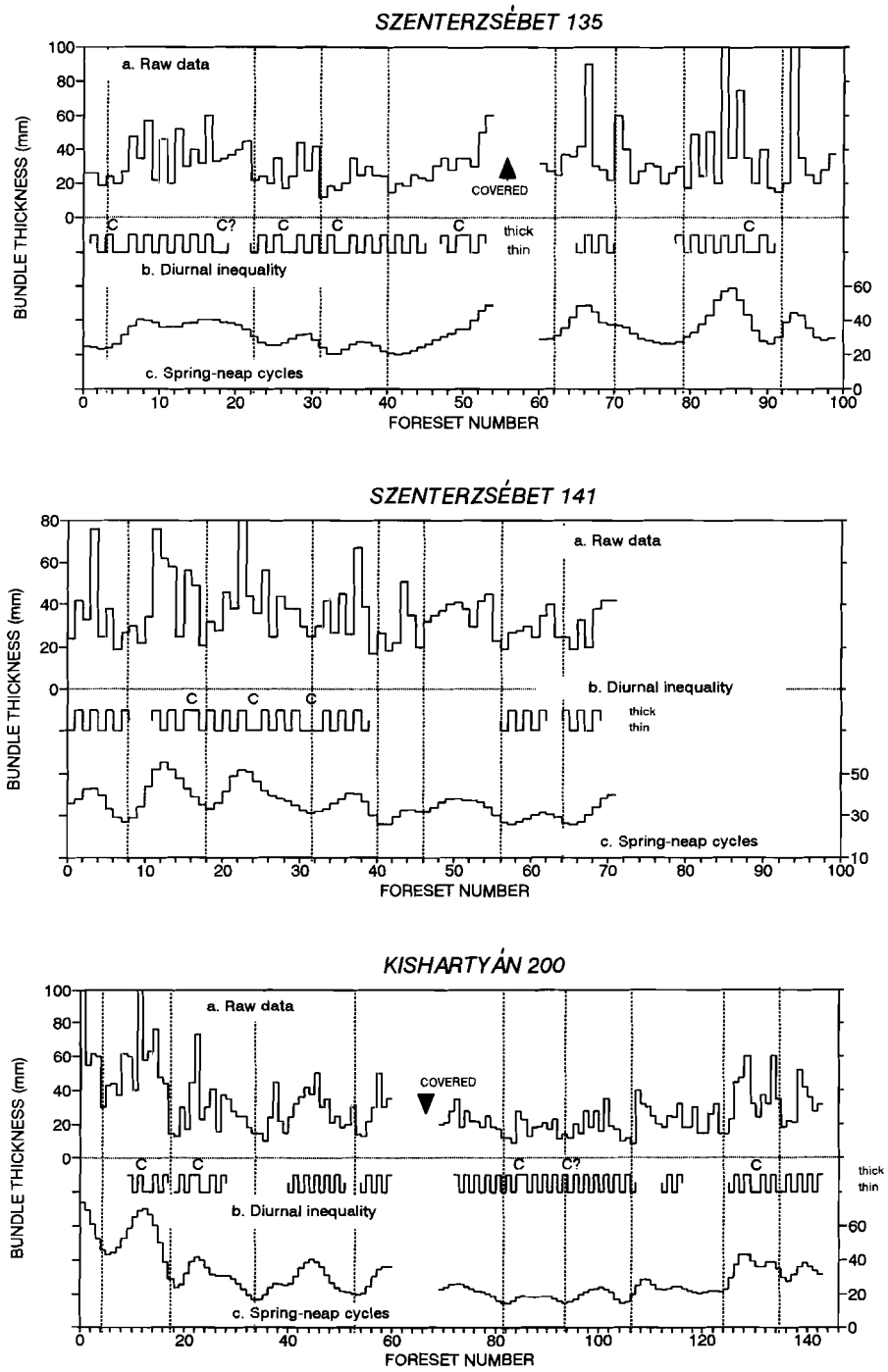

Fig. 5.5. Tidal cyclicity at two locations in Szenterzsébet valley and at Kishartyán. See explanation in text.

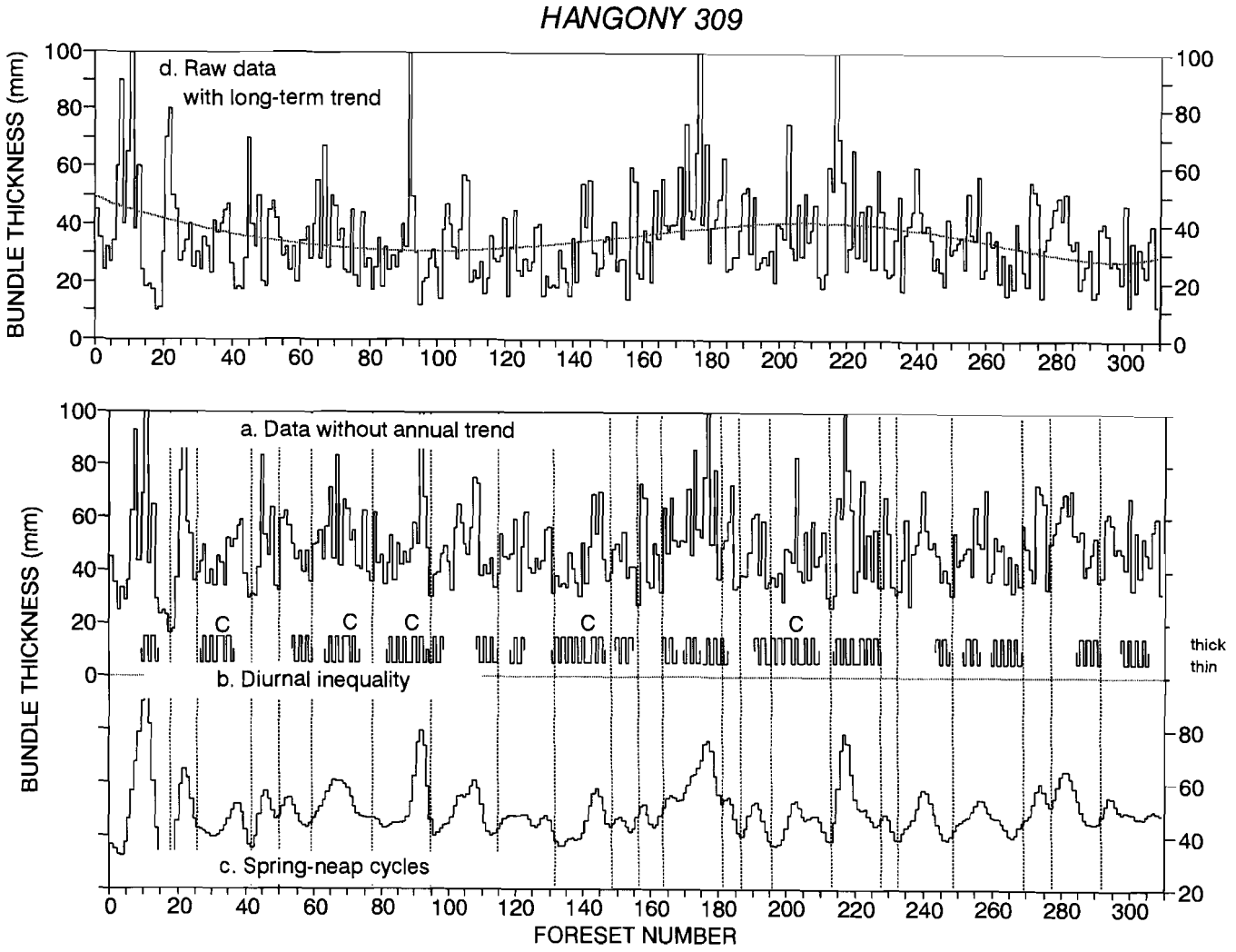

Fig. 5.6. Tidal cyclicity at Hangony 309. See explanation in text.

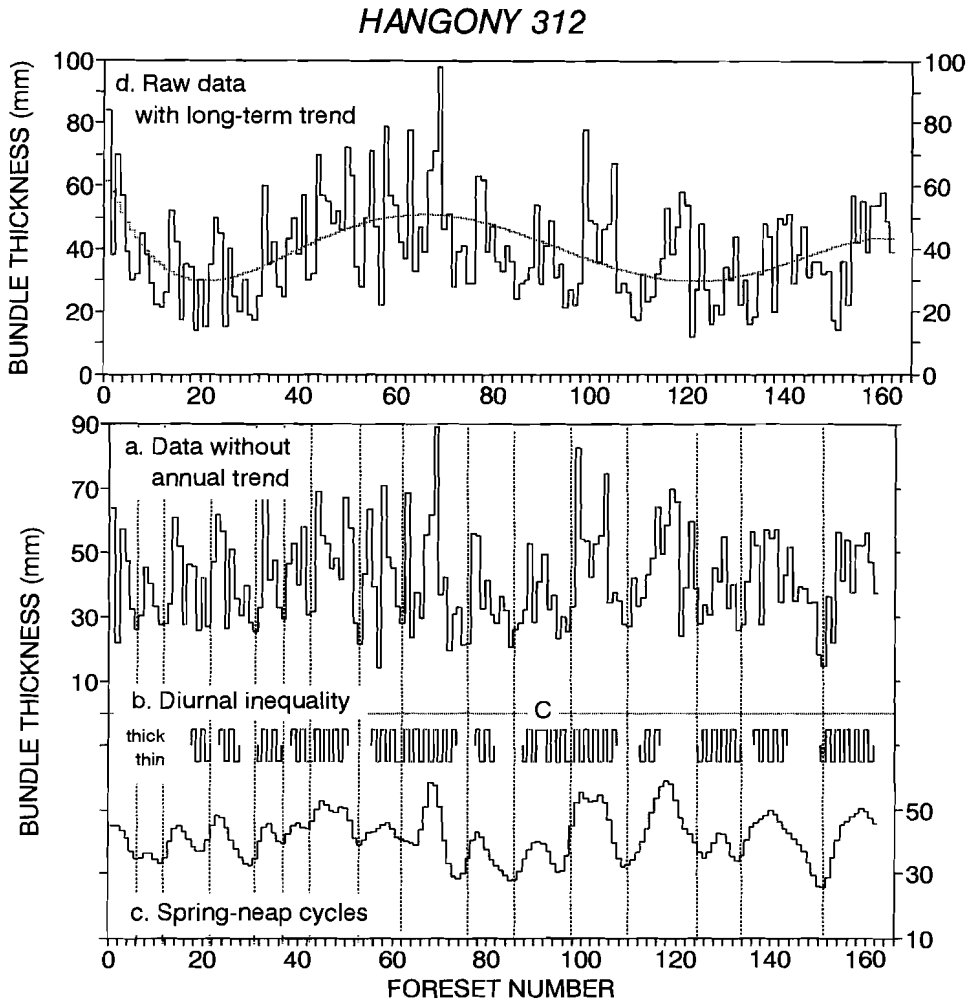

Fig. 5.7. Tidal cyclicity at Hangony 312. The same data set was interpreted by Tari et al. (1989). See explanation in text.

Spring-neap cycles

Spring-neap cycles are clearly reflected by the gradually thickening and thinning of the foreset thickness (Figs. 5.5c-5.7c). The approximate position of neaps is indicated by the vertical lines. The length of the cycles is fairly variable. The number of bundles in sequences ranges between 6 and 20, the most common value is 8-10, and the average is 11. These relatively short bundle sequences led Tari et al. (1989) to conclude that a diurnal tidal system governed deposition in the North Hungarian Bay during the early Miocene. The statistically significant presence of the diurnal inequality, however, provides a clear evidence for a semidiurnal system. The bundle

sequences thus are incomplete. This must be the result of insufficiently strong current velocities around neap tides. The relatively long quiescence is accentuated by the often abrupt appearance of thick lamina after neap. In some cases spring tide was followed by a gradual decrease of sedimentation (e.g. Fig. 5.5 first three cycles on 141.c). In other cases the lateral growth of the dune stopped as abrubtly as it began (e.g. Fig. 5.6).

The alternation in height of springs can be followed for 3-4 neap-spring cycles (Fig. 5.5a at locality 135 after foreset number 60, at locality 141 up to foreset 40, at locality 200 between 70 and 105). It is even more convincing at Hangony, where 5 subsequent spring tides alternate in height (Fig. 5.6.a between 20-80). As will be demonstrated below, many complete spring-neap cycles are often missing from the record. This explains the incidental lack of systematic changes in the height of springs.

Time recorded by spring-neap bundle sequences is 2 months at Szenterzsébet 135, 2 and 3 months at Kishartyán and 4 months at Szenterzsébet 141. These cycles thus are too short to determine which part of the year is represented. At Hangony, many bundle sequences were measured, covering the record of deposition during 8 and 11 months.

Semiannual cycles

The succession at Hangony 312 (Fig. 5.7) shows 160 subsequent foreset laminae. Two minima, and a maximum in between are clearly present. The complete series covers 16 neap-spring-neap cycles. Between the two minima about 10 cycles are counted. During half a year about 13 spring-neap cycles can be formed in a semidiurnal tidal system. This obviously can be less, if sediment transport ceases for some longer time. The long-term signal in exposure Hangony 312 is interpreted as the product of the semiannual solar tide. Compared to the present day conditions, the minima correspond to the summer and winter solstices, and the maxima to the passage through the equinoxes. Thus the record represents a complete year.

Annual cycles

A long-term signal was recorded at Hangony 309 (Fig. 5.6), too. It shows two weakly developed minima and one maximum in the data set. From Fig. 5.6 one may count about 17 bundle sequences from minimum to minimum. It cannot represent the semiannual equinoctical tide, because it is more than can be formed during half a year (13). On the other hand the record of 17 bundle sequences is less than a full annual record of 26 spring-neap events. Considering that in the studied examples obviously a number of complete neap-spring cycles did not leave a record in the sediment, the trend in Fig. 5.6 is interpreted as the product of the solar annual cycle. The maximum part corresponds to perihelion, the minimum part to aphelion. This record therefore covers more than a year, and is incomplete because during many neap-spring cycles current velocities were insufficent for the transport of sediment and migration of the bedforms.

Estimates of tidal range and current velocities

The recognition of tide-influenced deposits provides very important information about the sedimentary environment. They also may offer insight into the paleo-tidal range and peak tidal current velocities which occurred during deposition. Commonly many factors needed for the establishment of these parameters are unknown in fossil settings. Nevertheless calculations were attempted in several cases and some estimates of hydrodynamic parameters were given (cf. Siegenthaler, 1982; Nio et al., 1983; Allen & Homewood, 1984). Below the method of these authors is applied.

The dimensionless bed load transport and sediment transport rate were determined from the thickness of the bundles and the height of the bedforms (3 m). For the calculations only the thicknesses of bundles formed at spring and close to neap tide were used in order to obtain an upper and a lower limit, respectively. Because the grain size of the sediment is not uniform, calculations were made for coarse (1 mm) and medium-grained (0.5 mm) sand separately, which gives a range of values. These were used to determine dimensionless shear velocity and shear velocity. Tidal range was estimated from the shear velocity and the average depth of the flow (Nio et al., 1983). The latter was most likely of the order of 10-30 m. Therefore calculations were made for depths of 10, 15 and 20 m. The shear velocity was, moreover, used to calculate surface current velocity (cf. Allen & Homewood, 1984). The estimated tidal ranges are shown in Table I, the estimated peak surface velocities are shown in Table II.

Table I.

	TIDAL RANGE (m)			
	maximum		minimum	
depth \ grainsize	1mm	0.5mm	1mm	0.5mm
10 m	5.6	4.1	3.9	2.8
15 m	4.6	2.3	3.2	2.3
20 m	4	2.9	2.8	2.1

Table II.

	PEAK SURFACE VELOCITY (cm/s)			
	spring		neap	
depth \ grainsize	1mm	0.5mm	1mm	0.5mm
10 m	83.1	70.7	69.6	59.2
15 m	89.5	76.1	74.9	63.7
20 m	94.0	80.0	78.7	66.9

The smallest value for the palaeo-tidal range, calculated from the very small thickness of bundles around neap tides is slightly above 2 m. The largest values exceed 4 m. Thus, based on the calculation method of Siegenthaler (1982) and Nio et al. (1983), at least a mesotidal, but rather a macrotidal environment seems to have defined the deposition of sand waves in the Pétervására Sandstone.

Peak surface current velocities thus must have been in the range of 70-90 cm/s during spring tides and less than 60-80 cm/s around neap tides. These values compare well with those measured in shallow, modern tide-influenced seas, where active migration of dunes (sand waves) is observed (cf. Belderson et al., 1982; Dalrymple et al., 1990 Harris, 1991).

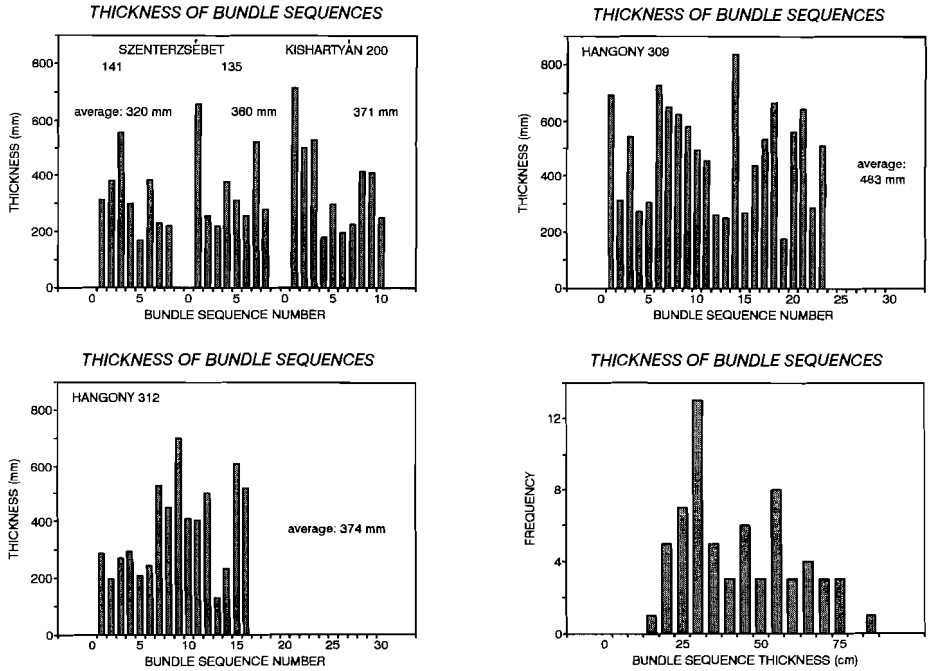

Fig. 5.8. Bundle sequence thickness at the discussed localities and their frequency distribution. The bundle thicknesses should be corrected for the difference between strike of outcrops and dip direction of foresets (maximum 30°). Bundle sequences of about 30-35 cm thick are the most common.

Celerity of dunes

The bundle thickness data demonstrate that months and years can be recognized in the laterally accreted dunes formed millions of years ago, which is fascinating even for a geologist's eyes.

Table III. Recent and ancient examples of dunes and their celerity as a function of bed height

		age	formation (locality)	height of dunes (m)	bundle thickness (cm)	celerity of dunes (m/14 days)	number of bundles in sequence
Richards	1986	Tr	SW Alps	0.64	1-7	14	~11
Kreisa & Moiola	1986	J	Curtis	0.8	18-80	30-40	15-20
Uhlier et al.	1988	J	Sundance	1	1-45	~4-6	9-13
v.d. Bergh	1982	R	Osterschelde	1.5	1-50	6.5	
J.R.L. Allen	1981	Cr	Greensand	1.8	3-150	~10	10-14
Siegenthaler	1982	R	Osterschelde	2	4-30	5-10	26-30
Lapido	1988	Cr	Ajali	2	2-40	~4	7-13
Harris et al.	1992	R	Moreton bay	2	~10	1.7	7
Allen & Homewood	1981	Mi	Swiss molasse	1-2	3-30	4.4-15	25
Berne et al.	1988	R	English ch.	3.5-7.5	?	5-10?	
Bartholdi (pers. com.)	1992	R	Faro inlet	4-5	?	0.15	
Kreisa et al.	1986	R	Rancho Rojo	6-12	1-22	~1-2	10
Sztanó	1994	Mi	Pétervására	3	1-11	0.6-0.8	13
			Pétervására	0.5	5-40	~7	?

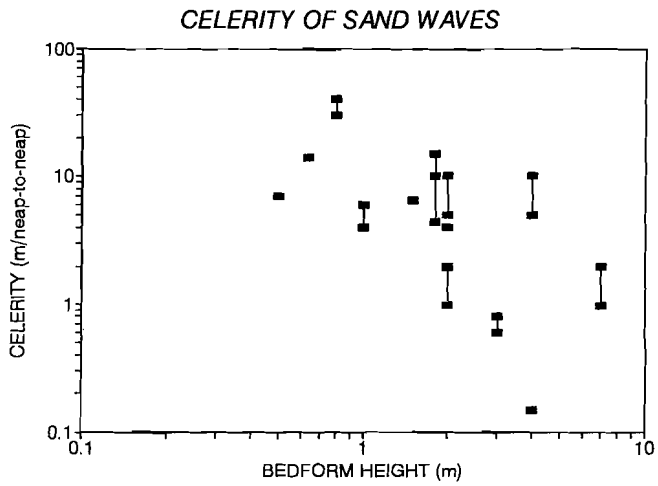

Fig. 5.9. Celerity decreases with an increasing height of sand waves. Data from Table III.

