RIVER DELTA MORPHODYNAMICS: EXAMPLES FROM THE DANUBE DELTA

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Abstract: The Danube, with its mouths at the Black Sea, has been economically and strategically one of the most important rivers in Europe; consequently, its delta has been studied since the Nineteenth Century. Although many morphological and sedimentological aspects of the Danube delta are well understood, its late Quaternary evolution remains ambiguous. This uncertainty reflects in part the complexity of the sea-level variations and water-chemistry changes related to the periodic isolation of the Black Sea during eustatic lowstands, but also a lack of accurate age control of the deltaic deposits. On the basis of a review of existing radiocarbon dates, we propose that the development of the delta at the open coast started approximately 6,000–5,500 ^{14}C years ago, much later than the 9,000 ^{14}C years BP previously suggested.

Morphodynamics of the open-coast Danube delta has been determined largely by the interaction between fluvial deposition and the strong southward wave-induced longshore transport. Morphological and facies asymmetry displayed by the marine lobes of the Danube delta indicate that a strong and sustained southward-directed longshore transport has been a persistent process along the delta shore. Coastal evolution on the adjacent nondeltaic coast is also strongly coupled to the delta morphodynamics via the longshore transport. Analysis of recent deltaic progradation of the youngest open-coast lobes of Danube delta indicates that river-mouth morphodynamics is highly nonlinear, involving multiple feedbacks between subaerial deltaic progradation, deposition on the subaqueous delta, current and wave hydrodynamics as well as wave–current interactions. First, a feedback loop is activated by the hydraulic groin effect of the river plume, which leads to a mutually sustained progradation of the updrift coast and subaqueous delta at the mouth. Second, the development of a shallow subaerial delta platform, strongly offset to the downdrift direction, helps dissipate waves reaching the platform, leading to entrapment of sediment on the platform. Third, increased flood-induced deposition on the subaqueous delta platform, followed by wave reworking, leads to recurrent emergence of barrier islands at its offshore edge; longshore transport is then channeled (i.e., intensified and guided) by the new coast along the barrier, leading to a rapid alongshore expansion of the subaqueous delta in the downdrift direction. Although the sedimentation processes are complex, the resulting morphology at the mouth exhibits a tendency to self-organize that is reflected and preserved by the facies architecture of wave-influenced lobes.

INTRODUCTION

The morphodynamics of deltaic coasts has been a less well explored aspect in the evolution of deltas. Coastal morphological units such as barriers, estuaries, or deltas have traditionally been studied in isolation, ignoring the fact they co-evolve interdependently (e.g., Cowell et al., 2004). Feedbacks between the evolving morphology of a delta and basinal hydrodynamics remain inadequately known, as are the interactions between contemporaneous lobes within a delta (e.g., Komar, 1973, 1998) or the influence exerted by a delta on the dynamics of adjacent nondeltaic coasts and vice versa (e.g., Penland and Suter, 1989; Jiménez et al., 1997; Stanley et al., 1997; Aslan et al., 2003; Corregiari et al., this volume). Although there has been much recent progress in understanding suspended-sediment deposition from river plumes (e.g., Syvitski and Bahr, 2001; Geyer et al., 2004), the dynamics of coarse sediments at river mouths has received little attention, in large part because of the inability of conventional techniques to provide reliable direct measurements. However, it is becoming increasingly clear that river-mouth processes involving bed-load transport exert a primary control on the larger delta morphology and facies architecture (e.g., Wright, 1985; Dominguez et al., 1983, 1987; Giosan, 1998; Rodriguez et al., 2000; Bhattacharya and Giosan, 2003, Fielding et al., this volume; Willis, this volume).

Our paper explores the morphodynamics of the wave-influenced Danube delta with the intent to clarify its Holocene evolution. Although monitored since the middle of the Nineteenth Century, mainly for navigation purposes, many aspects of the evolution of the Danube delta, including a reliable chronology, remain uncertain. We critically review less accessible literature on the subject as well as discuss new data to outline a coherent model for the development of the Danube delta. We further argue that a large-scale morphodynamics approach is necessary for establishing key controls not only for the evolution of any delta plain but also for the adjacent nondeltaic coast and subaqueous delta. This is particularly important for the management of Danube delta, Europe's largest wetland ecosystem, where economic and environmental interests among and within riparian nations are often divergent, leading to environmental pressure both from upstream factors as well as downstream changes in the Black Sea basin (Margesson, 1997; Lancelot et al., 2002).

ENVIRONMENTAL SETTING

Fluvial Regime

The Danube River is the second largest European river after the Volga in terms of catchment area (817,000 km^2) and length

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(2870 km) as well as water and sediment discharge (Milliman and Meade, 1983; Meade, 1996). The climate in the drainage basin, which covers most of the Central and Southeastern Europe, is continental, with Atlantic and Mediterranean influences in the western and southern regions of the basin, respectively (Rimbu et al., 2002). The average annual precipitation, evaporation, and runoff are 816 mm, 547 mm, and 246 mm, in that order (Panin and Jipa, 2002).

The decadal variation of the Danube flow is influenced by the North Atlantic Oscillation (NAO) via precipitation anomalies in the drainage basin, with a lower than normal river flow when the NAO index is positive (Rimbu et al., 2002). The average annual discharge reported between 1840 and 1995 is ~ 6240 m^3/s, with the highest measured annual discharge of ~ 9250 m^3/s in 1970 (Bondar and Blendea, 2000). The peak discharge during the year occurs in late spring (Fig. 1, Diaconu and Mihailov, 1963a; Vörösmarty et al., 1998), following snowmelt; there is roughly two times as much discharge in the spring compared to the fall (Reschke et al., 2002). Extensive damming, especially in the lower basin (Fig. 1), has reduced the suspended-sediment discharge at the apex of the delta from ~ 67 x 10^6 t/year between 1921 and 1960 (Diaconu and Mihailov, 1963b) to ~ 25–35 x 10^6 t/year at present (Panin and Jipa, 2002). The bed-load sediment discharge at the Danube mouths has been estimated to range between ~ 4.6 and ~ 5.3 x 10^6 t/year (Bondar et al., 1992).

In the delta region, the Danube river splits into three main distributaries: the Chilia (Kilia), the Sulina, and the Sf. Gheorghe (St. George). Since 1856, when the first estimates of the flow were made by the European Danube Commission (EDC), discharge through the main navigation route, the Sulina arm, has steadily increased (from 7% to 19%), while the Sf. Gheorghe and the Chilia distributaries have correspondingly decreased their flow (from 30% to 23% and from 63% to 57%, respectively) as a consequence of sustained maintenance and improvement projects for navigation on Sulina and intensive channelization in the delta (Diaconu and Mihailov, 1963a; Panin, 2003). After similar human interventions on the Sf. Gheorghe arm during the 1980s, discharge in this distributary started to increase (Panin, 2003). Suspended load is

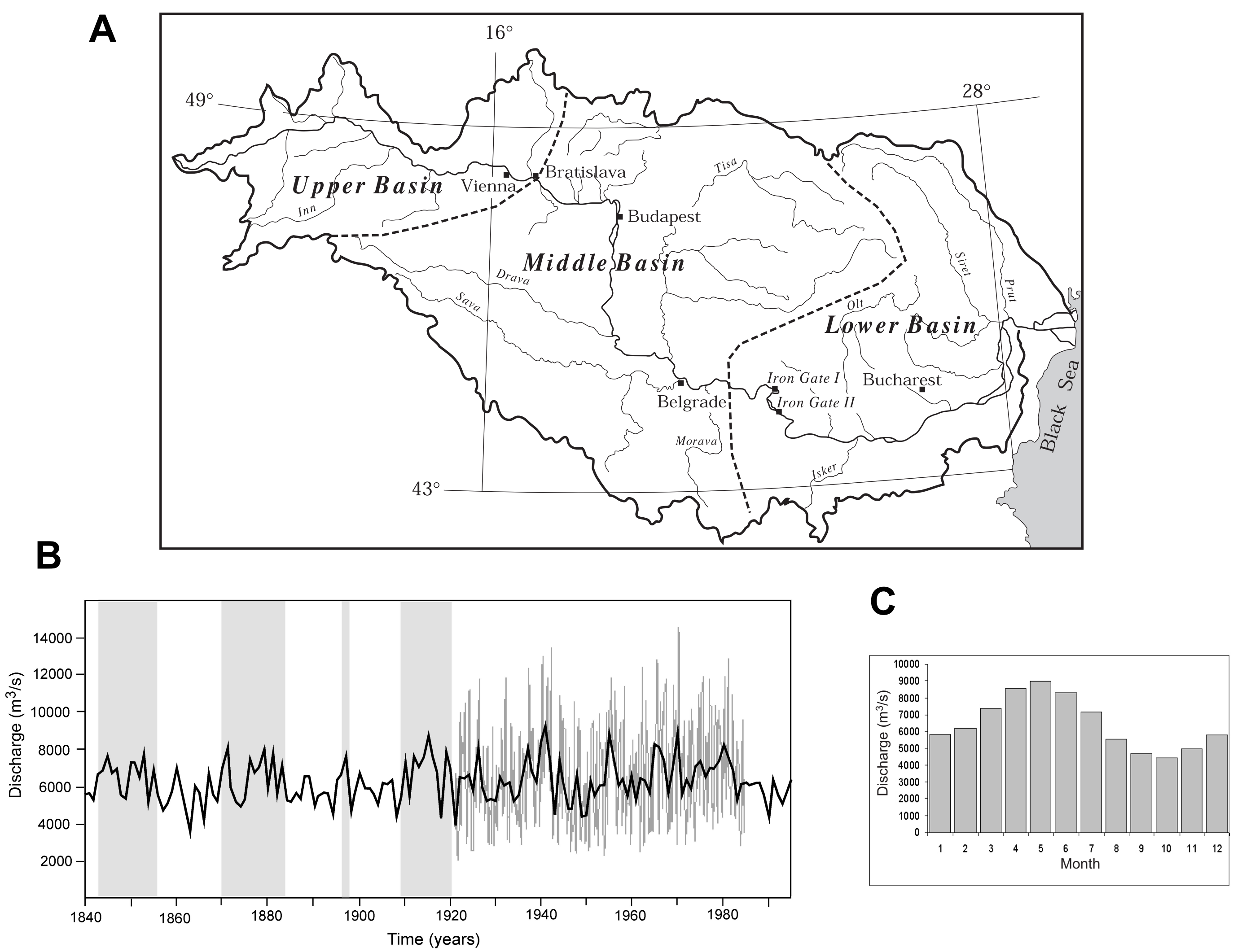

FIG. 1.—**A)** Characteristics of the Danube river drainage basin and discharge. **B)** Discharge between 1840 and 1995 (bold black line, annual average values from Bondar and Blendea, 2000; gray thin line, monthly measurements, Rimbu et al., 2002). **C)** Hydrograph for the discharge of the Danube between 1921 and 1985 (Vörösmarty et al., 1998).

redistributed among the Danube's main distributaries, with Chilia transporting ~ 55% of the total load, Sulina ~ 21%, and Sf. Gheorghe ~ 23%, whereas the bed-load sediment discharge is split 57–65% to Chilia, 19–25% to Sulina, and 19–21% for Sf. Gheorghe (Bondar et al., 1992; Bondar and Harabagiu, 1992).

Basin Characteristics

The Danube flows and builds its delta into the Black Sea, a semi-enclosed basin connected to the World Ocean via the Sea of Marmara and the Mediterranean. Although evaporation (350 km^3/year) exceeds precipitation (300 km^3/year), the average salinity of the Black Sea is low (~ 18 per mil) because of the freshwater (350 km^3/year) provided by rivers draining a large part of Europe and Asia (Özsoy and Ünlüata, 1997). The Danube annually discharges 77% of the total river runoff to the Black Sea and 85% of the runoff entering the northwestern shelf. The resulting buoyant plume, augmented with freshwater by southward coastal current fed by the Nistru (Dniestr), Dniepr, and Southern Bug rivers, propagates onto the shelf and along the coast toward the southwest (Stanev et al., 2002; Yankovsky et al., 2004). The coastal plume further interacts with the western cyclonic gyre of the Rim Current, forming quasi-persistent anticyclonic eddies over the middle shelf (Fig. 2; Oguz et al., 1993; Korotaev et al., 2003; Yankovsky et al., 2004). Via its plume, the quantity and quality of Danube water and suspended-sediment discharge exerts a strong influence on sedimentation, water-column and sediment chemistry, and ecosystems of the northwestern shelf of the Black Sea (Bacescu et al., 1971; Humborg et al., 1997; Lancelot et al., 2002 and papers therein).