The thickness of bundle sequences, representing neap-spring-neap tidal cycles, was between 32 and 48 cm on the average in the Pétervására Sandstone (Fig. 5.8). The migration rate (celerity) of sand waves, calculated from thickness of bundle sequences and dip angle of foresets (30°), was about 64-82 cm from neap-to-neap (Table III) parallel to the flow, corrected for the true migration direction of bundles. These sand waves of 3 m height thus moved very slowly. The sum of bundle sequence thicknesses is 5.5 m and 9.5 m at Hangony 312 and 309 respectively, which were covered during about a year. Thus, considering the about 30° dip angle of foresets, the migration rate has been twice as much as the thickness of foresets and were of the order of 11 to 19

m/year. Such values can be compared with the celerity of about 20 m/half-a-year of 2 m high dunes in a modern embayment in Australia. These dunes only move during every second strong flood around spring, thus producing about 7 bundles from neap to neap (Harris et al., 1992 Table III and Fig. 5.9). Supposing a migration rate of the same order of magnitude for the sand waves at Istenmezeje (Fig. 3.6, Chapter 3), which are exposed over a length of 500 m, these would represent a development over a period of 25-50 years.

The 13 listed examples (Table III.) show that the thickness of bundles decreases with increasing height of the sand wave. Thus migration rates of dunes are inversely related to their height, altough grain size, depth of deposition and tidal current velocities are obviously different. Such relation is obvious whereas more sediment is needed for the migration of larger dunes. The analyses of tide-influenced characteristics of sand waves in the Pétervására Sandstone indicate a rate of migration of the order of 0.6-0.8 m/14 days. Considering the height of the sand waves this value fits well to other data described in the literature.

Conclusions

Measurement of foreset thickness variation in the sand waves of the Pétervására Sandstone allows a detailed reconstruction of the tide-influenced environment. It was shown that:

- Foreset laminae are tidal bundles deposited by the dominant tidal current. The presence of mud drapes and/or reactivation surfaces at the bundle boundaries is not neccessary to identify tidal influences.

- Deposition occurred in a semidiurnal tidal system, as is indicated by the thick-thin thickness variations of foreset laminae.

- Spring-neap tidal cycles are highly incomplete, shown by the average of 11 bundles in neap-to-neap sequences (instead of about 28). Sediment transport was confined only to the strongest currents around spring.

- Some of the sand waves are sufficiently long laterally to recognize semiannual and annual variations of tidal currents. On that base the annual down-current migration of sand waves can be estimated to have been about 10-20 m.

- The annual record moreover reveals that tidal currents remained below the threshold of sediment transport not only around neap tides, but occasionally also during low spring tides.

- Calculations based on the thickness of foresets indicate a meso- or rather a macrotidal range (above 4 m), with spring-tide current velocities of about 70-90 cm/s.

CHAPTER 6
PALAEOGEOGRAPHIC SIGNIFICANCE OF TIDAL DEPOSITS: AN EXAMPLE FROM AN EARLY MIOCENE PARATETHYS EMBAYMENT, NORTHERN HUNGARY

Abstract

Tide-influenced deposits can be used in palaeogeographic reconstructions because they are good indicators of open marine connections. An example is presented from the late Aquitanian - early Burdigalian (~Eggenburgian) of the North Hungarian Bay. This bay was part of an inland sea, the Paratethys. Tide-influenced deposits prove that tidal motions in the North Hungarian Bay were locally amplified. This required a free propagation of tidal waves from the open ocean through the Paratethys into this embayment. Since all seaways towards the Mediterranean were closed during the late Aquitanian - early Burdigalian, the only connection between the North Hungarian Bay and open marine waters, which could allow the transmission of tidal waves was the outlet towards the East Slovakian Basin in the northeast. The presence of tide-influenced deposits in the North Hungarian Bay implies that tidal waves entered the Eastern Paratethys from the east through a wide passage.

Other examples of Lower Miocene tide-influenced deposits in the Mediterranean and Paratethian regions were reported from slightly different periods. In these cases amplification of tidal motions in various embayments and straits also could occur because of the local basin morphology. It is demonstrated, that significant palaeogeographic changes during the Early Miocene resulted in changes of current pathways and related shifts of loci of tide-influenced deposition.

Introduction

General

Tides, which result from the gravitational attraction of the Moon and the Sun, are generated everywhere on Earth. Their effect is most pronounced in large bodies of water. Tidal waves are induced in oceans and are amplified when propagating onto shallow shelves. Therefore, wide shelf areas are often tide-dominated (Elliot, 1986). The interaction of the tidal waves with the bottom topography, its depth and morphology in particular, and with the geometry of the coastline determines its amplification or damping (Lisitzin, 1974).

The propagation of the tidal wave is obstructed when it passes through one or more narrow passages (e.g. Magellan Strait, Medeiros & Kjerfve, 1988). In inland seas, such as the present-day Mediterranean and Baltic seas, tidal amplitude is usually very small. The reason is that the tidal

waves of the open ocean cannot enter and the volume of the basin itself is too small to effectively generate tides (cf. Johnson & Baldwin, 1986). Straight, wide and shallow straits can locally intensify tidal motions (e.g. Torres Strait, Harris, 1988). The physiography of semi-enclosed bodies of water, such as some epicontinental seas, embayments and estuaries, can be favourable for the amplification of tidal motions. If they are wide enough, the tidal wave moves around with zero amplitude in the center and maximum amplitude along the coasts (amphidromic amplification, e.g. North Sea; Lisitzin, 1974). If the relatively shallow semi-enclosed water body is funnel-shaped and if its length is close to a given portion of the tidal wave length, resonant amplification develops (e.g. Bay of Fundy; Lisitzin, 1974; Dalrymple et al., 1978).

Local amplification of tidal motions can lead to a meso- or macrotidal setting (Elliot, 1986; Johnson & Baldwin, 1986;). Therefore, any meso- or macrotidal deposit formed in a shallow marine or littoral environment indicates local or nearby amplification of tidal motions. This thus implies a connection to open marine areas and a favourable physiography for the amplification of tidal motions in the ancient sea.

The Paratethys and the North Hungarian Bay

Numerous Lower Miocene deposits in the Paratethys reflect the presence of strong tidal motions (Fig. 6.1), although the Paratethys was a relatively closed inland sea (Báldi, 1980; Rögl & Steininger, 1983, 1984, Steininger et al., 1985; Nagymarosy, 1990a). Tide-influenced deposits were reported from the early Aquitanian of the Venetian Basin in northern Italy (Massari et al., 1986), from the late Aquitanian - early Burdigalian of the North Hungarian Bay (Tari et al., 1989; Sztanó, 1992b; Chapter 5), from the Burdigalian sequence in the Upper Marine Molasse in Switzerland (Homewood & Allen, 1981 Allen et al., 1985; Allen & Bass, 1993), and from the middle Burdigalian Molasse in Austria (Faupl & Roetzel 1987, 1990; Krenmayr 1991).

From palaeontological and stratigraphical data it has been evident since long that the North Hungarian Bay was in communication with other basins of the Paratethys (Steininger & Senes, 1971; Báldi & Senes, 1975). In order to explain the dispersion of the aquatic biota, narrow straits and/or complicated seaways were sufficient. Thus three pathways were suggested as the direction of possible connections (Fig. 6.1; Rögl & Steininger, 1983; Báldi, 1986; Nagymarosy 1990a). However no sedimentary record of these possible marine connections has been preserved between the North Hungarian Bay and other basins.

Apart from the fauna associations, also the presence of well-developed late Aquitanian - early Burdigalian (mainly Eggenburgian) tide-influenced sediments in the North Hungarian Bay (Sztanó, 1992a,b, Chapter 5) indicate that this bay must have had a wide passage towards open marine areas. Only in that case could tidal waves enter into the basin without a strong reduction of amplitude.

The objectives of this chapter are: 1. to analyze the palaeogeographic significance of the presence of tide- influenced deposits in the North Hungarian Bay; 2. to demonstrate how this bay

was connected to other Paratethys basins; and 3. to discuss other examples of tidal deposits in relation with palaeogeographic changes, such as the closure and reopening of Paratethian seaways.

Palaeogeographic setting

Paratethys

From the Oligocene onward the Paratethys and the Mediterranean Sea developed (Fig. 6.1). In a geographic sense they are the descendants of the former Tethys. The Paratethys was a large "inland" sea, consisting of a chain of basins extending from the Alpine foredeep to the Caspian area (Fig. 6.1). Although these basins had a different structural development, a series of events, demonstrating a free exchange of aquatic biota, can be recognized and correlated throughout the area (Báldi, 1980, 1989; Rögl & Steininger, 1984; Nagymarosy, 1990a).

The Paratethys is divided into two slightly different provinces: the Central and the Eastern Paratethys (Fig. 6.1). The Alpine-Carpathian region belongs to the Central, and the Ponto-Caspian area to the Eastern Paratethys.

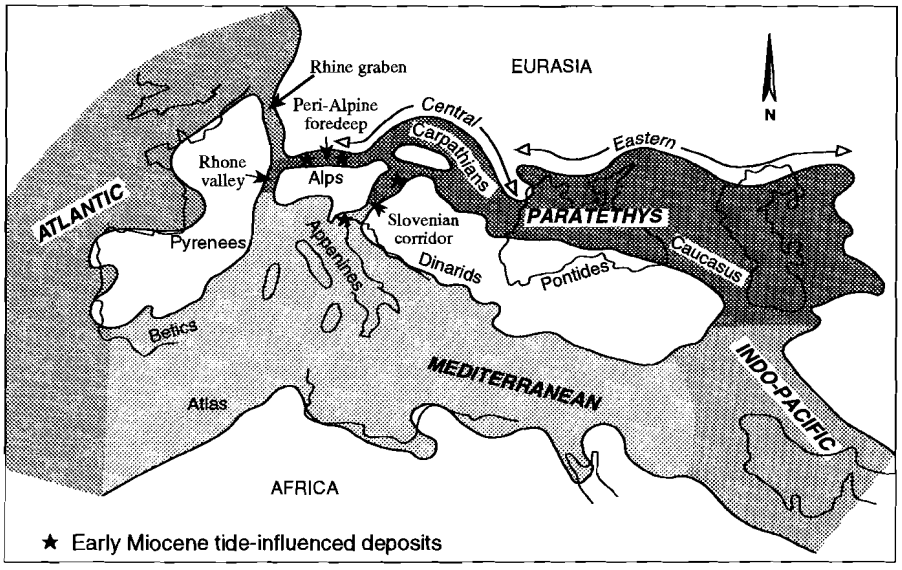

Fig. 6.1. General palaeogeography of the Paratethys during the Oligocene and Miocene (after Rögl & Steininger, 1983). Well known Early Miocene tide-influenced deposits are found in or near seaways.

The Paratethys had various connections to other marine basins, like the Atlantic, the Mediterranean and the Indo-Pacific (Fig. 6.1). These pathways, however, were blocked occasionally, giving way to the development of endemic biotas within the Paratethys (Báldi, 1980,

1989). Consequent characteristic changes of floral and faunal elements provided the base for a local bio- and chronostratigraphy (Fig. 6.2; Steininger & Senes, 1971; Báldi & Senes, 1975; Steininger et al., 1990).

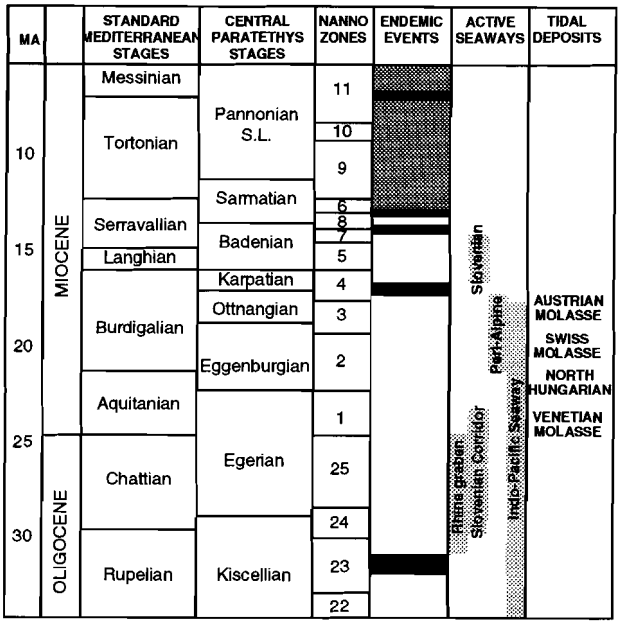

Fig. 6.2. Correlation of standard Mediterranean and Central Paratethys stages based on nannoplankton zonation (after Nagymarosy & Müller, 1988; Nagymarosy & Báldi-Beke, 1988). Periods of isolation, characterized by endemic faunal communities, are indicated by black bars. Periods, in which tidal influence was recorded somewhere in the region are also indicated.

The Central Paratethys and the Eastern Paratethys were connected through the foreland of the Eastern Carpathians. The eastern outlet of the Eastern Paratethys was the Indo-pacific "seaway" (Fig. 6.1; Rögl & Steininger, 1984). In the Central Paratethys three pathways towards the "west" have been recognized. The Rhine graben formed a connection to the Atlantic during mid Oligocene (Fig. 6.1). Communication with the Mediterranean Sea occurred through the Slovenian corridor, through the peri-Alpine foredeep and the Rhone valley (Fig. 6.1; Rögl & Steininger, 1984; Steininger et al., 1985; Nagymarosy, 1990a). These pathways were strongly influenced by eustatic sea-level changes and by the structural evolution of the Alpine-Carpathian-Dinaridic orogenic belt (Nagymarosy, 1990a).

The "first separation" of the Paratethys occurred during the early Oligocene (NP23 nannoplankton zone, Fig. 6.2), when the three seaways to the west and the north were all closed. During the mid-Oligocene (NP24) the connection to the oceans was restored through the Rhine graben and the Slovenian corridor (Báldi, 1980). The Rhine graben lost its seaway function again before the end of the Oligocene. Before the late Aquitanian the Slovenian corridor was also closed (Rögl & Steininger, 1984; Steininger et al., 1985; Massari et al., 1986; Nagymarosy, 1990a). This implies that during the Late Aquitanian the Central Paratethys must have been connected with open marine areas through the Eastern Paratethys. In the Burdigalian the Rhone valley and the

peri-Alpine foredeep were inundated over their full length for a second time, and thus the Upper Marine Molasse was formed (cf. Schoepfer & Berger, 1989). Before the late Early Miocene (Karpatian), the Peri-Alpine seaway was closed again by intense uplift of the Eastern Alps and the resulting high rates of clastic sedimentation. Meanwhile, the Eastern Paratethys had also been shut by the closure of the Indo-pacific seaway (Rögl & Steininger, 1984). Therefore, during the late Early Miocene a second isolation of the Paratethys occurred. From that time onwards only reduced marine (brackish) influences are known from the Eastern Paratethys (Kókay, 1985). The Slovenian corridor opened once more during the middle Miocene (Badenian) (Nagymarosy & Müller, 1988) as a result of extensional tectonics in the Intra-Carpathian area. From the late Middle Miocene (Sarmatian) onward the Central Paratethys became completely isolated (Rögl & Steininger, 1984; Steininger et al., 1985; Nagymarosy, 1990a). From that time onwards, endemic events were mainly the result of climate changes (Báldi, 1980).

The North Hungarian Bay

The Oligocene to early Miocene sedimentary successions in Slovenia, northern Hungary and south Slovakia are analogous, indicating deposition within the same basin (Nagymarosy, 1990b). The Slovenian part of the basin represents the Slovenian corridor. The Slovenian and the Hungarian parts of the basin were torn apart by large-scale transpressional wrench tectonics, most likely during the Late Oligocene to Early Miocene period (Csontos et al., 1992). The north Hungarian - south Slovakian part is traditionally called the North Hungarian "Palaeogene" Basin, where marine sedimentation occurred from the Late Eocene to the Early Miocene (Báldi & Báldi-Beke, 1985). During the early Miocene the North Hungarian "Palaeogene" Basin became, paleogeographically, an embayment. Because the name "Palaeogene" is anachronistic for the early Miocene situation, this area here will be referred to as the "North Hungarian Bay".

From the Late Oligocene onwards silty sediments (Szécsény/Lucenec Schlier) were deposited in the deepest parts of the embayment. Schlier is a deep neritic - shallow bathyal deposit formed in a water depth of 60 m to about 300 m (Báldi, 1986). During the Late Oligocene - Aquitanian (Egerian) shallow marine sands, the Törökbálint Formation and the "Tura" Sandstone, were deposited along the western and southeastern margins of the bay (Fig. 2.8; Báldi, 1986). In the early Aquitanian the sea flooded new areas in the northeast. Above the basal transgressive deposits the Szécsény Schlier was formed, indicating an extension and a deepening of the basin towards the northeast (Báldi, 1986; Vass et al., 1988). At the same time deposition of shallow marine sands continued along the southern side of the embayment (Nagymarosy unpublished, 1988; Lakatos et al, 1991). During the late Aquitanian - early Burdigalian (around the Egerian/Eggenburgian boundary; Fig. 6.2) a pronounced basinward shift of the shallow marine sandy facies occurred in large areas along both the western and eastern margins of the basin (Fig. 2.8). Thus deposition of the Pétervására Sandstone and of the Budafok Sand started (the Slovakian

counterparts are the Filakovo Sandstone and Lipovany Sands respectively; Vass et al., 1988). The appearance of wide-spread shallow marine sandy facies was the result of an eustatic sea-level fall prior to the Burdigalian (Sztanó & Tari, 1993, Chapter 8). As the sandy sediments prograded, the deposition of bathyal siltstones was restricted to a relatively narrow trough in the northern part of the North Hungarian Bay (Fig. 2.8; Báldi, 1986). At the same time, the shallow southeastern part of the bay most likely became subaerially exposed (Fig. 2.8; Sztanó & Tari, 1993).

Tidal influence in the North Hungarian Bay

Along the western coast of the North Hungarian Bay the Budafok Sand was deposited in the narrow littoral-sublittoral zone, at a water depth of up to 10-30 m (Fig. 6.3; Báldi & Nagy-Gellai, 1990). Molluscs found in the sand are indicative of the inter- to shallow subtidal zone (Báldi, 1986). At this side of the bay, the slope towards the relatively deep center of the bay was much steeper than the slope along the eastern side (Fig. 6.3). Sedimentary structures along the eastern margin, such as large-scale composite crossbedding, indicate a dominant sand transport to the south. Horizontal bedding, with current ripples and mud drapes is also characteristic for the Budafok Sand. Accumulation of a thick Egerian to Eggenburgian sedimentary succession is known from the southernmost tip of the embayment (Fig. 6.3). From the same place (Budafok, Pacsirta-Hill) vertically accreted sand/silt laminites were reported, which were interpreted as intertidal mudflat deposits (Báldi, 1959). These structures, in concert with the palaeoecological data, suggest a depositional environment of sandy inter- and subtidal flats cut by channels.

Tidal influence was clearly demonstrated in the lower Miocene Pétervására Sandstone, which was deposited in the northeastern segment of the North Hungarian Bay (Tari et al., 1989; Sztanó, 1992.b, Chapter 5). Conglomeratic lobes and lee sides of 2-6 m high sand waves, facing to the north, indicate highly asymmetrical tidal motions, with a dominant current towards the north (Fig. 6.3). Cyclic thinning and thickening of foreset laminae of the sand waves is interpreted as tidal bundle sequences, the result of a semidiurnal tidal regime (Fig. 5.5-7; Chapter 5). Peak tidal-current velocities are estimated to have been around 0.7-0.9 m/s (Chapter 5). The depositional depth was between 10-30 m, and all bedforms clearly indicate subtidal deposition (cf. Allen, 1984, see Chapter 5).

Smaller dunes and finer grained sediments were formed in deeper water, gradually interfingering with the bathyal siltstones (Sztanó & Tari, 1993). At the northeastern side of the bay a gradual deepening and a decrease of the tidal energy occurred towards the central depression of the bay (Fig. 6.3).