The Black Sea is a microtidal basin with semidiurnal tides ranging between 7 and 12 cm. Tides are negligible in comparison to other water-level fluctuations such as seiches and storm surges, which can elevate the sea level at the coast between 1 and 2 m (RCMGG, 1994). Interannual and interdecadal sea-level variations, ranging from 20 to 55 cm along the coast influenced by the Danube plume, appear to correlate with changes in the Danube discharge (Malciu and Diaconu, 2001).

Long-term relative sea-level change on the Danube delta coast at Sulina is 2.47 mm/year (Vespremeanu et al., 2004). South of the delta, at Constantza, reported values for the relative sea-level change vary between 1 and 3 mm/year (see references in Vespremeanu et al., 2004). However, a cursory comparison of the sea-level data from Sulina and Constantza with nearby tide gauges in Bulgaria and Ukraine (reported on the Permanent Service for Mean Sea Level website http://www.pol.ac.uk/psmsl/) suggests that data from Romanian coast is likely affected by significant datum and/or reporting errors and need to be reconsidered in an in-depth examination before using them for studies of sea-level change.

The average wind speed in the northwestern Black Sea is between 5 and 6.5 m/s (Bulgakov et al., 1992). Winds from the north-east quadrant are dominant (Ciulache, 1993). Long-term visual observations of wave height and directions are available offshore Constantza (Giosan et al., 1999). The distribution of wave directions was highly skewed, with most waves (82%) arriving dominantly from the left-hand side relative to the regional direction of the coast (i.e., north-east quadrant); the annual average significant wave height was 0.8 m, with an annual standard deviation of 1 m, and a mean period of 5 seconds (Giosan et al., 1999).

GEOMORPHOLOGY OF THE DELTA PLAIN

With no significant tides in the Black Sea, dividing the delta plain into upper and lower parts, separated at the upstream limit of the tidal influence (Coleman and Wright, 1971) is not warranted. However, two distinct delta-plain regions are recognized (Fig. 2; Antipa, 1915) based on the distribution of channels, flood basins and lakes, and sand-ridge geometries. The "fluvial delta" developed in the former Danube Bay delimited by the Bugeac loess plateau and the North Dobrogean Orogen (Figs. 2, 3). The "marine delta" developed largely outside Danube Bay and exhibits a clear wave-influenced morphology (Figs. 2, 3). Spratt (1860) was the first to propose that the Danube has built its initial delta into an embayment protected by a baymouth barrier extending from the Bugeac toward Dobrogea (Fig. 2); subject to an intense debate among Romanian researchers (e.g., Antipa, 1915; Bratescu, 1922), the barrier hypothesis was strengthened by subsequent work (Zenkovich, 1956, Panin, 1989; see also discussion below). To avoid misnomers, we will use "internal delta plain" as the generic name for bayhead and lacustrine delta lobes built within the Danube bay and "external delta plain" for the wave-influenced lobes developed outside the bay, in front of the Danube baymouth barrier.

Marine beach ridges are absent in the internal delta plain, indicating a total lack of wave influence (Fig. 4B); a network of bifurcating/anastomosing, active and abandoned distributary channels/levees indicates development in a sheltered environment (Fig. 4A). The channel network partitions the delta plain into numerous flood basins covered by marshes and lakes (Fig. 4C). The distributary channels are organized into drainage systems that define three separate lobes (Fig. 4A). The largest drainage system, corresponding to the Tulcea lobe, which probably developed initially as a bayhead delta (Antipa, 1915), consists of channels associated upstream with the Tulcea distributary and the upper course of the Chilia distributary, and farther downstream with the Sulina and the Sf. Gheorghe, after they split from the Tulcea branch (Fig. 2, Fig. 4 A ,D). The channel density decreases between the upper and the lower Tulcea lobe (Fig. 4 C, D).

The morphology of the predeltaic relief played a limiting role in the direction and rate of progradation of the delta, as proposed by Murgoci (1912), Antipa (1915), and later Ghenea and Mihailescu (1991). The Tulcea lobe probably advanced in the direction of the alluvial valley along the general direction of the modern Sf. Gheorghe distributary (see data in Liteanu and Pricajan, 1963), which continues on the shelf with a network of buried channels leading to the Danube (Viteaz) canyon on the slope (Popescu et al., 2004). Two smaller drainage systems are associated with the Chilia arm (Fig. 4 A, D); they are separated from the Tulcea lobe by lacustrine beach ridges and from each other by the Chilia loess promontory, which juts south from the Bugeac plateau (Fig. 4B; Antipa, 1915; Popp, 1961; Panin, 1983). The loess relief below these drainage systems (Ghenea and Mihailescu, 1991) suggests that the upstream system, Chilia I lobe, had evolved as a lacustrine delta in a shallow depression on the loess platform. Chilia II lobe developed subsequently as a lacustrine or a bayhead delta in a small protected embayment (Fig. 4D; Popp, 1961). Both the morphology of Chilia I and II lobes as well as lithology of short cores described there (Popp, 1961; Munteanu, 1996) suggest sediment deposition on channel levees as well as in flood basins, which is typical for anastomosing channels (Makaske, 2001), rather than for braided channels as previously proposed (Panin, 2003).