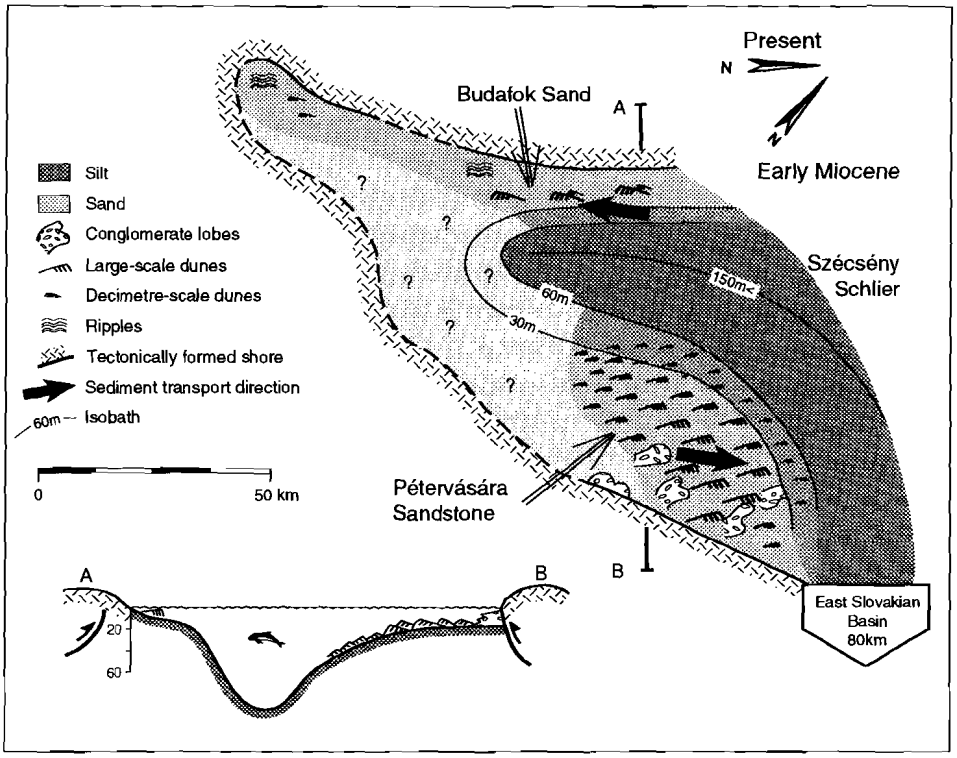

Fig. 6.3. The reconstructed dimensions of the North Hungarian Bay with its characteristic sediments and bathymetry. The estimate of the bathymetry is based on palaeocological data (Báldi, 1986) and on size of the bedforms (cf. Allen, 1984; Chapter 3, 7). The bay was connected to the East Slovakian basin, through a seaway where no sediment has been preserved (to be discussed later in this chapter). The geographical North during the early Miocene is based on palaeomagnetic data (Márton et al., 1992).

Timing of tide-influenced deposition

Fossils are rare in the Pétervására Sandstone. Molluscs are found only in the conglomeratic intercalations both in the lower and upper units (Fig. 4.2). The reconstructed mollusc assemblage represents the widespread Eggenburgian radiation of giant bivalves and allows dating of the Pétervására Sandstone (Báldi, 1986). During deposition of the Pétervására Sandstone strong tidal motions were recorded in the North Hungarian Bay (Tari et al., 1989). Paradoxically a bathymetry, which was favourable for the amplification of tidal motions, developed after a relative sea-level fall at the Egerian/Eggenburgian boundary (shortly before the Burdigalian, Sztanó & Tari, 1993).

The sequence of tidally influenced deposits can be divided into two major units: a lower aggradational and an upper progradational one (Fig. 4.2). The aggradation of the lower unit implies a delicate balance between sedimentation and relative sea-level rise. Deposition kept pace with a slow relative sea-level rise and the tidally-influenced environment was maintained (Sztanó & Tari,

1993). The influence of strong tidal motions is evident in the lower unit, but was weaker in the upper part, where bedforms are slightly smaller and bundle sequences are not common. Between the two units a minor flooding event can be recognized. This indicates a temporarily more intense relative sea-level rise. Dating of this flooding event in the North Hungarian Bay is crucial for further correlation with other tide-influenced deposits in the region. Despite the world-wide Burdigalian sea-level rise (Rögl & Steininger, 1983; Steininger et al., 1990; Haq, 1991), the North Hungarian Bay was filled up to sea-level during the following progradational phase of the Pétervására Sandstone (Fig. 4.2; Sztanó & Tari, 1993).

In conclusion, deposition of the tide-influenced Pétervására Sandstone started during a sea-level fall (Sztanó & Tari, 1993) prior to the Burdigalian (at the Egerian/Eggenburgian boundary; Fig. 6.2) and continued during the Burdigalian sea-level rise.

Implications of tidal deposits in the North Hungarian Bay

Physiography of the North Hungarian Bay

A shallow marine basin should fulfill some physiographic conditions in order to allow for the amplification of tidal motions. Most important are width, length, bottom topography, bathymetry and shape of the basin (Chapter 7).

The North Hungarian Bay was a funnel-shaped relatively narrow, but long embayment. The innermost, southern part is thought to have been shallow, but the outer, northern part was relatively deep (Fig. 6.3; cf. Báldi, 1986). The North Hungarian Bay was not wide enough to form an amphidromic system, but the dominant sediment transport towards the south and the north along opposite coasts of the bay and the time asymmetrical tidal motions were the result of topography-induced residual currents in combination with the basic tides (Chapter 7). Amplification of the tidal motions in the North Hungarian Bay occurred because the physiography was near conditions of resonance (Chapter 7). Considering the estimated "average water depth" of the North Hungarian Bay of around 40 m, resonant amplification would require a basin length of about 240 km (Chapter 7). From the southernmost tip to the northernmost outcrops at the mouth of the bay the length was 150 km (Fig. 6.3). At the mouth of the bay deep marine silty deposits (Szécsény Schlier) terminate abruptly, without any shallow marine equivalent to the north or to the east. The greatest izopachs of the Schlier (thickness over 1000 m) remain open there as well (Fig. 2.8). This indicates that the bay must have continued to the northeast and thus must have been considerably longer than 150 km.

In the northeastern direction, deepening upwards successions of the Szécsény Schlier reflect that during the Aquitanian the sea transgressed onto Triassic carbonates. At present, these Triassic rocks are exposed at the surface, because of Late Miocene - Quaternary uplift (Dunkl, personal communication, 1993). Sediments deposited in the potential northeast continuation of the bay thus must have been eroded.

Farther towards the northeast, where a continuation of the North Hungarian bay should have occurred, the East Slovakian Basin is situated. There, thick siltstones of Eggenburgian age are known, i.e., Presov beds up to 1000 m thickness (Vass & Cvercko, 1985). These deposits are supposed to be the lateral continuation of the Szécsény Schlier. The East Slovakian Basin is located at a distance of 80-100 km from the northernmost outcrops of the North Hungarian Bay. The length of the North Hungarian Bay (150 km) and the supposed continuation in Slovakia (80-100 km) together account for the length, which is required for resonant amplification of the tidal wave (Chapter 7).

Tidal pathways terminating in the North Hungarian Bay

There are only three possible marine connections (Báldi, 1986; Nagymarosy, 1990a), along which the tidal waves may have entered when the tidally influenced deposition in the North Hungarian Bay started (i.e., during the earliest Eggenburgian /latest Aquitanian/, Sztanó & Tari, 1993).

The first potential pathway is the Slovenian corridor, which formerly connected the North Hungarian Bay with the Mediterranean (Figs. 6.1 and 6.4). However, migration pattern of mammals, sedimentation in the south Alpine-Dinaridic foredeep in northern Italy and proofs of subaerial erosion in the southern part of the North Hungarian Bay indicate that the Slovenian corridor was closed before the late Aquitanian (Fig. 6.2; Steininger et al., 1985; Massari et al., 1986; Nagymarosy 1990 a,b).

The second potential pathway may have been through the "intramontane basins" of west Slovakia to the northwest (Fig. 6.4). Small basin remnants at Mále Karpaty Mountains., Dobra Voda Depression, Brezovske Karpaty, Váh River Valley, Bánovska Basin, Upper Nitra Valley were palaeogeographically part of an island archipelago (Kovác, 1986; Baráth & Kovác, 1989). At these localities Eggenburgian fan-delta conglomerates and deeper water siltstones (Luzice Schlier) occur. This area was connected with the Alpine Molasse Basin through the Vienna Basin (Fig. 6.4; Kovác et al., 1989; Jiricek & Seifert, 1990). In the Alpine foredeep marine deposits of late Aquitanian age can be followed only to the area of Munich (Bachmann & Müller, 1992). Further to the west, beds of the Lower Freshwater Molasse are known (Schoepfer & Berger, 1989). The Alpine foredeep was mostly dry land, and the tidal wave cannot have propagated along the Alpine foredeep. The west Slovakian "intramontane basins" were also in communication with the West Carpathian foredeep (Fig. 6.4). However, it is very unlikely that tidal motions first propagated towards the west, then turned to the southeast, crossing the island archipelago in west Slovakia before arriving to the North Hungarian Bay, and then were still powerful enough to produce a macrotidal setting.

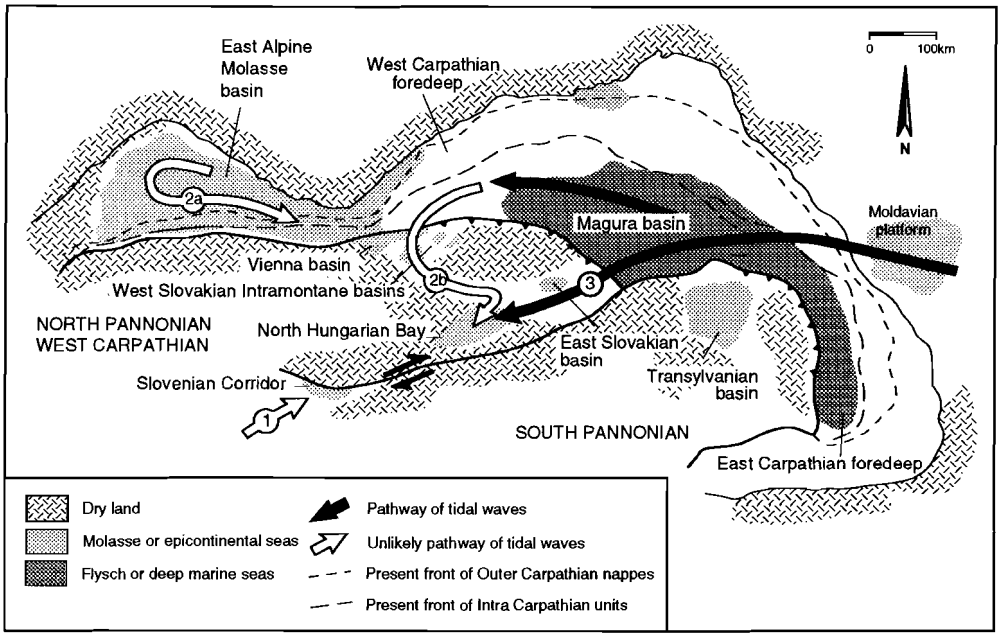

Fig. 6.4. Palinspastic and palaeogeographic reconstruction of the intra-Carpathian area. Numbers on the arrows refer to different routes described in the text.

The third possible pathway connects the North Hungarian Bay to the East Slovakian basin (Fig. 6.4), although the sedimentary record of the direct connection is eroded. In the East Slovakian basin some brackish-water deltaic deposits indicate nearby dry land (Vass et al., 1988). Fully open marine, deep water marly siltstones of Eggenburgian age (Presov Formation) are also known from this area (Vass & Cvercko, 1985). It is supposed that the East Slovakian basin was directly connected with the Magura basin and the Carpathian foredeep (Fig. 6.4; Vass et al., 1988; Kovác et al., 1993). Taking the Neogene shortening into consideration (cf. Cieszkowski, 1992), this foredeep was wide enough during the early Miocene to have allowed the propagation of tidal waves. This particularly holds for the eastern part of the West Carpathian foredeep, where deposition occurred in the Magura "residual" flysch basin (Fig. 6.4; Cicha & Kovác, 1990). The pathway from the Magura basin through the East Slovakian basin to the North Hungarian Bay (Fig. 6.4) therefore seems to be the most appropriate one for the transmission of tidal waves.

Tidal pathways beyond the Carpathian region cannot be demonstrated directly. As shown above, all seaways towards the Mediterranean were closed, so the only possibility is that tidal waves entered from the direction of the Eastern Paratethys through the Moldavian platform (Fig. 6.4). Few data are available about the early Miocene sedimentation in the Eastern Paratethys. The above presented data, however, lead to the conclusion that tidal waves, which finally arrived in the North Hungarian Bay, must have entered the Eastern Paratethys through the Indo-Pacific Seaway.

Early Miocene palaeogeographic changes and tidal deposits in the region: a discussion

The main palaeogeographic changes of the Paratethys region are known since long (Rögl & Steininger, 1983, 1984; Steininger et al., 1985). According to the analysis of various tide-influenced deposits in the region, the previously proposed Early Miocene scenario has to be modified.

During the early Aquitanian, tide-influenced estuarine deposits, estuary mouth bar deposits and offshore bioclastic sand ridges developed in the Venetian basin in northern Italy (glauconitic Orzes Sandstone, Massari et al., 1986). These deposits were formed in the Mediterranean, but near the southern outlet of the Slovenian corridor (Fig. 6.5). Tidal currents passing through the Slovenian corridor must have contributed to the amplification of tidal motions in the Venetian basin. When the Slovenian corridor was closed, i.e., before the late Aquitanian, the tidal influence ceased and restricted marine conditions developed in the Venetian basin (Massari et al., 1986).

During the late Aquitanian tidal influence started in the North Hungarian Bay, north of the closed Slovenian corridor (Fig. 6.5), and continued during the early Burdigalian. Tidal waves could not enter from the Mediterranean anymore, nor could they have come from the peri-Alpine foredeep. Instead, they must have travelled through the Eastern Paratethys, entered the Carpathian flysch basin and continued their way up into the North Hungarian Bay (Fig. 6.5).

The eustatic sea-level rise during the early Burdigalian was significant in the Paratethys, as well as in the Mediterranean (Rögl & Steininger, 1983). Nevertheless, the Slovenian corridor, which had been closed by transpressional tectonics, remained inactive. The peri-Alpine foredeep, however, became flooded from the direction of the Rhone valley and from the Central Paratethys as well (Rögl & Steininger, 1984; Schoepfer & Berger, 1989). In addition to eustasy, subsidence of the Alpine foredeep may also have contributed to this flooding (cf. Allen & Bass, 1993). Thus gradually a "new" seaway between the Mediterranean and the Paratethys was formed (Fig. 6.5).

In about the same time only a minor flooding took place in the North Hungarian Bay (Sztanó & Tari, 1993). The intensive sedimentation kept the basin shallow enough (Sztanó & Tari, 1993) to maintain tide-influenced sedimentation. By the time the sea completely flooded the peri-Alpine foredeep, the North Hungarian Bay had been gradually filled up above sea-level, and the "Palaeogene sedimentary cycle" came to an end in Hungary.

During the Burdigalian the rising sea flooded some of the islands in the western Slovakian archipelago, and thus the connection to the Alpine foredeep widened (Fig. 6.5; Kovác et al., 1989). Tidal waves from the Eastern Paratethys, formerly "trapped" in the North Hungarian Bay, may have continued their way to the long peri-Alpine strait (Fig. 6.5). Tidal waves also could propagate directly from the Mediterranean through the Rhone valley. One of these or both have been amplified and interacted in the Peri-Alpine strait, and have produced the famous tidal deposits of the Upper Marine Molasse in Switzerland (Homewood & Allen, 1981; Allen et al., 1985; Allen & Bass, 1993) and in Upper Austria (Faupl & Roetzel, 1987, 1990; Krenmayr, 1991). Amplification

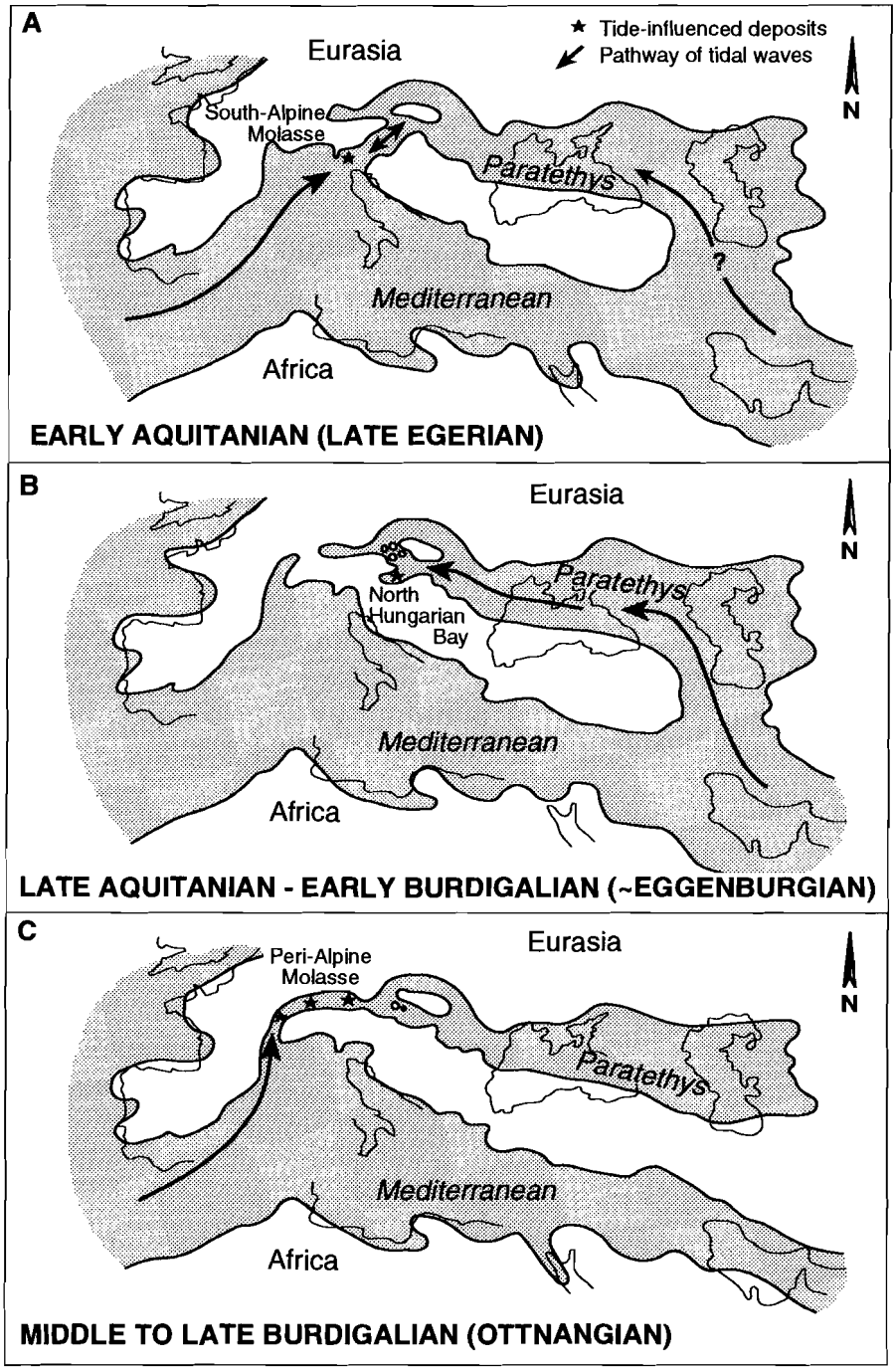

Fig. 6.5. Palaeogeographic sketches of the Early Miocene Paratethys (A and C after Rögl & Steininger, 1983; B intermediate scenario proposed in this chapter).

probably occurred by means of resonance. Reflection of the tidal wave may have occurred near the tip of the Bohemian Massive, where the strait was narrowest.

In Switzerland these tidal deposits were dated as early Burdigalian (late Eggenburgian) (Schoepfer & Berger, 1989) and in Austria as Ottnangian (late Burdigalian) (cf. Faupl & Roetzel, 1987, 1990). The shift in time and locus of tidal amplification within the Peri-Alpine strait most likely occurred due to the gradual flooding of the strait. Due to the relatively poor resolution of biostratigraphic data it should not be excluded, however, that both tidal depositional systems are of the same age, most probably Ottnangian, and thus amplification took place in the same oceanographic setting. Further biostratigraphic studies are needed to verify this assumption.

Sedimentary successions at various localities in the Paratethys show that before the late Early Miocene the Indo-Pacific seaway was blocked by the collision of the Arabian and Eurasian plates (Rögl & Steininger, 1983). This must have resulted in a gradual change of the circulation system. After this blocking, tidal influence came only from the Mediterranean to the Paratethys. Shortly afterwards the marine connection through the peri-Alpine foredeep was closed, resulting in the second isolation of the whole Paratethys (Rögl & Steininger, 1984; Steininger et al., 1985). Afterwards no more tide-influenced sediments were formed in the Paratethys.

As is demonstrated by the examples of tide-influenced deposits in the Paratethys, amplification of tidal forces in all cases occurred locally, either in shallow embayments or in straits which connected the Paratethys and the Mediterranean Sea. Tectonic activity of the Alp-Carpathian-Dinarid orogenic wedge resulted in changes of subsidence, sedimentation rates, and thus in changes of palaeogeography, and the opening and closure of seaways. The resulting changes of oceanic current systems were, in turn, reflected by shifts in loci of tide-influenced deposition.

Conclusions

In the North Hungarian Bay local resonant amplification of the tidal wave occurred during the Late Aquitanian-Early Burdigalian (mainly Eggenburgian), and led to the deposition of the tide-influenced Pétervására Sandstone. This happened because during the Eggenburgian, the physiography of the bay met the conditions needed for amplification of the tidal wave, and because a "simple" pathway towards fully open marine areas existed at the same time. The only pathway, which was wide and straight enough to allow the propagation of tidal waves, was the connection with the East Slovakian Basin. This basin was connected through the Carpathian flysch and molasse basins with the Eastern Paratethys. Tidal waves thus must have entered the Eastern Paratethys through the wide seaway.

The presence of tide-influenced depositional systems at various places in the Mediterranean and the Paratethys during the early Miocene can be explained by the local

palaeogeographic scenario controlling the amplification of tidal forces and the sedimentary system. In all cases a freely propagating tidal wave was obviously present.

The example of the Pétervására Sandstone and the other tide-influenced deposits in the region clearly demonstrates their significance for palaeogeographic reconstructions. The presence of meso- and macrotidal deposits, which are the result of local amplification of tidal motions, can be used as a tool for such reconstructions. They reveal the contemporaneous existence of connections wide and deep enough to allow the propagation of the tidal wave induced in open marine water masses.

CHAPTER 7
AMPLIFICATION OF TIDAL MOTIONS IN THE EARLY MIOCENE NORTH HUNGARIAN BAY

Abstract

Following the Late Aquitanian sea-level fall, tide-influenced deposition started in the North Hungarian Bay, an embayment in the Paratethys open to the northeast. The relatively narrow, funnel shape of the bay supported amplification of tidal movements, resulting in the generation of strong tidal currents. The length and the depth of the North Hungarian Bay and the connecting seaway through East Slovakia fulfilled the conditions needed for resonant amplification of semidiurnal (M2) tides. Tide-influenced deposits were formed at both sides of the North Hungarian Bay. They reflect dominant currents into opposite directions and of different strength at either side of the basin. This difference was the result of bottom-tide interactions. Cyclonic (anticlockwise) residual currents were induced above the asymmetrical central depression in the bay and were superimposed upon the tidal currents, producing an anticlockwise, asymmetrical current system.

The North Hungarian Bay and other examples show that amplification of tidal motions and formation of tide-influenced deposits may occur if basin dimensions pass through the "Tidal Amplification Window". This window represents ideal conditions for resonant or amphidromic amplificatation of tidal currents and determines an ideal length/depth or width/depth ratio relative to the wave length of the astronomical tides. Thus signs of strong tidal influence in fossil basin fills may be used to reconstruct the dimensions (length, depth and width) of such basins.

Introduction

The physiography of a marine basin determines the potential controls of the dynamics of the system. The shape, depth and orientation of a basin control whether, and to which extent, currents generated by winds, tides or both can affect sedimentation. The influence of tides is most often recognized in relatively shallow marine sediments, particularly when they were formed in a meso- or macrotidal setting (cf. Johnson & Baldwin, 1986). Commonly microtidal settings can be recognized in the geological record only if the tidal basin was large enough for considerable water masses to flow in and out during each tidal cycle, and other potential forces (e.g. winds) did not overshadow the effects of the relatively small tidal amplitude. An example of such conditions is found in the present-day Wadden Sea, where a relatively small tidal amplitude combined with a large drainage basin leads to relatively strong tidal currents (Oost & de Boer, 1994).

Tidal waves generated in the open ocean, can be amplified when they propagate into

shallow marine basins (Pugh, 1987). Amplification of the tidal wave depends on the morphology of the receiving basin, which thus determines the tidal range and the strength of the resulting tidal currents.

The palaeogeography of ancient seas is less well known than that of modern seas. However, in fossil cases, the physiography of a basin can be deduced from indirect evidence provided by tidal deposits. Tide-influenced deposits, particularly in semi-enclosed or partly isolated fossil basins, may provide information about connections to the open sea (Chapter 6), and about the morphology, the depth and the length of the basin.

Early Miocene clastic deposits in the North Hungarian Bay clearly show the characteristics of a tide-influenced depositional system (Sztanó, 1992a; Sztanó & Tari, 1993; Chapter 3 & 5). Considering the scattered distribution of lower Miocene sediments in adjacent areas, this bay seemingly had restricted connections to other marine basins of the Paratethys (cf. Nagymarosy, 1990a). However, the tide-influenced deposits in this bay indicate that an open marine connection must have existed, e.g. through the East Slovakian basin, in order to allow the access of tidal motions (Chapter 6).

The aim of this chapter is to analyse how amplification of tidal motions occurred in the North Hungarian Bay during the early Miocene. This will be done by using simple equations, which describe the behaviour of tidal waves in a given basin setting. The result is a semi-quantitative estimate of the local hydrodynamic regime and of the dimensions of the basin and of the seaways connecting the North Hungarian Bay with open marine waters.

Geological and palaeogeographical setting

From the Oligocene onward a large "inland" sea, the Paratethys developed in the Alpine-Carpathian orogenic belt (Fig. 6.1; Báldi, 1980; Rögl & Steininger, 1983). It consisted of a chain of basins, each with a different structural development. Characteristic changes of the aquatic biota can be correlated between the different basins. This indicates that a free communication of water masses within the Paratethys existed (cf. Báldi, 1980; Nagymarosy, 1990a; Steininger et al., 1990).

During the early Miocene (regional Eggenburgian stage; Fig. 6.2) tide-influenced deposition started in the North Hungarian Bay, an embayment of the Paratethys (Fig. 6.1). Although the palaeogeography of the Central Paratethys is not completely clear, geological evidence suggests that all seaways towards the Mediterranean Sea and towards the North Atlantic Ocean were closed when tide-influenced deposition began within the North Hungarian Bay (Sztanó & Tari, 1993). This implies that the Paratethys should have been sufficiently wide towards the east to allow tidal waves to propagate and to approach the North Hungarian Bay (for a discussion of potential tidal pathways see Chapter 6).

The tidal movements in the Paratethys may have been relatively weak, as often happens in

"open" marine systems. However, in shallow marginal seas amplification may occur. This must have been the case in the North Hungarian Bay during the early Miocene. This bay was connected with the Paratethys through the East Slovakian Seaway, northeast of the North Hungarian Bay (Fig. 6.4). The sedimentary record in this seaway hardly survived the later geological evolution of the area; lower Miocene (Eggenburgian) deposits have been preserved only in the East Slovakian basin (Vass & Cvercko, 1985). Nevertheless the tidal wave must have travelled a relatively long way (more than 100 km) in shallow "epicontinental" seas, through the East Slovakian Seaway before entering the North Hungarian Bay (Fig. 6.4).

A relatively strong influence of tidal motions has been recorded by sand waves in the shallow-neritic Pétervására Sandstone along the eastern coast of the bay (Fig. 6.3). A weaker tidal influence was observed in the littoral Budafok Sand along the western coast of the bay (Fig. 6.3; Báldi, 1986). These deposits along the margins of the bay were formed when a rapid, significant shallowing occurred, as the result of a relative sea-level fall before the Burdigalian (Sztanó & Tari, 1993). In the same time deposition of silty formations (Szécsény Schlier) continued in the deeper part of the embayment, at a shallow bathyal to deep neritic depth (cf. Báldi, 1986; Vass et al., 1988).

Physiography of the North Hungarian Bay

The North Hungarian Bay was an elongate, funnel-shaped bay with a wide mouth, and was progressively narrower to the south (Fig. 6.3). It was open to the northeast, where it joined the East Slovakian Seaway (Chapter 6).

The highly asymmetrical bathymetry of the bay (Fig. 6.3) was most likely the product of the tectonic evolution of the basin (Tari et al., 1993). The sedimentary structures of the Pétervására Sandstone, particularly the size of the large sand waves, indicate a depositional depth of around 20 m at the eastern side of the bay (cf. Allen, 1984). Towards the centre of the bay, where the Szécsény Schlier was formed simultaneously, a gradually steeper slope is supposed. Deposition of the Schlier may have started at a depth of 60 m, but generally occurred in a range of 100 - 300 m, as is indicated by faunal communities (Horváth, 1972; Báldi, 1986; Báldi & Nagy-Gellai, 1990). A relatively steep slope must have been present between the central depression and the Budafok Sand in the west, which was formed in the narrow littoral - sublittoral zone at depths of up to 30 m (Báldi, 1986).

The "average depth" of the bay is an important element for the estimate of the behaviour of the tidal wave. It is deduced from the above depth estimates in combination with the corresponding surface areas. In case of the North Hungarian Bay a small difference in the estimate of the depositional depth of the shallow marginal areas has no significant influence on the "average depth" of the bay. Instead it strongly depends on the great depth (over 150 m) of the relatively small central depression (Fig. 6.3). The avearage depth of the bay is estimated to have been around

40 m.

From an oceanographic point of view the East Slovakian Seaway and the North Hungarian Bay should be regarded as one basin. The length of the North Hungarian Bay in the SW-NE direction was 150 km (Fig. 6.3). The length of the East Slovakian Seaway from the North Hungarian Bay to the open Paratethys must have been more than 100 km. This gives a total length of over 250 km. It is difficult to estimate the depth conditions in the East Slovakian Seaway, since hardly any sediment has been preserved there (Vass & Cvercko, 1985). Nevertheless it is postulated that deposition of shallow bathyal-deep neritic silty sediments occurred over the full length of the seaway, from the Szécsény Schlier in the North Hungarian Bay to the Presov Schlier (Vass & Cvercko, 1985) in the East Slovakian Basin. In addition schlier deposits must have had time-equivalent deposits at shallow depths along the margins of the seaway. Therefore the average depositional depth in the East Slovakian Seaway may have been similar, or even greater than in the North Hungarian Bay. This implies that the average depth of the combined North Hungarian Bay and East Slovakian Seaway must have been in the range of 40-50 m on the average.

Sedimentary facies of the tide-influenced deposits

ThePétervására Sandstone

The shallow marine Pétervására Sandstone crops out in the northeastern part of the bay, and covers an area of about 1500 km^2 (Fig. 2.8). It consists of four shore-parallel facies units (Fig. 3.15, 4.2; Sztanó 1992a,b; Sztanó & Tari, 1993 Chapter 3), which were formed from offshore to shore in increasingly shallower water. The most distal facies unit consists of fine, rippled, silty sand, which represents the transition from the shallow marine Pétervására Sandstone towards the time-equivalent part of the Szécsény Schlier which was deposited further offshore (Fig. 6.3). In the direction of the shore the rippled silty sand interfingers with fine- to medium-grained sandstones, which are characterized by decimetre-scale (0.1-0.6 m) crossbedding. The next facies towards the shore is medium- to coarse-grained sand with large-scale crossbedding with sets of 1 m to 12 m height. Nearest to the coast, 4-8 m high conglomerate lobes occur (Fig. 3.15). The coastline was tectonically controlled (Chapter 4), therefore only a very narrow, steep beach may have been present.

The influence of tidal motions is less evident in facies units deposited in deeper water and more expressed in the relatively shallow water. The foresets of the decimetre-scale crossbedding are occasionally covered by mud drapes (Chapter 3), which must have formed during slack water periods in a tide-influenced system (cf. Visser, 1980). An area of about 500 km^2 was covered by such decimetre-scale dunes (Fig. 3.15).

The foresets of large-scale cross-beds provide even better evidence of tidal influences. Mud drapes on the large foresets and reactivation surfaces are rare, but tidal influence is clearly

reflected by periodical variations in grainsize and thickness of foreset laminae. Spring-neap tidal activity with a clear reflection of the diurnal inequality of the tide is apparent from variations in foreset thicknesses (Chapter 5). These large-scale crossbeds with a record of tidal cyclicity were produced by migrating sand waves (cf. Allen, 1980). The internal structure of these deposits indicates strong, highly asymmetrical tidal currents (cf. Allen, 1980). A large area of more than 500 km^2 was covered by these sand waves driven by strong tidal currents (Fig. 3.15).

The distribution of both the medium- and large-scale dunes points to a sheet-like morphology of the tide-influenced deposits. No channel-fill or intertidal deposits were found. West of the tectonically controlled coast, the bottom of the bay was covered by a field of sand waves at a depth of about 20 m (Chapter 3). Further offshore the large sand waves were rimmed by a field of decimetre-scale dunes gradually diminishing and interfingering with offshore silts. This areal arrangement of facies units is similar to the one observed in some modern macrotidal, sandy offshore areas (Belderson et al., 1982).

The Budafok Sand

The Budafok Sand crops out in a narrow patch along the western coast of the North Hungarian Bay. It covers a significantly smaller area and has a smaller volume than the Pétervására Sandstone. A variety of sedimentary structures is present, but poor outcrop conditions prevent a detailed reconstruction of facies units, as well as their lateral and vertical relationships.

There are massive beds, up to 3 m thick, consisting of poorly sorted granule to very coarse sand with accumulations of large shells. At a few localities large-scale crossbedding was found in coarse and pebbly sand. Upon these metre-scale gently (5-15°) inclined surfaces, decimetre-scale crossbedded foresets are superimposed, dipping into the same direction. Upward coarsening successions are present, with massive medium-grained sand at the base and fine sandy conglomerate at the top. In the fine-grained sand horizontal bedding, occasionally with ripples alternating with organic matter-rich silty laminations is the characteristic structure. From one locality (Budafok, Pacsirta Hill; Fig. 2.8) vertically accreted sand/silt laminites were reported, and interpreted as tidal mudflat deposits (Báldi, 1959). Small- and large-scale current-generated structures in the Budafok Sand reveal the influence of moderate tidal currents, weaker than at the other side of the bay where the Pétervására Sandstone was formed simultaneously.

Burrows and a great variety of well-preserved mollusc shells in the Budafok Sand indicate a prosperous faunal life in shallow normal salinity water (e.g.: *Anomia* indicates 2-3 m depth, some *Ostrea* species are representative of intertidal to shallow subtidal conditions, and the giant *Pectinids* are indicative 20-30 m depth; Báldi, 1986). The above sedimentary structures in concert with the palaeoecological information suggest a coastal tide-influenced depositional environment.

Sediment transport in the North Hungarian Bay

Foreset dip directions of decimetre-scale dunes, tidal sand waves and conglomerate lobes indicate an unimodal transport towards the north during deposition of the Pétervására Sandstone along the eastern coast of the bay (Fig. 6.3). Considering the tide-influenced depositional environment this points to a highly asymmetrical current system, with dominant (ebb) currents towards the mouth of the bay in the north. The lack of clear reactivation surfaces produced by the subordinate tide (cf. Visser & de Mowbray, 1982), indicates that the subordinate current was usually below the critical velocity needed for sediment transport. Statistical analyis of the variation of foreset thicknesses (Chapter 5) revealed that also only part of ebb currents, i.e., those around spring-tides, were strong enough to transport the coarse sand and to move the sand waves. The sand waves of the Pétervására Sandstone fall into class IIIA of Allen (1980).

Bedforms in the Budafok Sand, at the western side of the bay reflect a net transport towards the south-southeast. Previous studies of faunistical and petrological characteristics of the Budafok Sand also suggest an overall southward sediment transport (Fig. 6.3; Báldi, 1986). The large-scale compound cross-beds resemble sand waves of class IVA-V of Allen (1980), composed of low angle master bedding (reactivation surfaces of Visser & de Mowbray, 1982) covered by medium-scale dunes. The reactivation surfaces show that the subordinate current must have been strong enough to partly erode the foresets produced by the dominant tidal current. The subordinate current was, however, less powerful than the dominant current; no structure built by the subordinate current has been preserved (cf. Allen, 1980). The relative asymmetry of the tidal currents thus was smaller, and the dominant current in the Budafok Sand at the western side of the basin was towards the south, while it was to the north in the Pétervására Sandstone at the eastern side.

The different style of sedimentary structures and the opposite transport directions reflect significantly different hydrodynamic conditions along the eastern and the western sides of the bay. The most important differences were tidal current velocities, depositional depth and sediment characteristics.

The considerably smaller volume of the Budafok Sand indicates that a smaller amount of sand was transported from the north along the western shoreline and deposited in the southern end of the embayment, than was brought into the system and moved northward along the eastern side. The petrographic difference between the dominantly quartzose Budafok Sand and the lithic Pétervására Sandstone (Báldi, 1986; Vass et al., 1988) moreover indicates that, although the current pattern in the bay seems to have been rotary (Fig. 6.3), the transport of sediment was not. The quartzose, coarse, pebbly sand introduced into the bay at the western side, thus never reached the eastern part of the bay. A major influx of lithic sand from large rivers and/or erosion of older deposits must have occurred in the south-southeastern part of the basin (cf. Chapter 4). This huge amount of lithic sand was subsequently mixed with a minor amount of ophiolite-derived clastic material which was supplied to the bay from the east along the Darnó fault (Fig. 2.8, Chapter 4).

Factors controlling the modification and amplification of the tidal wave

Tidal waves have a very long wave length. They are modified in shallow marine areas, mainly due to tide-bottom interactions (Pugh, 1987), which are largely governed by the depth of the shallow marine basin. Depending on the size of the basin, tide-bottom interactions may result in the amplification of the tidal wave. Before analyzing the mechanism which produced the strong tidal currents in the early Miocene North Hungarian Bay, the processes which may amplify the tidal wave in semi-enclosed basins will be shortly discussed.

Amphidromic systems

As any moving body in the northern hemisphere, the propagating tidal wave is deviated to the right by the Coriolis force due to the rotation of the Earth. Therefore in a basin the propagating tidal wave tends to turn around an amphidromic point in an anticlockwise direction, with an increasing amplitude away from the amphidromic point. The increase of amplitude with the increasing distance from the amphidromic point is well known along the coasts of the North Sea (e.g. the German Bight). However, a basin should have a certain width to accommodate the rotary tidal wave and thus to form an amphidromic system (Pugh, 1987). The so-called Rossby radius of deformation (R), which describes the exponential decay of a tidal wave (Kelvin wave), should be comparable with the basin width (Pugh, 1987).

$$R=(g*d)^{1/2}/f \qquad (1)$$

where g is acceleration of gravity (9.81 m/s^2), d (m) is the average water depth of the bay and f is the Coriolis coefficient ($10.3*10^{-5}/s$ at 45° latitude; Fig. 7.1). For a given latitude the potential development of an amphidromic point thus depends on the water depth and on the width of the basin. For any basin a range of these related variables defines if an amphidromic tidal system can develop (Fig. 7.1).

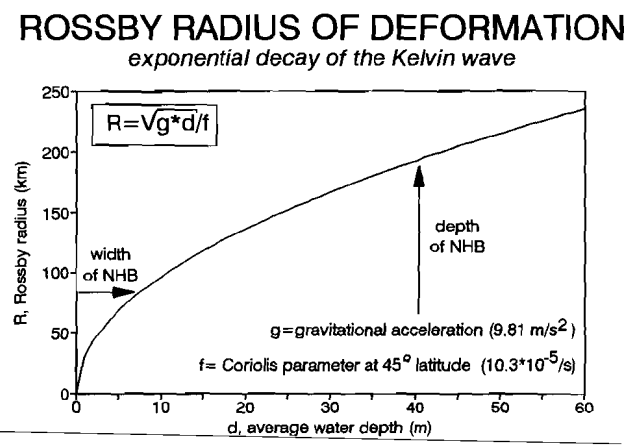

Fig. 7.1. The Rossby deformation radius of the Kelvin wave describes how wide a basin should be for the development of an amphidromic system. The North Hungarian Bay (NHB) was too narrow for the rotation of the tidal wave.

The effect of the Coriolis force, however, is felt even in narrow basins, such as channels (e.g. major tidal inlets of the Dutch Wadden Sea). This simply means that during the flood the water table is higher at the right side of the incoming flow, and that during the ebb it is higher at the right side of the outgoing flow. Thus at one side of the channel the flood, at the opposite side the ebb is the dominant current. A similar situation may occur in large but narrow basins (Gerritsen, 1990).

Resonant amplification

The tidal wave may also be intensified, particularly in long embayments, where it is reflected at the head of the bay. The resulting pattern of interference depends on the length of the wave with respect to the length of the basin. A standing wave or resonance may develop, if the length of the basin approximates a given proportion of the tidal wave length (Pugh, 1987). E.g. in the southern part of the western Wadden Sea a standing wave occurred, because the length of the sea was approximately half of the length of the propagating tidal wave (Klok & Schalkers, 1980). After closing part of the embayment with a dyke, by which the length of the basin was about halved, the tidal amplitude in the remaining part increased.