The external delta plain consists of several laterally offset lobes developed by deltaic distributaries after they reached the open coast. Amalgamated beach ridges form extensive beach-ridge plains that are capped in places by dune fields with heights up to ~ 12 m (Fig. 4B). Non-amalgamated beach ridges organize in barrier-beach plains, resembling cheniers, with a succession of

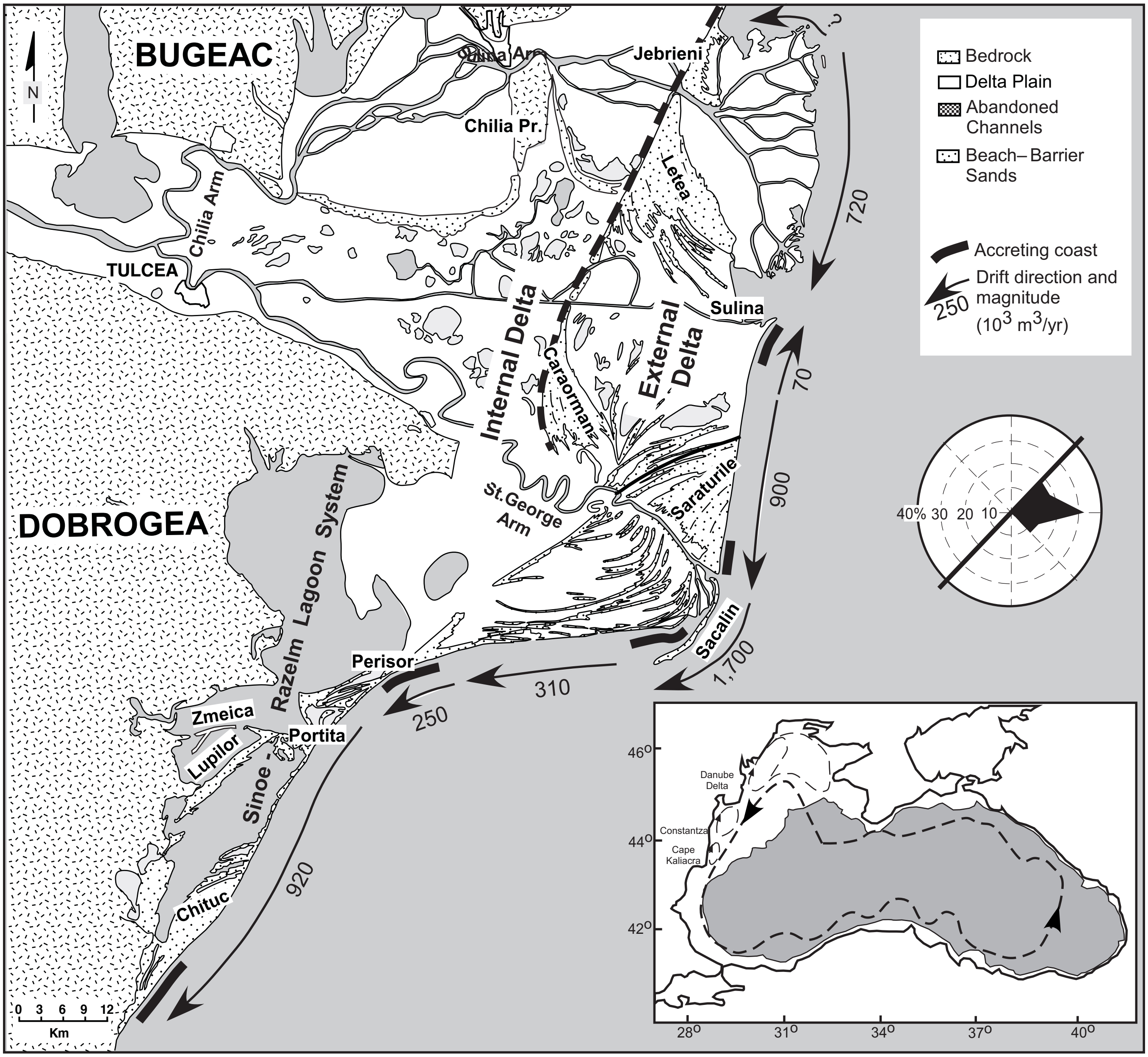

FIG. 2.—Danube delta basemap and oceanographic processes. Lower inset surface circulation in the Black Sea. Thick dashed line, the Rim Current; thin dashed lines, quasi-persistent anticyclonic eddies over the shelf influenced by the Danube plume; the deep basin beyond the shelf is in gray (after Yankovsky et al., 2004). Middle inset: wave direction frequency relative to the average orientation of the coast (thick line). Under this wave climate, the longshore drift along the coast is extremely strong with a general southward direction. Prograding sectors of the coast south of Sulina are indicated. Main figure: The internal delta consists of several baymouth and lacustrine delta lobes that were built inside Danube bay, separated from the Black Sea by a baymouth barrier (thick dashed line indicates the probable orientation of the barrier; Panin, 2003). The external delta is composed of lobes built at the exterior of the baymouth barrier. Once reaching the open coast, the southern and central main distributaries (the Sf. Gheorghe and the Sulina) built three wave-dominated lobes—the beach ridge plains of the Caraorman, the Letea, and the Saraturile represent the northern halves of these lobes and were built by longshore drift. The other main distributary (the Chilia) built a fluvial dominated lobe that is now strongly modified by waves—north of this lobe, the longshore drift built another beach ridge plain, the Jebrieni.