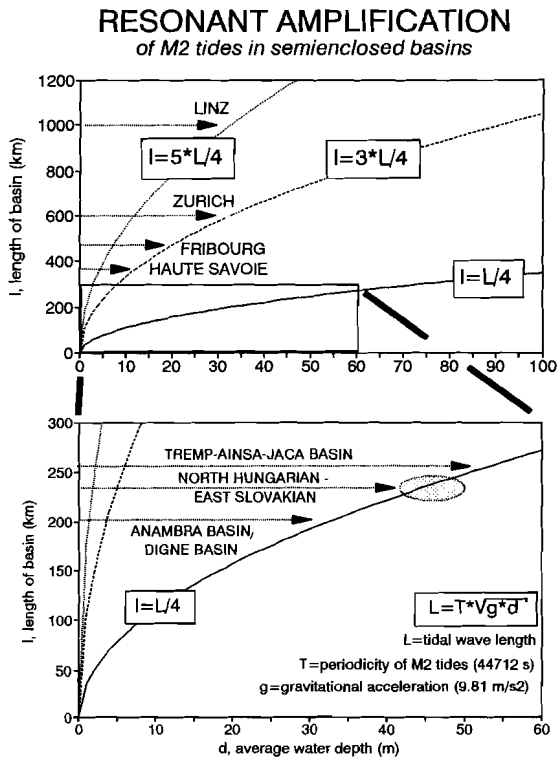

Fig. 7.2. The average depth of a basin determines the velocity and the length of the propagating tidal wave. For resonant amplification the critical basin length should be an odd multiple of the quarter of the tidal wave length. Lengths of basins is shown, in which resonant amplification of the M2 tide may have occurred (see text in the

discussion for references).

Resonant amplification can generate extremely high tidal ranges in case of favourable geometric conditions (cf. Lisitzin, 1974; Komar, 1976). A best known example of resonant amplification occurs in the Bay of Fundy in Canada, where it produces the highest tidal range observed in modern seas (Dalrymple et al., 1990). Resonant amplification occurs if the length of the bay (l) is close to odd multiples of the quarter of the tidal wave length (L) (Pugh, 1987):

(2) $l=(2n+1)*L/4, \; n=0, 1, 2, 3, ...$

thus $l=L/4, 3*L/4, 5*L/4, ...$

The smallest basin length required is about the quarter of the length of the tidal wave. The tidal wave length (L) can be determined, if water depth (d) is known:

(3) $L=T*(g*d)^{1/2}$

where, T is the period of the tide (M2=12.42 h). If the average depth of a basin is known, a calculation of the quarter tidal wave length indicates how long such basin must have been to produce resonance of a given tide (Fig. 7.2).

In a similar way, if geological evidence provides data on the length and depth of the bay the natural period of oscillation (t) of the water-mass can be calculated (Pugh, 1987), which is

(4) $t=4*l/(g*d)^{1/2}$

in the simplest case. This is the period for which the water mass of a basin is sensitive for externally forced oscillations, such as tidal motions. Thus if the natural period of oscillation (t) of the basin approximates the periodicity (T) of tidal motions, it easily oscillates on the tidal frequency (Fig. 7.3; Pugh, 1987). Thus even the type of the tide (semidiurnal, diurnal, etc) can be determined, if the size of the basin is known (Pugh, 1987).

Modification

The topography of the bottom is also crucial in modifications of the propagating tidal wave. The most important phenomena produced by bottom interactions are vorticity, which in advection causes residual currents, and higher-order harmonics of basic tidal motions (e.g. M4 tides with periods of 6 h). These, in combination with the basic tide, induce time and velocity asymmetrical tidal motions (Pugh, 1987).

Vorticity may develop when oscillating currents pass slightly obliquely over a positive relief (Robinson, 1983). Thus an *anticyclonic* (clockwise) residual vorticity is generated on the northern hemisphere (Zimmerman, 1981). This is the result of a net advection by the tidal current, the bottom friction and the Coriolis effects. This explains why sand is transported clockwise all around sand banks and tidal sand ridges, as has been observed in several studies in recent environments (cf. Stride, 1982). This process also explains the oblique orientation of sand ridges with respect to tidal currents.

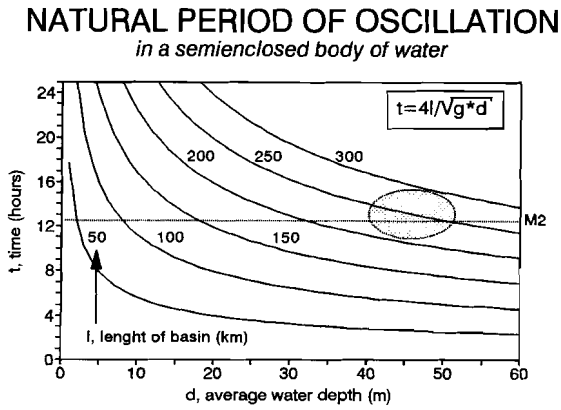

Fig. 7.3. The natural period of oscillation of basins of different length depends on water depth. The period of the M2 tide (12.42 hours) is indicated by the horizontal line. Crosspoints of the M2-line with time-lines indicate the length and depth, which are required for conditions near resonance of the semi-diurnal tide. In the 150 km long North Hungarian Bay resonant amplification would have occurred in case of only 20 m average depth. The North Hungarian Bay and the East Slovakian Seaway together exceeded a length of 250 km; for that situation an average depth of 40-50 m would have provided the conditions needed for resonant amplification.

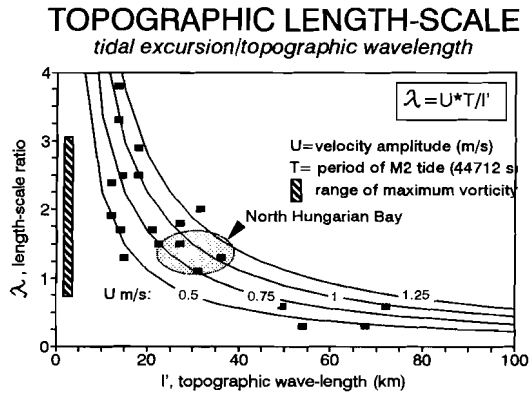

Fig. 7.4. The relation between the dimensions of bottom irregularities, tidal current velocities and the topographic length scale ratio. The range of values where maximum vorticity and thus residual currents can be formed is shown by the vertical bar (Zimmerman, 1981). The lines show the theoretical values given by the equations, the dots are examples after Zimmerman (1981).

The same physical process is also valid in case of depressions. Tidal currents passing obliquely over a depression also produce a residual vorticity, which is *cyclonic* (anticlockwise) in the northern hemisphere (Robinson, 1983). The length of any negative or positive topographic feature (e.g. depression or sand ridge) should be of the same order of magnitude as the horizontal tidal excursion is, which is the route of an imaginary particle travelling with the tidal currents during one tidal cycle. This is expressed by the topographic length-scale ratio (λ; Zimmerman, 1981):

$$\lambda = U*T/l', \qquad (5)$$

where U is the velocity amplitude of the tidal currents (i.e., the maximum current velocity), T is the period of the tidal wave, and l' is the wave length of the topography (i.e., the size of the depression or the ridge in the direction of wave propagation; Fig. 7.4). Both field data and theoretical calculations show that, for common tidal velocities and dimensions of tidal sand ridges, λ of about 1.9±1.2 provides conditions for maximum residual circulation (Zimmerman, 1981).

Generation of strong tidal motions in the North Hungarian Bay

Amphidromic systems

The sedimentary structures in the Pétervására Sandstone and to a lesser extent in the Budafok Sand indicate that during deposition strong tidal currents were active. Considering that there was not any storage basin south of the North Hungarian Bay (Fig. 2.8), these strong tidal currents must have been due to a high tidal range in the bay (Chapter 5). Moreover, no signs of tidal motions have been recognized in adjacent parts of the Paratethys system. Therefore tidal motions in the bay must have been amplified.

Dominantly southward sediment transport in the western part of the North Hungarian Bay and northward-directed transport along the eastern side suggest a net anticlockwise circulation of water currents within the bay. Therefore it would be attractive to propose a rotary tidal front model, i.e., a tidal current system with an amphidromic point. Considering the sediments and the sedimentary structures found (see above) the dominant tidal forces must have been weaker at the western side and stronger at the eastern side of the bay. In such case an amphidromic system would have neccesarily been eccentric, similar to those observed in the English Channel and in the Irish Sea, (cf. Komar, 1976; Pugh, 1987). The width of the North Hungarian Bay, however, was 80 km at maximum (Fig. 6.3), which is insufficient to develop any amphidromic sytems in a basin with an average water depth of 40 m (Fig. 7.1). For the development of amphidromic systems in an 80 km wide basin, the average depth should have been about 8 m (Fig. 7.1), which is clearly in conflict with the available palaeoecological (Báldi, 1986) and sedimentological data. Thus the presence of an amphidromic tidal system during deposition of the Pétervására Sandstone and the Budafok Sand can be excluded.

Resonant amplification

Another way of amplification of tidal motions in semi-enclosed basins is resonance. An elongate, funnel-shaped bay with a wide mouth, which becomes progressively narrower inward, such as the North Hungarian Bay (Fig. 6.3), supports resonant amplification of tidal motions (Pugh, 1987). With respect to potential resonant amplification of the tidal wave the North Hungarian Bay should be considered in conjunction with the East Slovakian Seaway, which formed the connection to more open parts of the Paratethys (Fig. 6.4). The average water depth of the North Hungarian Bay and the East Slovakian Seaway was about 40-50 m. At this water depth the tidal wave length is of the order of 1000 km (Fig. 7.2). The combined length of the basins was more than 250 km. This approximates a quarter tidal wave length, and thus may have provided favourable conditions for near-resonance with a natural period of oscillation of around 12 hours, close to that of M2 tides (12.4 h; Fig. 7.2 & 7.3). This, together with data on lateral growth of sand waves (Chapter 5), strongly suggests that resonant amplification of semidiurnal tidal motions occurred in the combined North Hungarian Bay - East Slovakian Seaway. It happened during a certain stage in their evolution, when a favourable depth and length for resonant amplification were met.

Tide-induced residual currents

As decribed above, the sedimentary structures in the Pétervására Sandstone along the northeastern part of the bay indicate a dominant tidal current (ebb) to the north, while those in the Budafok Sand in the western part of the bay indicate a dominant tidal current (flood) to the south.

The direction of propagation of the tidal wave and the tidal currents must have been subparallel to the depression in the centre of the bay (Fig. 7.5). This depression was comparable in length (tens of kms) to large tidal sand banks (cf. Stride, 1982; cf. Belderson, 1986), although in magnitude it was much larger (minimum depth more than 100 m; Báldi, 1986; Fig. 7.5). Velocity amplitude of the tidal currents must have been 0.75-1 m/s (Chapter 5). Combined with the length of the depression (25-50 km), this gives a topographic length scale ratio in the North Hungarian Bay of around 0.7-1.8 (Fig. 7.4), which falls into the range where residual vorticity can be formed (cf. Zimmerman, 1981; Fig. 7.4). Considering the observed sedimentary facies and the geometric configuration of the North Hungarian Bay it is very likely, that a strong anticlockwise residual circulation around the central depression of the bay, also contributed to the observed oppositely directed asymmetry of the tidal currents at either side of the bay.

Typical residual current velocities are of the order of 1-10 cm/s, in general they are less than 10% of the tidal currents (Zimmerman, 1981). Since an amplified tidal wave was present in the embayment, tidal current velocities were also enhanced, which thus must have resulted in relatively small, but significant residual currents. These residual currents enhanced those tidal currents which flew in the same direction and weakened currents which flew in the opposite direction (Fig. 7.5). Over one tidal cycle the vector sums of the resultant currents, which finally determined sediment transport, were opposite at both sides of the bay. This fits well to the observed net sediment transport to the south in the western part of the bay and to the north in the eastern part of the bay (Fig. 7.5).

Towards a semi-quantitative estimate of the current velocities: a model

A combination of the residual currents and the "basic" ebb and flood currents, which were evolved in the generation of vorticity, well explains the observed asymmetry in the direction of the dominant sediment transport along opposite sides of the North Hungarian Bay (Fig. 7.5).

Above a number of factors has been discussed which could have caused the asymmetry of those ebb and flood currents within the North Hungarian Bay. The deformation of the tidal wave likely was induced by bottom-interactions, developed during propagation in the East Slovakian Seaway, by the asymmetric geometry of the North Hungarian Bay, the different depositional depths and the different sediment properties (e.g. grain size) on opposite sides of the bay.

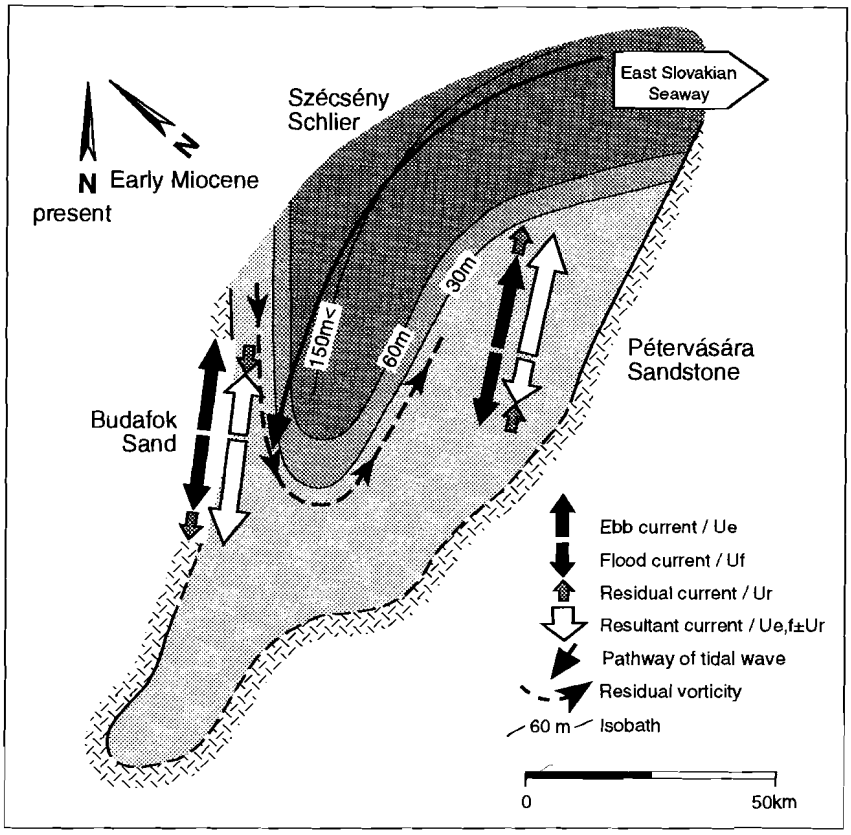

Fig. 7.5. "Basic" flood and ebb currents (U_e, U_f) in combination with bottom-topography-induced residual currents (U_r) acted on the water masses of the North Hungarian Bay. This resulted in flood dominance along the western coast, and in ebb dominance along the eastern coast of the bay. Note that the flood and ebb currents envolving in the generation of vorticity are asymmetrical (U_e>U_f).

The potential effect of the Coriolis force, to raise the water level at opposite sides of a bay during opposite phases of the tides may also added to the development of the tidal asymmetry. Higher harmonics, like forth-diurnal tides (M4 with a period of 6h), may have been induced by the irregularities of the bottom and/or the shoreline, and superimposed on the basic tidal periodicities they could have produced any asymmetry of the effective tides (Pugh, 1987). These factors together obviously have significantly enhanced the asymmetry of the tidal wave in the North Hungarian Bay. The aim of the following simple model is only to approximate the rate of asymmetry of the "basic" tidal currents.

The internal geometry of the sand waves in the Pétervására Sandstone indicates that during the ebb large amounts of sediment were transported northward, and that during the flood the tidal currents were not strong enough to erode lee faces of the sandwaves and to transport sediment along the eastern side of the bay. Thus on the Pétervására side the combined residual and tidal

currents (see resultant vector in Fig. 7.5) produced a maximum current velocity in the ebb direction and a minimum in the flood direction. The resulting ebb current must have been above a critical value, needed for sediment transport (U_{cr}), while the resulting flood current should have been below that value (Fig. 7.5):

(6) resultant current during the ebb: $U_e+U_r>U_{cr}$,

(7) resultant current during the flood: $U_f-U_r<U_{cr}$,

where U_e, U_f, U_r are the amplitude of the "basic" ebb and flood currents and the residual currents, respectively. Plus and minus sign in front of U_r indicates that it acts in the same or in the opposite direction with regards to flood or ebb currents.

On the western, Budafok side the compound internal structures of dunes indicate that the resultant flood current was stronger than the resultant ebb current. The resultant flood current transported sand to the south, and the ebb current produced gentle erosion and reactivation surfaces. Transport of sediment by the subordinate (ebb) current therefore must have been relatively small. Nevertheless both resulting currents must have been above a critical value (U_{cr}):

(8) resultant current during the ebb: $U_e-U_r\geq U_{cr}$,

(9) resultant current during the flood: $U_f+U_r>U_{cr}$,

(10) and also $U_e-U_r<U_f+U_r$

The solution of these unequalities for the basic flood and ebb velocities, which evolved in the generation of vorticity at both sides of the bay, is the following:

(6) - (10) $U_{cr}-U_r<U_f<U_{cr}+U_r\leq U_e$.

It is obvious from the above unequalities that the amplitude of the "basic" flood currents should have been smaller than that of the ebb currents ($U_f<U_e$). Their difference can be estimated from (10), as

$$U_e<U_f+2*U_r.$$

Since residual currents are relatively small (usually less than 10% of the parent current; Zimmerman, 1981), this difference is also small. Considering peak tidal current velocities in the range of 0.75-1 m/s, the difference between the velocities of the ebb and flood currents (i.e., $2*U_r$) must have been in the order of 0.15-0.20 m/s. Altough this difference seems to be small, there are recent examples when even smaller seasonal differences in current speeds and directions cause significant change in direction of net sediment transport (Harris, 1991). Therefore this small difference in combination with all other above described factors are assumed to have produced the asymmetrical sediment transport at both sides of the bay.

A velocity asymmetry of the tides also implies time asymmetry: long duration and low velocity flood currents and shorter duration but higher velocity ebb currents in combination with residual currents acted on the sediments at both sides of the North Hungarian Bay (Fig. 7.5). This is in contrast with the most common observation in shallow areas (Pugh, 1987), where the flood is

of shorter duration and of higher magnitude than the ebb. Such an asymmetry, which was present in the North Hungarian Bay can be well explained e.g. supposing the presence and superposition of out of phase higher harmonics (e.g. M4 tides, Lisitzin, 1974; Pugh, 1987).

It is thus concluded that the main factor creating the circular pattern of water movements in the North Hungarian Bay is an anticlockwise residual current interacting with time- and velocity-asymmetrical tidal currents both governed by the asymmetric bottom morphology. These currents could have been effective in sediment transport only because resonant amplification related to an approximately ideal length/depth ratio of the basin occured.

Analogues in the sedimentary record: discussion

Interpretations of ancient tide-influenced sediments, even if they were deposited along coasts of ocean basins or major seas, are not straightforward. The tidal regime depends on too many factors - e.g. along the modern Atlantic ocean the full range of micro- to macrotidal coasts is found, depending on the width and depth of the shelf, and the shape of the coastline among many other factors (Pugh, 1987). In semi-enclosed seas, not necessarily directly connected with oceans, the conditions for the development of tide-influenced sediments are even more puzzling. Studies of such small basins commonly lack detailed data about features such as basin depth and morphology which are available for modern systems. Nevertheless small, semi-enclosed fossil basins with signs of strong tidal currents may provide information about dimensions and palaeogeography. E.g. reconstructions of the palaeotidal regime were ventured in the shallow Western Interior Seaway, based on numerous examples of Jurassic-Cretaceous offshore tidal sand ridges (Bridges, 1982 and references therein). This seaway was large enough (4000 km long, 500-1000 km wide) to accommodate several amphidromic systems (Bridges, 1982). In the above example, the North Hungarian Bay was not wide enough to allow the development of an amphidromic system, but basin dimensions did allow resonant amplification of the tidal wave. The recognition of such phenomenon can be useful for reconstructions of other ancient seas.

Marine basins, particularly in tectonically active areas, commonly show an evolution in which basin dimensions (length, width and depth) change gradually. Thus, if some basic requirements with respect to basin dimensions and and connections to the open ocean are fulfilled, dimensions of such basins may, somewhere during their evolution, pass through a stage, in which amplification of tidal motions may occur. We shall call this stage the "tidal amplification window".

An obvious example is the Peri-Alpine Molasse Basin, where tidal currents have been very active during the early Miocene. As a result, a series of tide-influenced deposits can be followed from the Rhodano-Provancal gulf up to the coasts of the Bohemian Massive (Fig. 7.6; Homewood & Allen, 1981; Allen et al., 1985; Faupl & Roetzel, 1987, 1990; Tessier & Gigot, 1989;

Krenmayer, 1991; Lesueur et al., 1990; Allen & Bass, 1993). Based on observations of wave ripples Homewood & Allen (1981) calculated a basin width between 30 and 100 km and fairly shallow water up to a few tens of metres. This width was obviously too small to permit the development of an amphidromic system anywhere in the long peri-Alpine foredeep (cf. Fig. 7.1). The length of the basin, however, increased gradually during the "Burdigalian" transgression, and must have fit from time to time to the requirements for resonant amplification of the tide. Supposing that the average depositional depth remained about 30-40 m, and the basin length was near to odd multiples of the quarter tidal wave-length (L/4, 3*L/4, 5*L/4... Fig. 7.6; Pugh, 1987). E.g. offshore tidal sand waves around Zürich (Allen et al., 1985), are found at a distance of about 3*L/4 (600 km) from the Mediterranean (Fig. 7.2). In the vicinity of Linz, off the Bohemian coast, shallow water tide-influenced deposits are known (Faupl & Roetzel, 1987, 1990; Krenmayer, 1991). Resonant amplification must have developed due to reflection at the coast of the Bohemian massive, which was situated aproximately at a distance of 5*L/4 (1000 km) from the entrance of the Peri-Alpine Molasse Basin (Fig. 7.2 and 7.6).