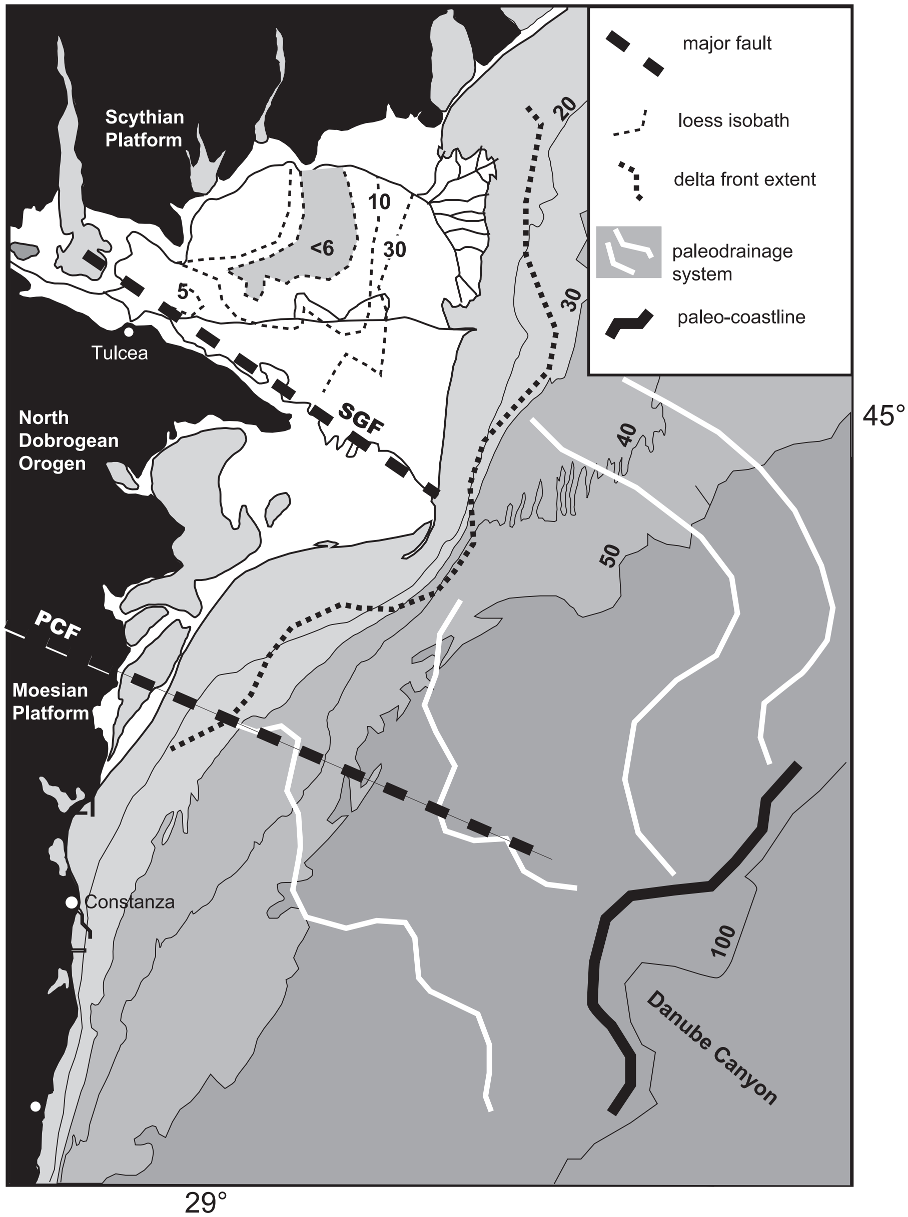

FIG. 3.—Tectonic setting of the Danube delta and the bathymetry of the Danube continental shelf (after Panin, 1989; Popescu et al., 2001; Popescu et al., 2004). The delta has developed in a structurally controlled embayment overlying the North Dobrogean Orogen and the Pre-Dobrogean Depression on the Scythian Platform. Major faults: SGF, Sf. Gheorghe fault; PCF, Peceneaga–Camena fault. Loess occurs within deltaic deposits (isobaths from Ghenea and Mihailescu, 1991). Extent of delta front on the shelf is after Panin (1989), and paleo-drainage systems on the shelf are after Popescu et al. (2004).

elongate sandy ridges separated by mud-filled marshy/lacustrine elongate depressions (Fig. 4 B, C). A detailed geomorphological description of the beach-ridge systems of the external delta is presented by Panin (1974).

Beach ridges have long been recognized to represent former shorelines of the delta (Bratescu, 1922); their groupings in sets that are laterally juxtaposed allows the identification of the lobe development sequence for the "marine delta" (Fig. 4 B, D). The Caraorman beach-ridge plain is the landwardmost ridge system and outlines the oldest marine lobe built by the Sf. Gheorghe distributary. The Letea beach-ridge plain system and its age-equivalent barrier-ridge plain located north and south of the Sulina branch, respectively, delineate a younger lobe, currently in the abandonment phase, built by the Sulina. On the seaward side of the Sulina lobe, the Sf. Gheorghe and Chilia constructed the youngest lobes of the Danube delta that are still currently active. The Sf. Gheorghe II lobe exhibits a similar morphologic asymmetry, with an updrift beach-ridge plain (Saraturile) and a barrier plain on the downdrift side, whereas the Chilia has built a largely river-dominated lobe (Chilia III). The number of secondary distributary channels is much reduced for the wave-dominated lobes compared to the bayhead/lacustrine deltas from the internal delta or the open-marine Chilia III lobe (Fig. 4C). Lakes and flood basins are substantially fewer in most of the external delta