Other examples of relatively small, semi-enclosed basins in which resonant amplification is likely to have occurred during basin evolution include the Eocene Catalan Basin (Santisteban & Taberner, 1988), the Aptian-Albian Lower Greensand (Allen, 1981; cf. Bridges, 1982), the Eocene Vlierze and Brussel Sand in the "Laon Strait" connecting the Paris Basin to the North Sea through Belgium (Houthuys & Gullentops, 1988a,b;) and the Upper Creataceous Ajali Sandstone in the Anambra basin in SE Nigeria (Lapido, 1988). This latter basin was an approximately 90 km wide, 200 km long basin opening to the long and narrow Palaeo-South-Atlantic (Frster, 1978; Reyment & Dingle, 1987).

An interesting case is also the Tremp-Ager Basin in the Southern Pyrenees, in Spain. During the Palaeocene, conditions for the amplification of tidal motions seem to have been fulfilled. This is indicated by the westwardly tidal reworking of fan-delta deposits (Roda Sandstone, Yang & Nio, 1985; Nijman, 1989) along the northern margin of the basin and tidal deltas with eastward transport along the southern margin (Ager Sandstone, Mutti, 1985; Nijman, 1989), strikingly similar to the case in the North Hungarian Bay. The width of the Tremp-Ager basin was in the order of 30 km and length is about 40 km with a more or less open connection towards the west, through the Ainsa and Jaca basins with the bay of Biscay (total length about 250 km; Nijman, 1989). Depth in the Tremp-Ager basin was some few tens of metres, increasing to below storm wave base in the Ainsa basin (Baas, 1993) and further towards the ocean. Thus the depth and length needed for the amplification of the tidal wave seem to have been provided. Intervals in the overlying Eocene deposits also reflect the effect of relatively strong tidal currents (Nijman & Nio, 1975), indicating that dimensions of the combined Tremp-Ainsa and Jaca basins passed the "tidal amplification window" several times.

It thus appears that the recognition and the analysis of tidal amplification in ancient sedimentary basins offers a valuable tool for the reconstruction of basin dimensions and for

identifying periods during which the basin dimensions, width, length and average depth, passed through the "tidal amplification window".

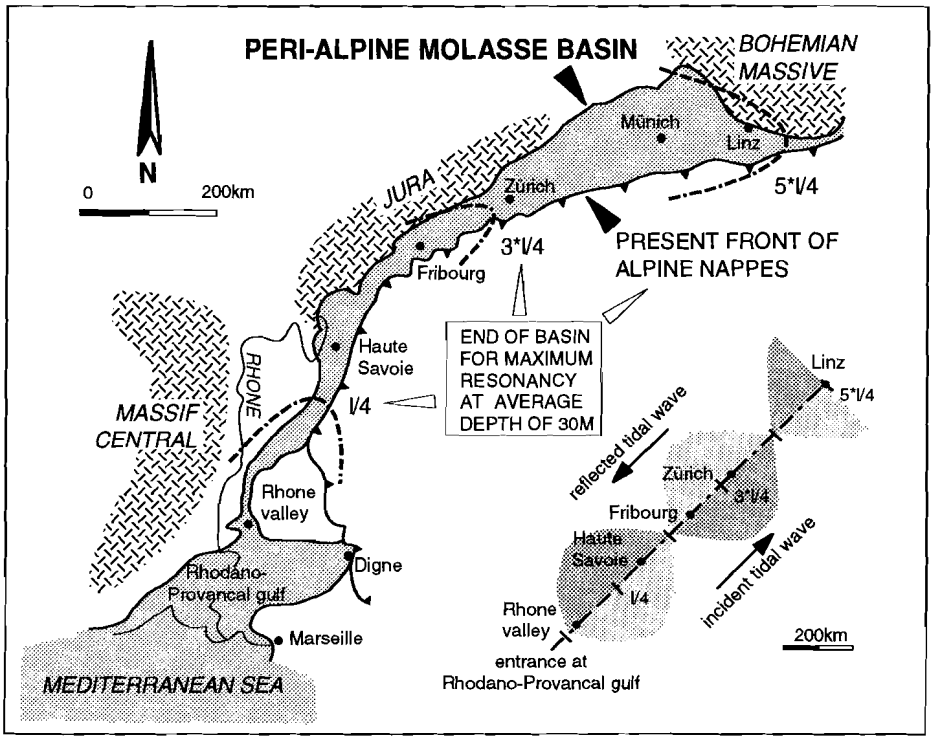

Fig. 7.6. Tide-influenced deposits in the long and narrow Peri-Alpine depression (see references in the text). It seems that the basin remained in the "tidal amplification window" during the Burdigalian transgression.

Conclusions

The sedimentary structures in the North Hungarian Bay reflect the activity of strong tidal currents, which were generated by the local amplification of tidal motions. The elongate, funnel-shaped bay with a wide mouth and a progressively decreasing width supported the amplification of tidal motions. Amplification could occure, because the depth and length of the North Hungarian Bay together with the connecting relatively shallow seaway through East Slovakia fulfilled the conditions for near-resonance with the semidiurnal tidal wave. Tidal motions cannot have been amplified exclusively in the North Hungarian Bay; the East Slovakian Seaway must have played an important role in the transmission and amplification of the tidal motions.

Sediments in the North Hungarian Bay reveal a dominance of ebb currents along the

eastern coast and a dominance of flood currents along the western coast of the bay. The tidal wave passing slightly obliquely over the central depression of the bay induced a cyclonic vorticity, resulting in small anticlockwise residual currents. These residual currents in combination with time asymmetric tidal currents, which both are determined by the bottom morphology, created the circular pattern of water movements in the North Hungarian Bay.

The case of the North Hungarian Bay and the shortly discussed analogues settings demonstrated that a hydrodynamic-oceanographic analysis of the tide-influenced deposits and their environment can provide information about the dimensions of marine basin.

CHAPTER 8
EARLY MIOCENE BASIN EVOLUTION IN NORTHERN HUNGARY: TECTONICS AND EUSTASY

Tectonophysics 226/1-4: 485-502

ABSTRACT

It is presumed that the Oligocene–early Miocene basin evolution in northern Hungary was primarily driven by compressional tectonics, producing a major second-order transgressive–regressive facies cycle. The early Miocene basin evolution is best understood in terms of "molasse" sedimentation in an overfilled flexural basin. During this time the gradual cessation of thrusting in the adjacent West Carpathian thrust-fold belt resulted in its uplift and subaerial exposure. Significant amounts of sediment were delivered to the flexural basin filling it up to sea level.

During the late-stage uplift of the flexural basin a shallow-marine depositional environment developed, as a result of isostatic rebound, and signals of third-order eustatic sea-level changes can be revealed. Sedimentological studies of outcrops of the Lower Miocene succession proved marked changes in facies. Along the gently dipping distal (southeastern) flank of the basin a sudden inception of shallow-marine coarse clastics on top of siltstones, deposited in significantly deeper water, may have been the result of a third-order eustatic sea-level fall shortly before the Burdigalian (at the boundary of the NN1/NN2 nannofossil zones). During the resulting lowstand, various tide-dominated facies aggraded, which were subsequently flooded by the "Burdigalian sea-level rise". The highstand is represented by upwards shallowing progradational units due to accelerating sedimentation and/or tectonic uplift.

The typically elongated and narrow flexural basin, characterized by small water depth, had dimensions which were particularly suitable for the amplification of tidal motions. This resulted in the deposition of tide-influenced sandy sediments. In contrast to examples where the evolution of a strong tidal influence is related to transgressions, here such conditions developed following a drop of sea level.

Introduction

The tectono-stratigraphic evolution of flexural basins commonly follows a two-phase scheme. An early ("flysch" or underfilled or deep-water) stage is followed by a late ("molasse" or overfilled or shallow-water) stage (Covey, 1986; P. Allen et al., 1991). The early stage is characterized by accelerating subsidence due to the load of an adjacent thrust-fold belt, resulting in an elongated and asymmetrical marine basin. During the late stage the rate of thrust propagation slows down, and isostatic uplift of the thrust belt and the adjacent flexural basin occur. The emergent wedge sheds large amounts of sediment into the basin filling it up to or above sea level. The late "molasse" stage infills of flexural basins are usually characterized by fluvial and/or shallow-marine depositional systems (P. Allen et al., 1986). Periods of widespread shallow-marine sedimentation are quite often represented by tidal deposits.

Tide-influenced deposits, however, have been mostly reported from transgressive settings (J. Allen, 1981; Homewood and Allen, 1981; Stride, 1982; Belderson, 1986; Faupl and Roetzel, 1987; cf. De Boer et al., 1988; cf. Smith et al., 1991; cf. Flemming, 1992). They often overlay fluvial or estuarine deposits and therefore are considered to be especially representative for transgressions. When incised valleys and subaerially exposed areas are flooded, the "new-born" sea is shallow with respect to the approaching tidal wave, which therefore increases in amplitude (Lisitzin, 1974). Tidal motions are intensified, resulting in a tide-

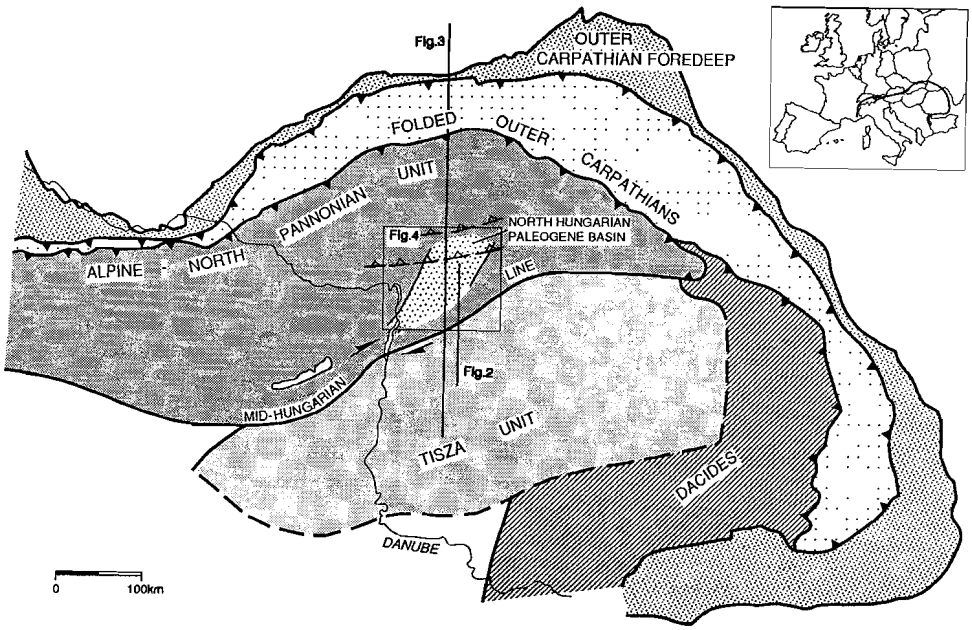

Fig. 1. Main tectonic units of the intra-Carpathian area and the location of the Palaeogene Basin in northern Hungary (after Csontos et al., 1992). The existence of SE-vergent thrusts in the basement at the northern margin of the basin is hypotethical. Position of sections of Figs. 2 and 3 are shown, as well as cut of detailed map (Fig. 4).

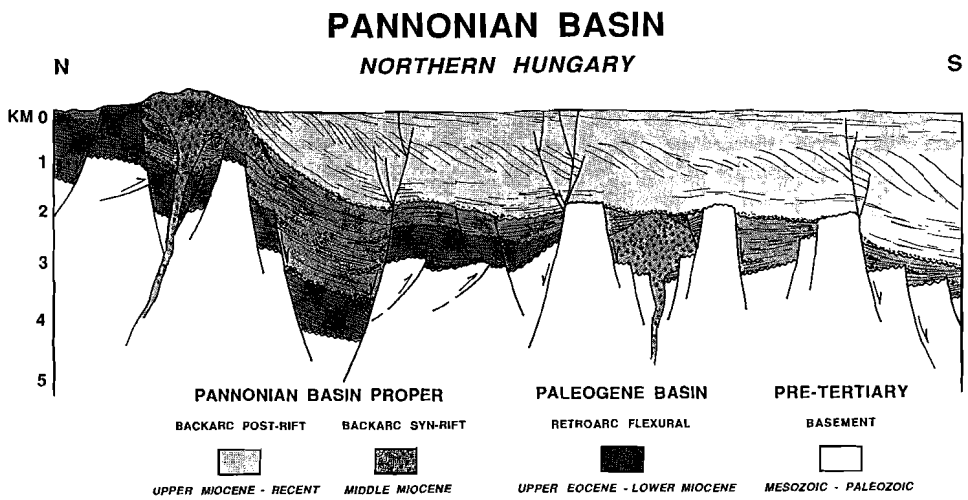

Fig. 2. Schematic section across northern Hungary showing the major Tertiary tectono-stratigraphic units of the Pannonian Basin. The vertical exaggeration is about tenfold. For a detailed explanation see text.

influenced sedimentary facies. The rising sea allows the fossilization of the tide-influenced deposits thus formed below open marine sediments.

Another effective way to enhance tidal forces by means of resonant amplification (Lisitzin, 1974) is provided in narrow seaways, such as the peri-Alpine foreland basin (Homewood and Allen, 1981; P. Allen et al., 1985) or in partly enclosed embayments (Amos, 1978; Rubin and McCulloch, 1980; Ramos and Galloway, 1990). These examples were also evolved in connection with marine flooding.

In this paper a sedimentological approach is used in order to arrive at a detailed structural reconstruction of basin evolution in northern Hungary. The evolution of the shallow-marine sedimentary architecture and the presumed effects of the late stage uplift of the flexural basin are discussed. The second goal of this study is to document the development of a tide-influenced depositional system in this basin in response to a marked sea-level fall. This scenario is, as yet, unique between the numerous examples of transgressive tidal systems.

Geological setting

Tectonic setting

The Neogene Pannonian basin proper is one of the Mediterranean back-arc basins (F. Horváth and Royden, 1981). It was formed as a typical transtensional basin where extension was coeval with compression in the surrounding Carpathian thrust-fold belt (Fig. 1; for a summary see Royden and Horváth, 1988). The two-stage fill of the Neogene Pannonian basin proper clearly reflects its extensional origin, as shown on a schematic cross-section through northern Hungary (Fig. 2; cf. also fig. 3 of Tari et al., 1993-this issue) The uppermost unit is the post-rift sedimentary succession developed from Late Miocene to Recent (10.5–0 Ma) and which is the result of regional thermal subsidence (F. Horváth and Royden, 1981). These strata are frequently cut by strike-slip related flower structures indicating that the inversion of the basin has already began. The locally pronounced uplifts (see e.g., the northern part of the section in Fig. 2) are thought to be the result of Quaternary tectonics. The underlying Middle Miocene (17.5–10.5 Ma) succession was formed in fault-bounded half-grabens and pull-apart basins (Fig. 2), reflecting syn-rift structural features. During the syn-rift stage widespread Miocene calc-alcaline volcanism (Fig. 2) was generated by the subducting European plate beneath the Carpathian orogenic belt (Szabó et al., 1992).

Beneath the well-known Neogene Pannonian basin proper, however, a Middle Eocene–early Miocene basin is found in northern Hungary (Fig. 2), which is traditionally called "Palaeogene Basin" (e.g., Csiky, 1961). In this paper we shall

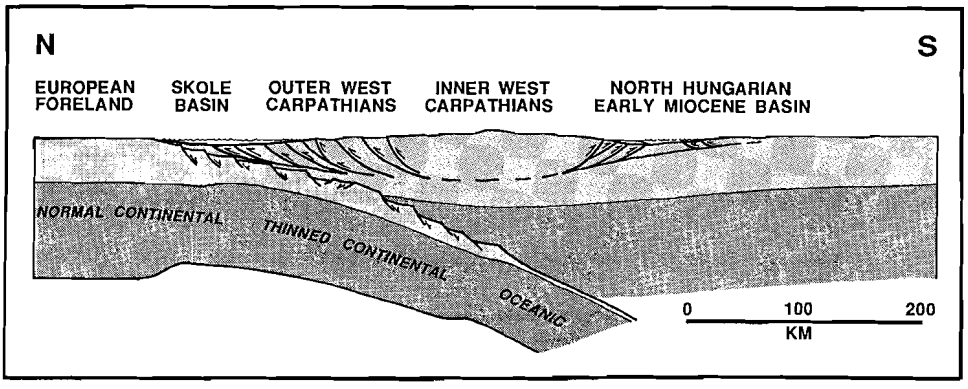

Fig. 3. Cartoon showing the early Miocene tectonic position of the Palaeogene Basin of northern Hungary in relation to the Western Carpathians (Tari et al., 1993).

use terms of "Palaeogene Basin" for the whole and "early Miocene basin" for the youngest part of this basin succession. The formation of the Palaeogene Basin of northern Hungary is much less understood than the overlying Neogene Pannonian basin. While some authors (e.g., Báldi and Báldi-Beke, 1985; Royden and Báldi, 1988) inferred a transtensional (pull-apart) origin for the Palaeogene sedimentary succession, it was recently reinterpreted as a flexural basin (Fig. 3; Tari, 1992; Tari et al., 1993-this issue).

Present-day boundaries of the Early Miocene basin

The sharp southern boundary of the basin (Figs. 1 and 2) is due to dextral strike-slip faulting along the Balaton–Mid-Hungarian wrench zone (Csontos et al., 1992), which occurred at the end of the Early Miocene (Tari, 1992; Tari et al., 1993-this issue). The large-scale right-lateral movement along this fault zone associated with the escape of the northwestern part of the Pannonian Basin from the Alpine realm (Csontos et

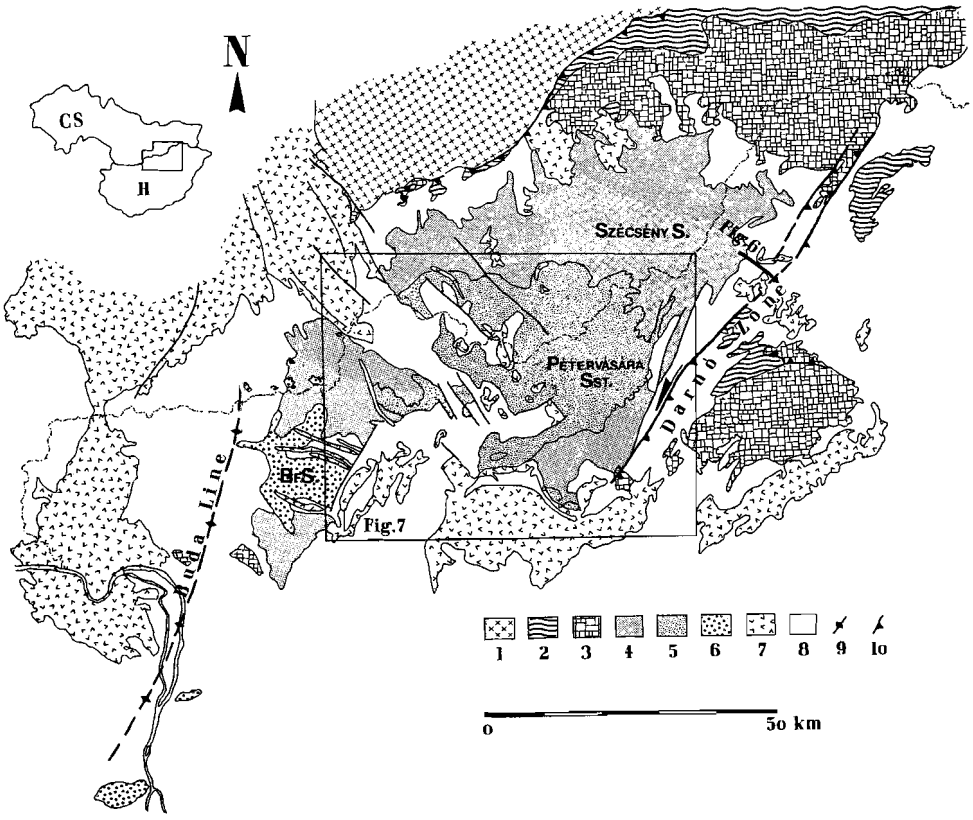

Fig. 4. Simplified geological map of the area showing present surface distribution of the studied formations (without Quaternary). Paleogene basin fill to the south is covered by sediments of the Pannonian basin proper. *1* = high-grade metamorphics, granitoids in Vepor; *2* = undifferentiated Palaeozoic rocks in Gemer, Szendrő, Uppony and Bükk; *3* = undifferentiated Mesozoic rocks in Gemer and Bükk; *4* = Egerian to Eggenburgian Szécsény Schlier; *5* = Eggenburgian Pétervására Sandstone; *6* = Eggenburgian Budafok Sandstone; *7* = various Miocene volcanics; *8* = undifferentiated post-Eggenburgian sediments; *9* = anticline; *10* = reverse faults or thrust fronts at the surface.

al., 1992) is primarily responsible for the present-day collage of disrupted Palaeogene basin fragments in the intra-Carpathian area (Nagymarosy, 1990).

The Darnó Zone, which consists of a bunch of related faults, is considered to be the eastern boundary (Fig. 4) of the Early Miocene basin. Recently acquired high-frequency reflection seismic data (Braun et al., 1989) shows a truncation of Upper Oligocene strata within the Darnó Zone indicating regional compression around the Oligocene/Miocene boundary (Fig. 6). Deposition of the Lower Miocene postdates the activity of reverse faults. However, the small anticlines (Fig. 6) indicate that compression continued with much less intensity during the Early Miocene. At the southeastern end of the seismic section the Darnó Line can be seen crosscutting the Lower Miocene sedimentary succession (Fig. 6). While the Darnó Zone was characterized by right-lateral displacements during the Palaeogene (Balla, 1987), it was reactivated in a left-lateral transpressional sense folding the whole Lower Miocene basin fill during Late Miocene times (Zelenka et al., 1983).

The western boundary of Lower Miocene succession approximately coincides with the NNE-trending Buda Line defined by Báldi and Nagymarosy (1976, see fig. 4). The Buda Line actually represents a broad zone of SE-vergent anticlines, which were active during Late Eocene and Early Oligocene times (Fodor et al., 1992). During the Late Oligocene, thrusting ceased in this region and the whole area suffered differential uplift as a result of isostatic rebound (Tari et al., 1993-this issue). The Late Eocene thrust front, however, was inherited as a topographic boundary during the early Miocene period, separating the subaerially exposed interior part of the thrust-fold belt west of the Buda Line from the shallow-marine basin to the east.

The northern boundary of the early Miocene basin is not entirely tectonic (Vass et al., 1979, 1988). Although the hypothetical thrust fronts (Figs. 1 and 3; Tari et al., 1993-this issue; and discussions in this paper), probably determined the position of the shoreline. Based on the character of the sedimentary sequence, however, an open marine outlet has to be assumed to the northeast towards the East Slovakian Basin.

Stratigraphy

The Pannonian Basin evolved in the Paratethys region which is characterized by its own faunal and tectonic evolution. Since the standard stages of the late Palaeogene and the Neogene could not be used, regional stages were defined starting

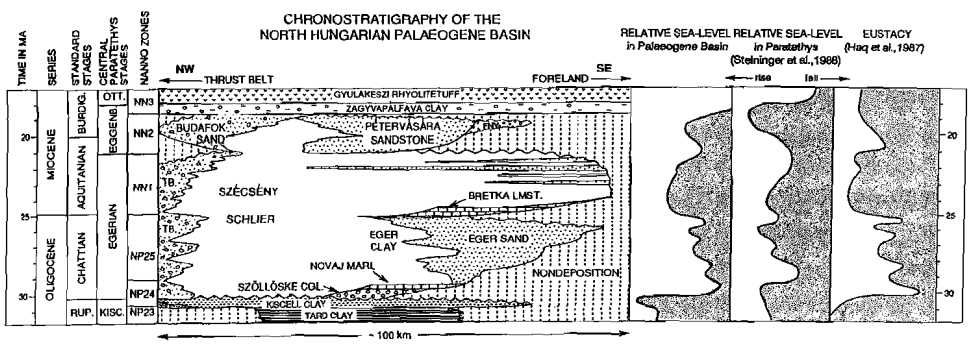

Fig. 5. Chronostratigraphy of the Palaeogene Basin in northern Hungary from Oligocene to early Miocene. Compare changes of relative sea level in Palaeogene Basin with those of Paratethys and the global chart. Depositional environments and facies of Eggenburgian are summarized in text. *TB* and *FNY* refers to Törökbálint Sand and Felsőnyárád Formation, respectively. Some of these deposits are also known from southern Slovakia under different names: e.g., Lucenec or Szécsény Schlier, Fil'akovo or Pétervására Sandstone and Lipovany or Budafok Sand (Vass et al., 1988).

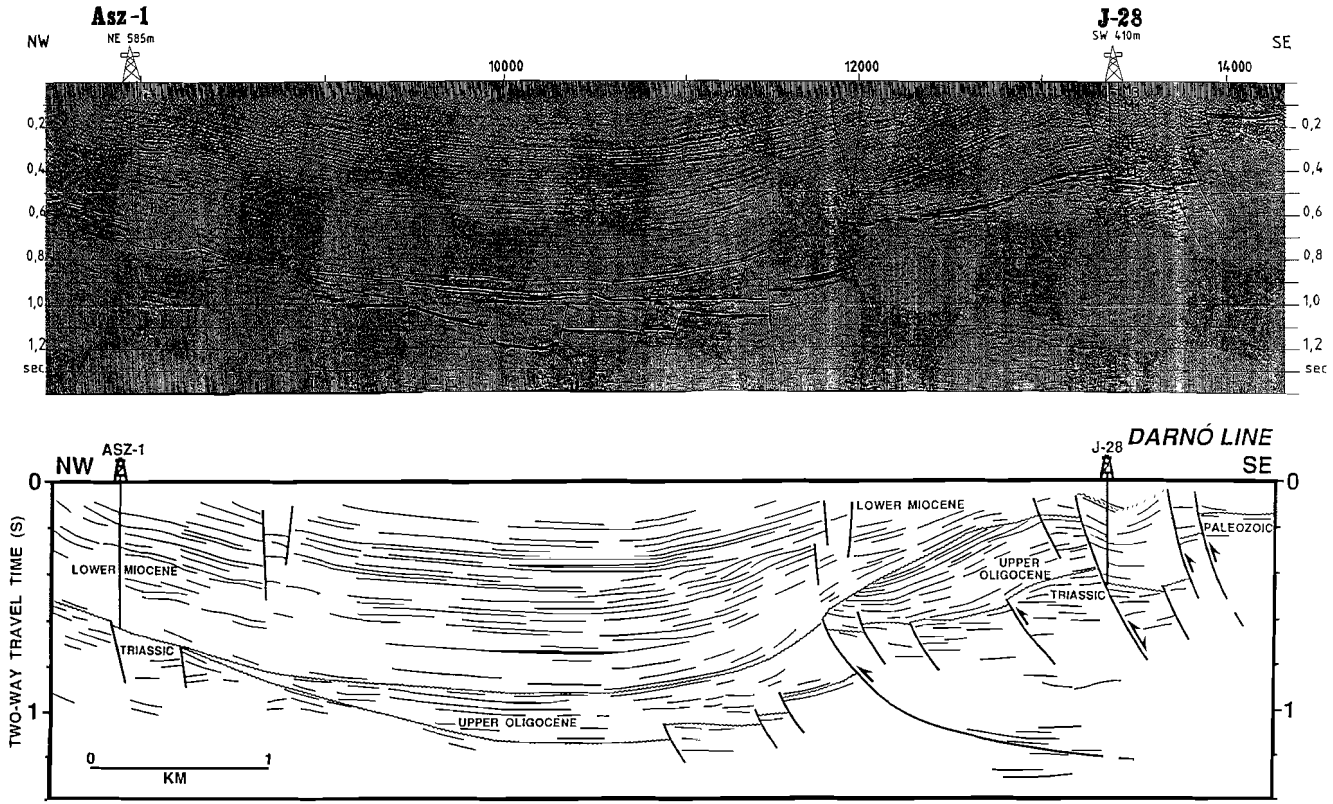

Fig. 6. Part of seismic section through the northeastern part of the basin (for location see Fig. 4) shows the transpressive character of Darnó Zone. For discussion see text. (top) Section published by Braun et al. (1987). (bottom) Recent interpretation.

from the base of the Oligocene. For a summary and correlation the reader is referred to Steininger and Senes (1971), Báldi and Senes (1975) Báldi (1980) and Rögl and Steininger (1984). A comprehensive study of the Palaeogene bio- and lithostratigraphy of Hungary is given by Báldi (1986). The distribution and correlation of formations in the Carpathian region was reviewed by Nagymarosy (1990).

The stratigraphy of the Palaeogene Basin from late Early Oligocene to early Miocene is shown in Figure 5 (modified from Tari and Sztanó, 1992). The calcareous nannoplankton biozonation served as the main tool for chronostratigraphic correlation (Nagymarosy and Báldi-Beke, 1988). Additionally, we adopted the absolute time scale of Haq et al. (1987).

The basin fill consist of a second-order transgressive–regressive facies cycle (Tari and Sztanó, 1992) from the Kiscell Clay to the Zagyvapálfalva Clay (Fig. 5). In this study we focus only on the effects of the final regressive processes during Eggenburgian times. Characteristics (thickness, bio- and lithofacies, depositional environments, etc.) of the Eggenburgian formations are summarized below.

The Szécsény Siltstone ("Schlier")

The 400–800-m-thick Szécsény Siltstone is Egerian to Eggenburgian in age (Fig. 5). Except for the deepest parts of the basin (Figs. 4 and 5), where Eggenburgian foraminifera (M. Horváth, 1972) and nannofossils (NN2) were reported (Nagymarosy and Báldi-Beke, 1988), deposition of Szécsény Schlier was confined to the Egerian. During the Egerian and Eggenburgian, the Schlier represents the deepest depositional environment in this area. It consists of strongly bioturbated sandy clayey silt, rich in thin-shelled molluscs (e.g., *Amussium* sp.). Palaeoecological studies (forams—M. Horváth, 1972; molluscs—Báldi, 1973) indicated a shallow bathyal to deep neritic environment for the Schlier, approximately at 60–300 m water depth. A decrease of depositional depth was observed in late Egerian times. A sandy atypical lithofacies of the schlier, still with typical biofacies was described as a transitional unit towards the adjacent and overlying Pétervására Sandstone (Báldi, 1986).

The Pétervására Sandstone

The Eggenburgian shallow-marine Pétervására Sandstone overlies the Schlier in the eastern part of the basin (Figs. 4 and 5). It crops out over 1500 km^2 in northern Hungary and southern Slovakia (Fig. 4). Thickness of the formation increases from southeast towards northwest from 150 to 600 m. It consists mainly of medium- to coarse-grained lithic sandstone and conglomerates (Vass et al., 1988) with intrabasinally reworked glauconites. The rock fragments have been derived from the crystalline basement, dominantly present in the east. Reworking of the sandy components from older sands, situated southward, is also supposed. The Pétervására Formation is poor in fossils, except for the conglomerates, which contain high amounts of mollusc debris. The *Chlamys–Ostrea–Balanus* assemblage encountered indicates high-energy conditions nearby (Fözy and Leél-Össy, 1985; Báldi, 1986). Based on palaeogeographical considerations even a rocky shore along the Darnó line might be inferred (Fig. 4). Detailed outcrop studies revealed a tide-influenced sedimentary environment (see below).

The Budafok Sand

The Eggenburgian Budafok Sand is restricted to a narrow patch along the Buda Line. It is an ill-sorted clean quartz arenit with conglomerate intercalations. Various sedimentary structures from ripples to large dunes and the *Chlamys–Anomia–Ancilla–Oliva* molluscs assemblage indicate a shallow neritic water depth (2–30 m) and littoral to nearshore deposition (Báldi, 1973). In contrast with the eastern part of the basin here strong tidal influence was not recognised. Transport partly happened by longshore currents from the north (Báldi, 1986). The bulk of the material comes from granitoid rocks (Vepor Unit, Fig. 4).

The Felsőnyárád Formation

Within and east from the Darnó Zone there is a different Eggenburgian section, which cannot directly be compared with the basin fill succes-

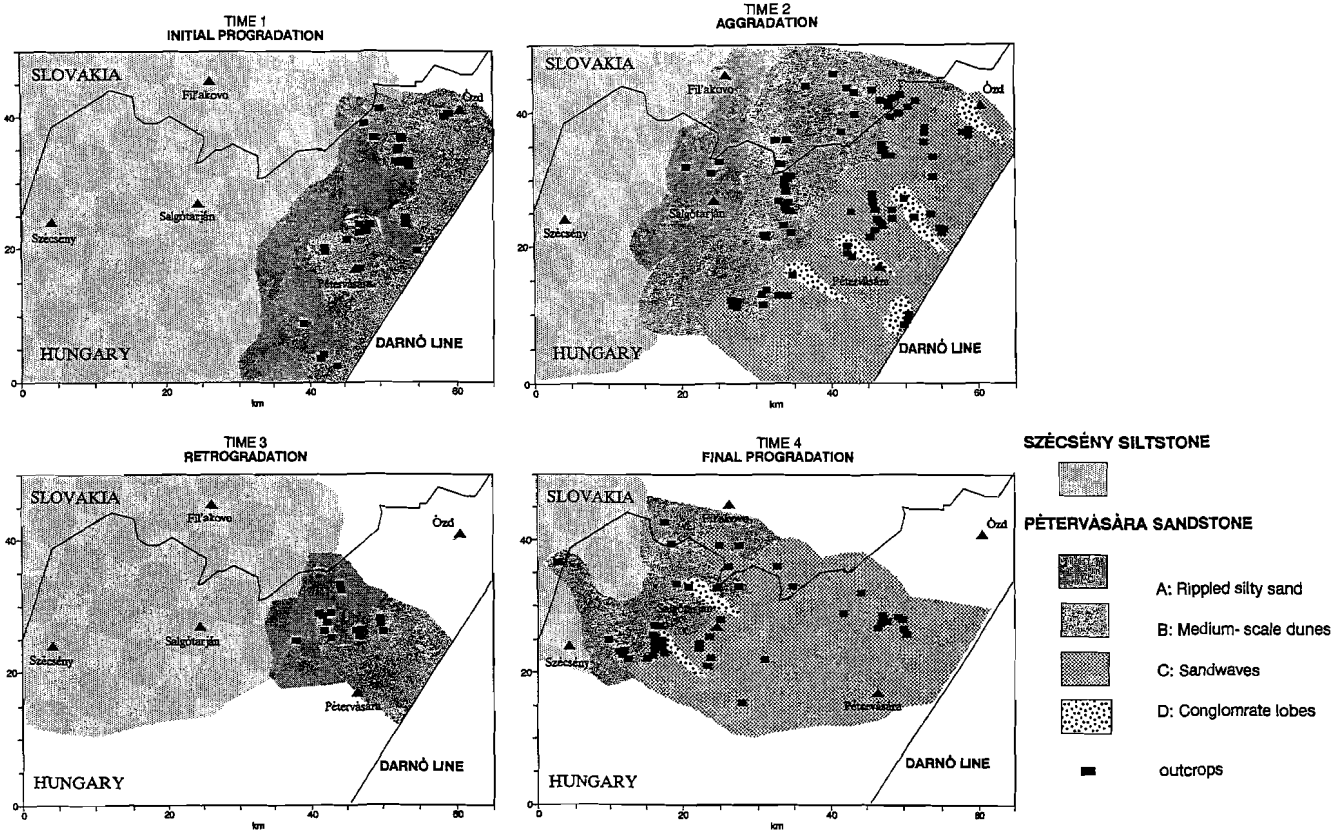

Fig. 7. Maps of facies distribution of the Pétervására Sandstone. Time 1 –facies units A and B appeared shortly after the most intensive relative sea-level fall, during initial progradation; Time 2—shore parallel distribution of facies units *C*, *B* and *A* during aggradation; Time 3—landward stepping of facies units *A* and *B* during a minor flooding. The significant landward shift is obvious in comparison with the previous time slices; Time 4—facies units *B*, *C* and *D* advanced far to the west, towards the center of the basin, during final progradation.

sion to the west. Although the Pétervására and Felsönyárád Formations does not interfinger at present, they must have been closely related environmental belts during deposition. Felsönyárád beds, limnic coal seams grading upwards into littoral to brackish-marine sands and muds, of maximum 300 m thickness are transgressive over basement rocks (Radócz, 1964). The top of the formation is erosive.

The Zagyvapálfalva Clay

Before the end of the Eggenburgian the basin was filled to sea level, and the terrestrial Zagyvapálfalva Clay formed. It consists of variegated mud, ill-sorted sand and conglomerates. The material was derived from a metamorphic and granitoid source area, that is inferred to be located in the north. The relatively thin sequence (maximum 60 m) was interpreted in terms of coastal plain, fluvial and flood plain environments (Hámor, 1985). In the flood plain deposits a rich assemblage of bird and mammal footprints were exceptionally well preserved (Kordos, 1985; Bartkó, 1985) below thick rhyolite tuff deposits at Ipolytarnóc.

Sedimentary facies units of the Pétervására Sandstone

The stratigraphic architecture in terms of facies units was reconstructed on the basis of correlation of field sections (Fig. 8) and two representative well logs (Fig. 9). Remarkable lateral shifts of facies units, deposited in shore-parallel zones (Fig. 7), reflect the changes in accommodation (Jervey, 1988; Cross, 1988) as the result of the interaction of synsedimentary tectonics and eustatic sea-level changes. The location of facies units during subsequent steps of evolution of the basin are shown in maps of Figure 7.

Facies unit A—thin-bedded, rippled silty sand

In this facies bedding is uneven, undulatory and flaser-like. Ripples in very fine sand were occasionally observed. Bioturbation is fairly intensive. Occasionally decimetre-scale glauconite-rich foresets consisting of medium-grained sand are intercalated, indicating the intermittent increase of current activity.

Although the lithology compares relatively well to the atypical type Szécsény Schlier, this unit lacks of molluscs characterizing the schlier. The rippled silty sand is a transitional facies unit between the Szécsény Schlier and the Pétervására Sandstone. The abundance of current-induced sedimentary structures and a higher amount of coarser grains points to a higher-energy environment for facies A, than for the adjacent Schlier. Also the presumed depositional depth of the rippled silty sand is smaller than that in which the Schlier was formed (deeper than 60 m, Báldi, 1986).

Facies unit B—thin-bedded sandstone with medium-scale crossbedding

In large areas (Fig. 7) the Pétervására Sandstone is fine- to medium-grained. It is structureless with cemented nodules and horizons or is associated with mud drapes and medium-scale crossbedded sets with heights of 0.2–0.6 m. The facies unit is structureless because it has been almost entirely bioturbated. Traces of crossbedding, however, have been preserved below cemented horizons, which indicate periods of non-deposition. Early diagenetic cementation, however, was not significant (Molenaar, pers. commun., 1991). Crossbedding is often shown by an alternation of glauconite-rich and glauconite-poor laminae. Tiny trace fossils are abundant too. Double mud drapes (cf. Visser, 1980) on foresets, and bottomsets with ripple marks are common. Palaeotransport directions are dominantly towards the north, but a few examples of structures produced by counter currents to the south were also measured.

Double mud drapes record slack water periods of successive ebb and flood events, and thus record deposition in a subtidal regime (Visser, 1980). The crossbeds of facies unit B represent a field of small- to medium-scale dunes, moved by tidal currents. A progressive decrease of depositional depth and an increase of current energy in the direction of the shore (SSE), suggests that facies units were deposited in shore-parallel belts.