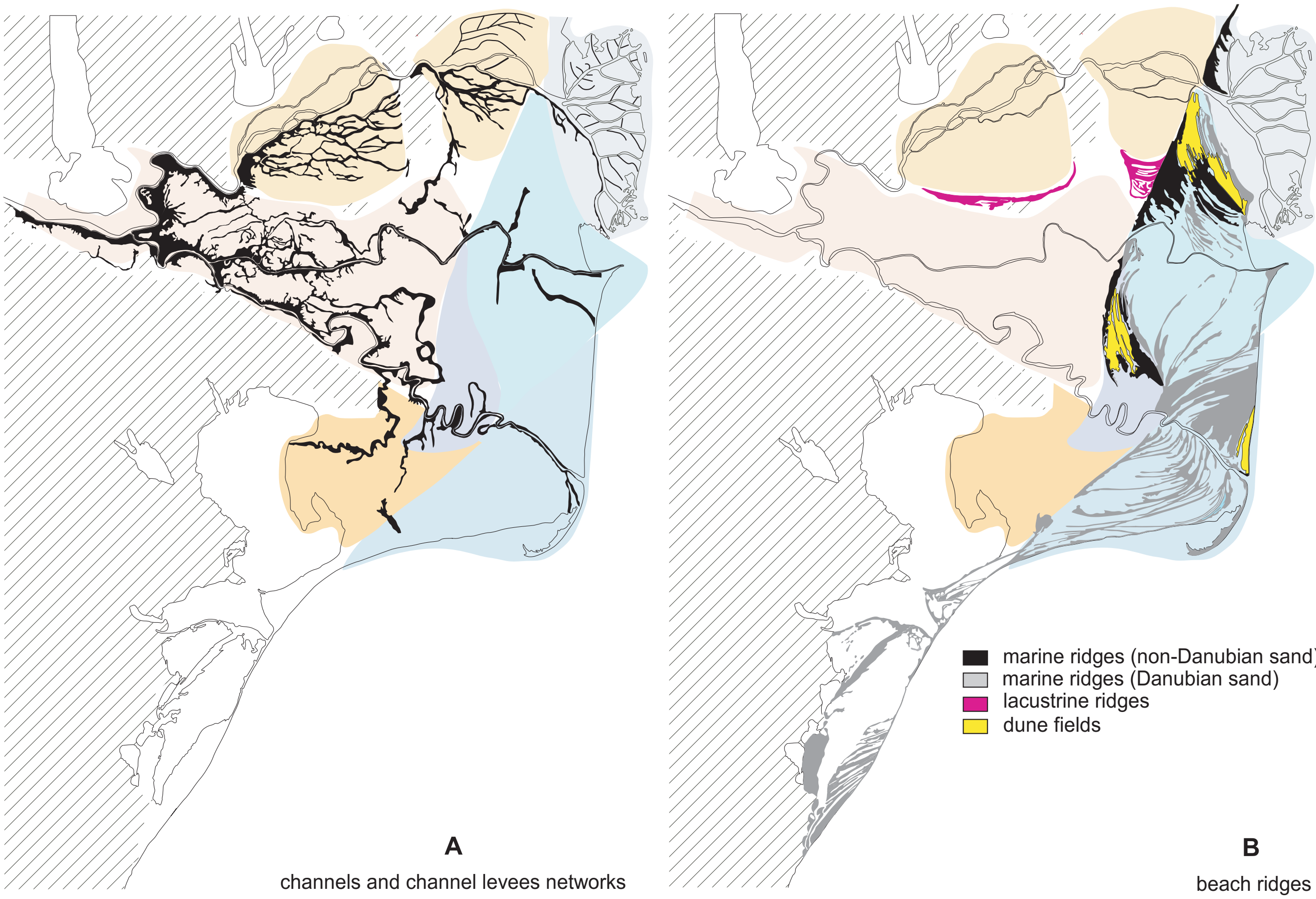

FIG. 4.—Geomorphology of the Danube delta plain (after Munteanu, 1996; Diaconu and Mihailov, 1964; Panin, 1989). Interpreted individual lobes of the delta are in warm colors (bayhead and lacustrine) and cold colors (open-coast lobes). **A)** Network of channels and channel levees (in black); **B)** sandy beach ridges.

compared to the internal delta plain, with the notable exception of the southern half of the Sulina lobe (Fig. 4C).

The deltaic coast continues to the south with a series of baymouth barriers fronting the large lagoon system of Razelm–Sinoe (Fig. 2, Fig. 4 B, C). At least three generations of barriers can be distinguished: Zmeica, Lupilor, and Chituc; the latter two display amalgamated beach ridges at their southern extremities. Another deltaic lobe, the Dunavatz, developed, at least in part, under bayhead and/or lacustrine conditions, filling the northeastern sector of the Razelm lagoon (Fig. 4D). No beach ridges are evident in the Dunavatz lobe morphology, and the number of channels/levees and lakes/flood basins is much reduced compared to the other bayhead/lacustrine lobes discussed previously (Fig. 4).

WAVE INFLUENCE ON DELTAS—IMPLICATIONS FOR THE DANUBE DELTA

On the basis of a survey of the morphology of several deltas and/or deltaic lobes, Bhattacharya and Giosan (2003) introduced a conceptual model for wave-influenced deltas that classifies them into symmetric, asymmetric, and deflected types (Fig. 5). At river mouths where the net longshore sediment transport is small, wave-influenced deltas are symmetric, with beach ridges developing on the interdistributary coasts, centered on each distributary mouth (Fig. 5). The delta planform is arcuate to cuspate, with straight or gently curved shorelines (Fig**.** 5; e.g., Wright and Coleman, 1973; Bhattacharya and Walker, 1992).

In contrast, where longshore transport is relatively strong and unidirectional at the river mouth, it can interact with the fluvial delivery of sediment, leading to an asymmetry in the morphology and facies distribution of the delta (Fig. 5). The river plume acts as a hydraulic groin and obstructs the longshore transport of sediment, which converges at the mouth (Todd, 1969; Komar, 1973); although some of the sand is probably bypassed by waves around the mouth, most of the sediment is deposited along the updrift shore, forming a new set of beach ridges (see Bhattacharya and Giosan, 2003, and references therein).

In the extreme case where the influence of longshore transport is dominant over a relatively low or episodic fluvial discharge, a deflected delta develops (Fig. 5; Bhattacharya and Giosan, 2003). In this case, the mouth of the river runs subparallel to the coast, with the river being separated from the sea by a sandy spit-levee (Wright, 1985). The delta progrades as a series of randomly distributed, quasi-parallel sand spits and channel fills.

The modern Sf. Gheorghe lobe of the Danube delta serves as a type example of an asymmetric delta. The updrift wing of the lobe consists of a succession of coalesced beach ridges, the