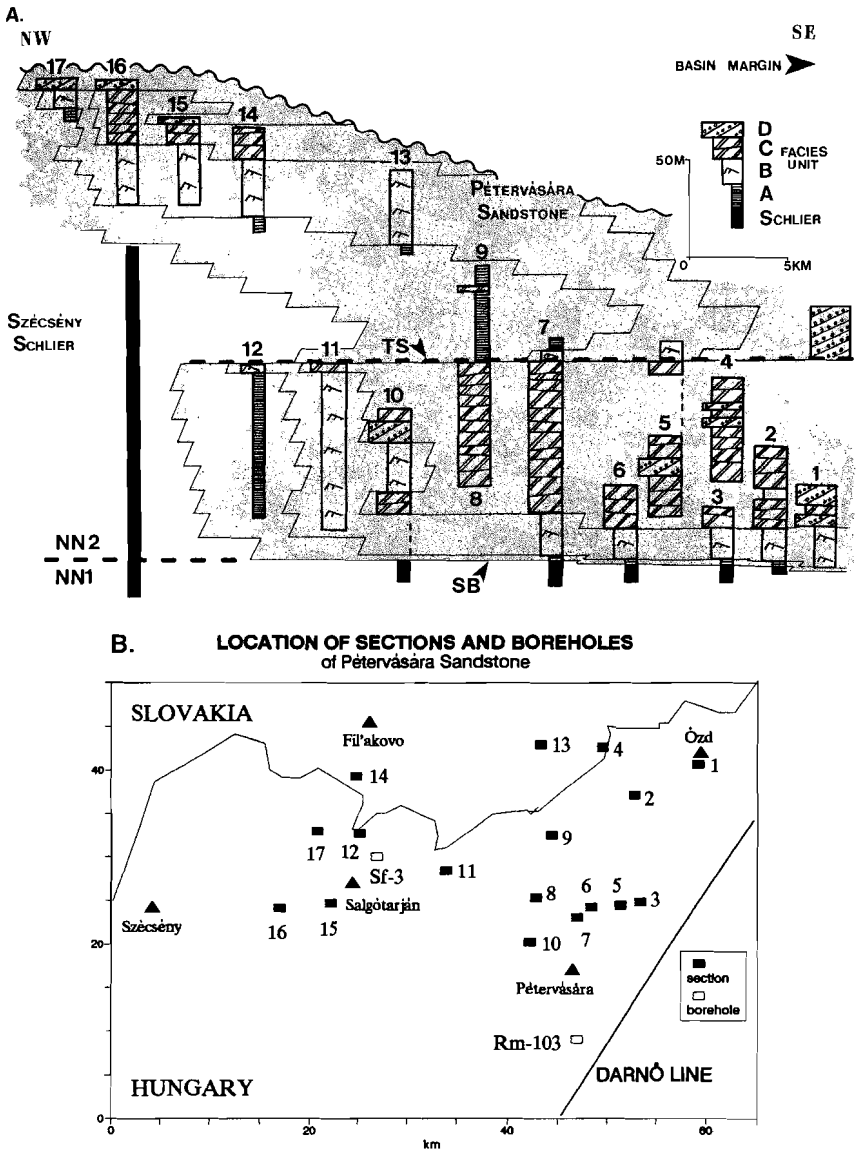

Fig. 8. (a) Composite profile reconstructed from field sections and borehole data of the Pétervására Sandstone shows the early phase of aggradation and the late phase of progradation of facies units. The transgressive surface (*TS*) was used as a datum. The border of NN1/NN2 nannozones coincides with the sequence boundary (*SB*) between the Szécsény Schlier and the Pétervására Sandstone. For explanation of facies units see text. (b) Map for location of sections and boreholes (Fig. 9). *1* = Ózd-Somsály hills; *2* = Borsodszentgyörgy west; *3* = Szentdomonkos, Kő hill; *4* = Drna south; *5* = Tarnalelesz, Peskó tető; *6* = Bükkszenterzsébet, Szappankő; *7* = Váraszó valley; *8* = Istenmezeje, Noah's vineyard; *9* = Zabar, road cut; *10* = Ivád, Nagy-Lyukaskő; *11* = Bárna, Szárkő; *12* = Somoskőújfalu railroad cut; *13* = Gemerske Dechtar; *14* = Cakanovce; *15* = Salgótarján, Csókás; *16* = Kishartyán, Kőlyukoldal; *17* = Karancslapujtő, apiary.

Facies unit C—medium- to coarse-grained sandstone with large-scale crossbedding

The dominant sedimentary structure in the Pétervására Sandstone is large-scale crossbedding consisting of relatively well sorted sandstone. Foreset morphology varies between tabular, tangential and sigmoidal. The average set thickness is 2–3 m, but locally it attains 8 m. Variations in shape, dip angle and grain size of foresets suggest

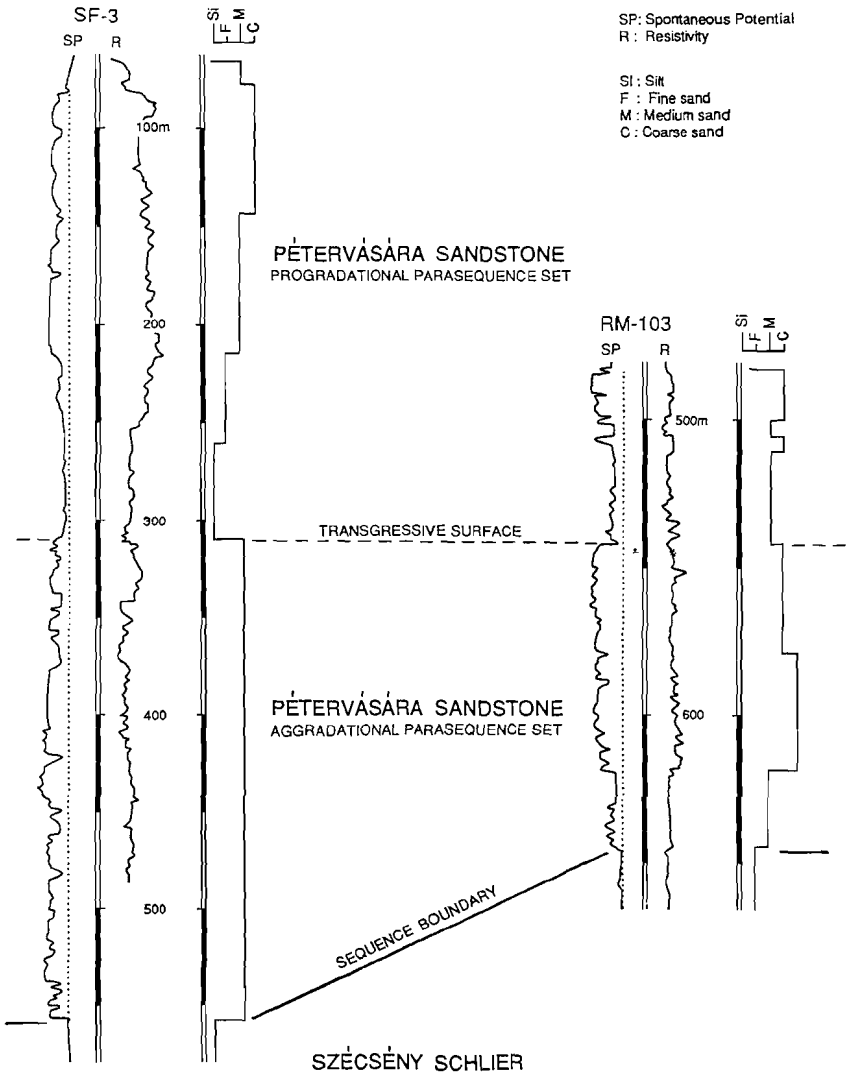

Fig. 9. Geophysical log and lithological column of two boreholes, cutting the top of Szécsény Schlier and Pétervására Sandstone. *RM-103* is near to basin margin, *SF-3* is from central part of the basin. For exact location see Fig. 8b.

periodical changes of flow energy. Mud drapes on foresets are scarce. The occasional occurrence of mud couplets, however, implies subtidal deposition (cf. Visser, 1980).

A fluctuation of grain size from very coarse to medium-grained sand is spectacular in large foreset laminae. A cyclic variation in foreset thickness was measured (Tari et al., 1989). These bundle sequences can be correlated with spring-neap tidal periodicity (Visser, 1980). Though many cycles are incomplete, semidiurnal tidal activity (two tidal cycles daily with diurnal inequality) is revealed by the presence of alternating thin and thick bundles (cf. De Boer et al., 1989). The foresets of dunes monotonously dip to the north, indicating a highly asymmetrical tidal influence (J. Allen, 1980). The palaeotransport direction was subparallel to the shoreline, which is supposed to have been along the Darnó Zone (Fig. 3).

Spectacular water escape structures in coarse sand might be the result of minor seismic shocks triggered by synsedimentary fault activity along the Darnó Zone.

The above described geometry of large-scale bedforms conforms well with the sandwave model of J. Allen (1980). Neither the sedimentary structures nor their distribution (Fig. 7) reflect channelized deposition, but rather a sheet-like morphology is inferred. Facies unit C is interpreted as a tidal sand sheet (cf. Stride, 1982; Belderson, 1986), consisting of northward migrating sandwaves, inward of the belt of medium-scale dunes. The dimensions of sandwaves might indicate a depositional depth around 15–30 m (J. Allen, 1982, figs. 8-20 and 11-25).

Facies unit D—graded conglomerates with large-scale crossbedding

Along the eastern margin of the basin, in the Darnó Zone, conglomerates are found. The pebbly material (radiolarite, quartzite, siliceous schists and mafic paleovolcanites) was derived from the basement in the southeast. To the west a decrease of grain size, and a reduction of less resistant components (e.g., dolerite) is observed. Conglomerates are associated with faults (Fig. 7). Steeply dipping strata of pebbly sand and sandy conglomerates are intercalated in a coarsening upward trend. The conglomeratic bodies are interpreted as small fan deltas relying on faults.

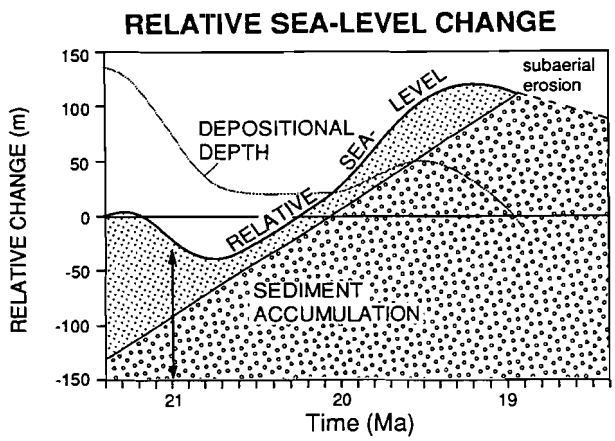

Fig. 10. Change of relative sea level was reconstructed from sediment accummulation (for simplicity a constant rate is supposed) and from changes in depositional depth (inferred from palaeoecologic indicators and sedimentary structures). Arrow indicates the position of the sequence boundary. Scale is approximative.

Their position and internal structure is a direct evidence of synsedimentary tectonic activity along the eastern margin of the basin (Fig. 4).

Also there are huge (8 m high) conglomeratic beds within the sandwave field (Fig. 7). Large-scale crossbedding and normal grading above the coarse bottomsets is characteristic. The height of the megaforesets approximates actual water depth. These conglomeratic units are interpreted as small, coarse-grained deltaic lobes, prograding to the west, and then modified by strong northward directed tidal currents.

Vertical and horizontal facies transitions

The transition from the Szécsény Schlier to the Pétervására Sandstone is gradual, but rapid. Below the base of the Pétervására Formation a few metres of atypical schlier is found. It is overlain by rippled silty sand of 2–10 m (facies unit A), followed by a thin (5–20 m) coarsening and thickening upward succession of medium-scale dunes (unit B), up to large-scale crossbedding of unit C. (Fig. 8, sections 1, 2, 3, 6 and 7). This rapid change from the Schlier to the tidal sandwaves reflects a seaward shift of facies units and a radical decrease in depositional depth (Fig. 10).

The thickness of facies unit B increases to the west, i.e., in the offshore direction, where it exceeds 100 m. It laterally interfingers with large-scale crossbedding of facies unit C to the east, and with facies unit A to the west (Figs. 7 and 8, section 11).

Higher upward in the stratigraphic record (Fig. 8 sections 7 and 9, Fig. 9), rippled silty sand (facies unit A) reappears and attains a thickness of 50–150 m. It covers a thick (100 m) series of large-scale crossbedded sets of unit C. This interval of "offshore" deposits indicates an increase of depositional depth.

There is a "younger" generation of small- to medium-scale dunes (facies unit B), towards the "center" of the basin. These overlay the Schlier Formation, which has an Eggenburgian age (NN2) there (Fig. 8, sections 14–17). On top of facies B sandwaves of unit C and conglomerates of D prograded reflecting a continuous decrease of accommodation.

Thickness of facies unit C is fairly variable throughout the study area. In outcrops 30–60-m-thick successions are abundant (Fig. 8, sections 1–9). Sandwaves of facies unit C are either abruptly overlain by rippled silty sands of unit A or by conglomerates of unit D. C/A transitions, however, are confined to the middle part of the Pétervására Fm. (Fig. 8, sections 8–9). Intercalation of conglomerates (D) is abundant in lower parts of the sandstone (Fig. 8, sections 1, 4, 5 and 10), while coarsening upward sequences of C and D are found close to top of the Pétervására Formation (Fig. 8, sections 15–17).

Speculations on structural evolution of the Palaeogene basin

During early stage of foreland basin evolution the load of the adjacent thrust-fold belt generates a rapidly subsiding, elongate marine basin. Because of the asymmetric geometry of the trough coarse clastics are trapped in front of the thrust front. On the foreland flank of the foredeep sediments might be shed from the uplifted peripheral bulge, but the volume of this input is insufficient with respect to the capacity of the basin. Therefore, the remarkable deepening is due to the "slow" response of sedimentation to subsidence (Heller et al., 1988). This early stage is traditionally called "flysch" (cf. P. Allen et al., 1986).

When the rate of thrust propagation slows down, basement-involved thrusting occurs typically in the rear side of the wedge. Thus another, flexural basin is formed, but in a retroarc position and it repeats the two-stage evolutionary scheme. When the activity of the thrust wedge ceases, the resulting isostatic uplift causes subaerial exposure. Thus, significant amounts of sediment are delivered into the adjacent flexural basin filling it to or above sea level. This late stage is called "molasse" (cf. P. Allen et al., 1986).

The classical "flysch" and "molasse" deposits are well known from the Western Carpathians. However, quite similarly the sedimentary succession of the Palaeogene Basin of northern Hungary also displays a two-stage transgressive–regressive facies cycle (Tari and Sztanó, 1992), which

was interpreted to be the result of compressional tectonics (Tari et al., 1993-this issue). The Palaeogene Basin developed to the south of the backthrusts of the inner Western Carpathian units (Fig. 3) and since the corresponding thrust system is antithetic to the subducting European plate margin (Fig. 3), it is considered to be a retroarc foredeep basin using the terminology of Dickinson (1976). It is important to realize (see Tari et al., 1993-this issue, for a detailed discussion) that while the Carpathians represented an extensional arc during the Neogene with a corresponding back-arc extensional basin (Pannonian basin proper; F. Horváth and Royden, 1981), the Palaeogene arc of the Carpathians was apparently a compressional arc with a corresponding retroarc foredeep basin (Palaeogene Basin).

During Middle Eocene–Late Oligocene times the Palaeogene Basin of northern Hungary was a generally deep, underfilled basin as the result of the thrust load emplacement. In the early Miocene, which is discussed in this paper, predominantly shallow-marine and continental deposition took place indicating the cessation of thrusting in the adjacent thrust-fold belt and the beginning of "molasse" sedimentation. During this time the abundant sediment supply derived from the uplifting and eroding thrust belt gradually filled up the basin. This overfilling eventually led to decay of marine sedimentation and widespread deposition of continental clastics by the mid-Early Miocene (Fig. 5).

However, it is to be noted that the existence of SE-vergent (basement involved) thrusts in the inner Western Carpathians north of the Palaeogene Basin is only hypothetical at present due to a lack of adequate reflection seismic data in northern Hungary. We consider the strongly asymmetric isopachs of the Upper Oligocene (e.g., the main axis of Schlier deposition is a WSW–ENE line, from Szécsény to Putnok, cf. Báldi, 1986) and Lower Miocene formations (e.g., the thickness of Pétervására Sandstone is typically 150 m at the basin margin, but it is more than 600 m in the center of the basin, Hámor, 1985), as indirect evidence for the ongoing compressional tectonic activity.

Superposition of a third-order eustatic signal on tectonics: a discussion

Revolution of depositional systems at the Egerian / Eggenburgian (NN1 / NN2) boundary

We interpret the surface between the shallow-marine Pétervására Sandstone and the underlying Szécsény Siltstone, deposited in significantly deeper water, as a sequence boundary. This surface can be followed in individual outcrops and represents a pronounced basinward shift of facies belts (approximately 20–30 km, see Fig. 7). This sequence boundary may have primarily resulted from a drop of eustatic sea level rather than being of tectonic origin. Accepting the flexural character of the basin, a marked short-term tectonic event such as a thrusting period would cause an overall rise of relative sea level on the foreland flank of the basin (cf. Flemings and Jordan, 1990). Another argument for the eustatic origin of the sequence boundary is its basinwide character. In the western part of the basin (Fig. 4) upward shallowing, regressive brackish water deposits were described (Báldi, 1973, 1986; Becske Beds of Hámor, 1985) and even the formation of a subaerial disconformity (Vass et al., 1979) was supposed at the Egerian/Eggenburgian Paratethyan stage boundary (NN1/NN2 biozone boundary).

At the Egerian/Eggenburgian boundary the size of the marine depositional system decreased dramatically (Nagymarosy and Müller, 1988), and a smaller and shallower sea was formed, when the deposition of the bathyal Szécsény Siltstone stopped in a considerable part of the basin. The corresponding eustatic sea-level drop must have been superimposed on the tectonic uplift of the basin, rather than on its subsidence. Thus, the demarcation of the Egerian Szécsény Schlier and the Eggenburgian Pétervására Sandstone, induced by a drastic decrease of accommodation space can be understood as a tectonically enhanced sequence boundary.

Recently we suggested forebulge uplift as a possible driving mechanism for the enhancement of the sequence boundary (Tari and Sztanó, 1992).

At present, however, we are not able to document the existence of a foredeep unconformity (e.g., P. Allen et al., 1991; Sinclair et al., 1991). Moreover, due to the gradual cessation of thrusting during the early Miocene, the development of a pronounced forebulge is not very likely.

The overall shallowing-upward trend of the late Oligocene to early Miocene sedimentation (Fig. 5), points to a second-order tectonic uplift of the basin, following a period of fairly intensive subsidence. Therefore, we interpret the uplift as a late-stage uplift, commonly observed in flexural basins, as the result of isostatic rebound (e.g., Heller et al., 1988). At present the magnitude and rate of this surface uplift cannot be quantified, but certainly this process dominated the long-term evolution of the early Miocene basin. The shorter-term third-order sea-level changes such as the one discussed above, were superimposed on the long-term uplift of the basin.

It is to be noted that within the resolution of the available biostratigraphic data, we obtained a good match with the eustatic curve of Haq et al. (1987). These authors indicated a marked sea-level drop at the boundary of the NN1 and NN2 nannoplankton zones, which they dated as 21 Ma old. This sea level drop, preceding the Burdigalian sea-level rise, is also well documented from other Paratethys basins (cf. Rögl and Steininger, 1984).

Evolution during the Eggenburgian

Due to the dramatic decrease of depositional depth tidally generated sandwaves and megaripples developed rapidly above the Egerian/Eggenburgian sequence boundary, as shown by facies transitions of the Pétervására Sandstone. Facies unit B and C reached a thickness of 100 m (Fig. 8) and are interpreted to be late lowstand aggradational units. The vertically stable facies boundaries show aggradation of these units over a longer period indicating equilibrium between sediment accumulation and the creation of accommodation space (Jervey, 1988; Posamentier et al., 1988).

Later an increase of accommodation rate occurred, demonstrated by a rapid landward shift of facies belts (Fig. 7). The rate of sea-level rise outpaced sedimentation and the above discussed tectonic uplift of the basin. This rise of sea level may correspond to the one, which caused the "Burdigalian transgression" (cf. Homewood and Allen, 1981; Rögl and Steininger, 1984; Haq et al., 1987).

After the transgressive event, coarsening upward sequences were formed grading from small dunes (B facies) to huge sandwaves (C facies) and to conglomerate lobes (D facies) (Fig. 8). They reflect progradation due to a decrease in accommodation. Finally the flexural uplift of the area and the continuous sediment supply resulted in filling of the basin up to sea level and above (continental beds of the Zagyvapálfalva Formation). The top of the Pétervására Formation is an unconformity, produced by the continued uplift and erosion of the previously deposited basin fill sequence.

Tide-influenced deposition during sea-level fall

The superposition of a eustatic sea-level fall on the late-stage isostatic uplift of the early Miocene flexural basin coupled with its shallow ramp geometry created a peculiar basin setting sensitive to the amplification of tidal movements.

These conditions are different from the commonly described situation in which tidal deposits are related to marine transgression (Van Wagoner et al., 1988, and many others). However, it should be realized that favourable conditions for the activity of tidal processes depend on features such as basin geometry (shallow water depth and confined basin morphology with a wide "inlet" towards the open sea). Such conditions may be met during a sea-level fall as well as during a rise. The important point, however, is that if sea-level fall continues, the tidal deposits which possibly may have formed, may become subject to subaerial erosion. Therefore, the preservation potential of regressive tidal deposits is significantly smaller than that of transgressive ones. This may explain the dominance of transgressive tidal settings and a virtual lack of regressive ones in the sedimentary record. The interference of synsedimentary tectonics was essential to maintain and

conserve the tide-influenced deposition in the early Miocene basin of northern Hungary.

Conclusions

A detailed sedimentological study of the early Miocene Pétervására Sandstone revealed a rapid change of depositional systems from a tranquil deeper marine into a high-energy shallow-marine environment. The new scenario was strongly influenced by tidal currents. The Pétervására Sandstone is consists of a lower aggradational and an upper progradational unit separated by a unit of considerable landward shift of shore-paralell facies belts.

This stratigraphic architecture is interpreted in terms of relative sea-level changes evolving in a flexural basin. The tide-influenced depositional system was undoubtedly initiated by a fall of eustatic sea level, and enhanced by the late-stage isostatic surface uplift of the flexural basin. It is concluded that uplift became the dominant factor in early Miocene basin evolution, though the typically gently inclined ramp geometry of the flexural basin made the depositional setting sensitive to third-order sea-level fluctuations.

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