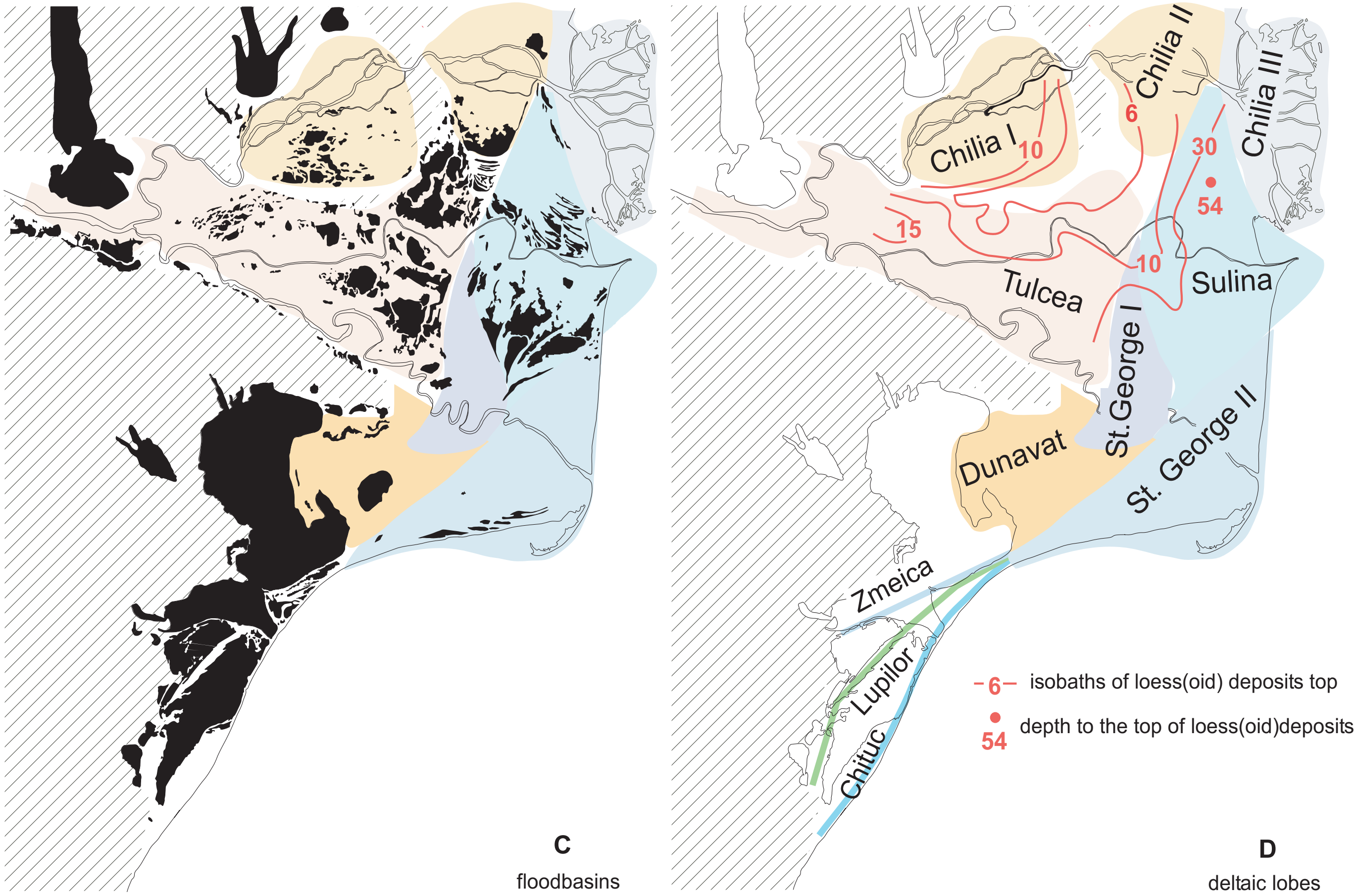

FIG. 4.—(Continued) **C)** flood basins (in black); **D)** individual lobes of Danube delta and associated baymouth barriers developed south of the delta (isobaths of loess deposits are from Ghenea and Mihailescu, 1991).

Sărăturile Formation (Fig. 4). The downdrift wing is formed by a subparallel series of sandy "shoestring" ridges encased in delta-plain muds (Diaconu and Nichiforov, 1963c; Banu and Rudescu, 1965). These ridges apparently originated as barrier islands on the subaqueous delta (e.g., Diaconu & Nichiforov, 1963c; Giosan, 1998). Sands updrift of the Sf. Gheorghe mouth are texturally more mature than the sands downcoast (Romanian Centre for Marine Geology and Geoecology, 1994; Giosan, 1993) suggesting that the updrift Sărăturile formation has not received a significant amount of fluvial material from the Sf. Gheorghe distributary, but instead has been built by sediment eroded from the Sulina lobe.

Other lobes of the external Danube delta plain developed asymmetrically or passed through asymmetric stages. The first lobe of the Sf. Gheorghe, the Sulina lobe, and even the incipient modern lobe of Chilia, exhibit beach-ridge plains on their northern wings (Fig. 4; Caraorman, Letea, and Jebrieni formations, respectively). Panin (1989) used the textural characteristics, mineralogy, and bulk chemistry of sediments to show that most of the Caraorman beach-plain ridge consists of sand transported alongshore by waves from north of the Danube delta, with little or no contribution from the Sf. Gheorghe distributary (Fig. 4B). The youngest sets of beach ridges on the Caraorman, however, are composed of sand with a distinct Danubian origin, suggesting that they formed after the Sulina distributary began to discharge updrift of the Sf. Gheorghe I lobe (Fig. 3B; Panin, 1989; see also discussion below about the evolution of Sulina lobe). The alternance between barrier islands and marshes/lakes on the downcoast wings is not preserved or has no surface expression in the of Sf. Gheorghe I lobe. Banu and Rudescu (1965) describe the subsidence and burial of barrier ridges within encasing muds on the downdrift wing of the modern Sf. Gheorghe lobe; a similar phenomenon must have affected the barrier ridges of the first Sf. Gheorghe lobe, in addition to the post-abandonment meandering of the distributary (Fig. 3A), which destroyed part of the initial architecture of the delta plain.

The Sulina has a more complex planform morphology than the conceptual model for asymmetric deltas introduced by Bhattacharya and Giosan (2003). The Letea formation is not a simple updrift beach-ridge plain. In its early stages, the Sulina split into several secondary channels (Fig. 3A; Panin, 1974), and the incipient beach-ridge plain formed updrift of the northernmost of these channels was made of non-Danubian material brought in via longshore transport from the north (Fig. 3B; Panin, 1989). Downdrift of this channel, beach ridges, built of Danubian sand or of mixed origin, are generally non-amalgamated and separated by lakes or lowland areas filled with fine sediments (Fig. 3C). When the northernmost distributary was abandoned, the beach-ridge plain of northern composition expanded southward to the mouth of the next active channel of Sulina (Fig. 3B).

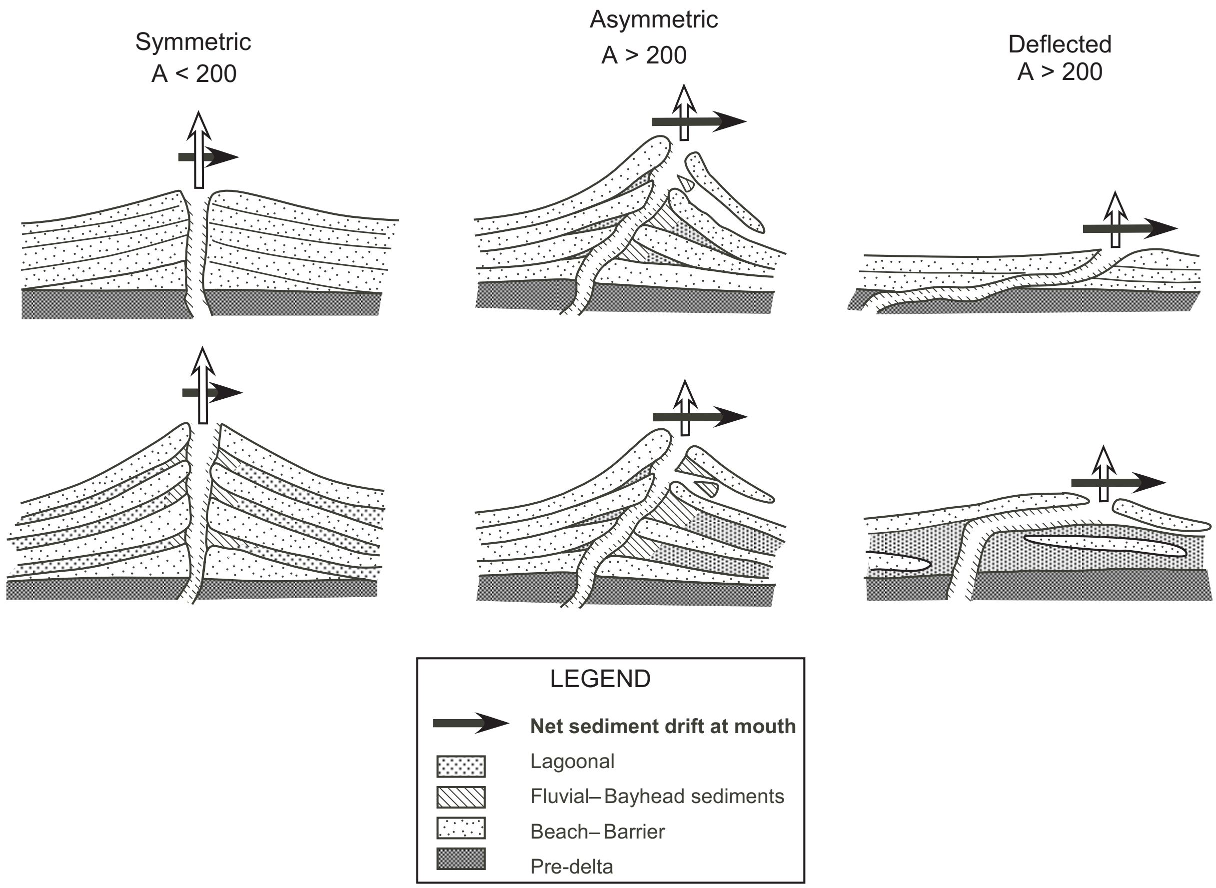

FIG. 5.—Process diagram for wave-influenced deltas (from Bhattacharya and Giosan, 2003). Generalized delta morphologies corresponding to different values of the asymmetry index are shown. The upper row includes deltas preserving a lower proportion of fluvially derived mud, and the bottom row represents examples of deltas comprising more heterolithic deposits. The ultimate proportion of sand relative to fine sediments in a wave-influenced delta may be affected by factors other than the ones considered explicitly in the asymmetry index, such as sediment caliber or flood frequency, that could translate to variations in the morphology of the symmetric, asymmetric, and deflected wave-influenced deltas.

The youngest sets of the beach-ridge plain, however, were built of Danubian material (Panin and Panin, 1969), and it has been proposed that once the Chilia distributary reached the open coast and started to build its delta, Danubian material became the dominant sediment of the longshore drift (Panin, 1989). However, it is also possible that the Danubian material has come entirely or at least in part from the Sulina itself. As the Sulina lobe prograded, its updrift shoreline rotated counterclockwise. Consequently, the ENE dominant waves approached the updrift coast of the delta from a progressively more normal direction, leading to a gradual reduction in the longshore transport rate toward the mouth (Komar, 1973; Pranzini, 2001). Further progradation could have lead to a reversal in the drift direction which provided sand of Danubian origin for the youngest set of beach ridges from the Sulina mouths and later from the reworking of the lobe's apex.

The ridges of the Jebrieni plain, updrift of the youngest Chilia lobe, are composed of northern, non-Danubian sand (Barkovskaya, 1948), which confirms their longshore-drift origin. However, it is unclear if the lobe exhibited asymmetry in an early wave-influenced stage, or if the formation of the first sets of the ridge plain are the result of the longshore drift being obstructed by the river-dominated lobe itself that was rapidly prograding (see discussion below).

The asymmetry displayed by older marine lobes of the Danube delta indicate a strong and sustained southward-directed longshore transport, suggesting that the wave-driven morphodynamics of the deltaic coast has played an essential role in the development of the external delta plain.

COASTAL MORPHODYNAMICS

At present, the Chilia lobe is prograding at the mouths of the main sub-distributaries, the Sf. Gheorghe lobe appears to be in equilibrium at the shore, and the Sulina lobe is being reworked by waves (Giosan et al,. 1999). The delta front extends to water depths between 15 and 45 m, depending on both the accommodation provided by the antecedent morphology of the continental shelf and by the intensity of progradation of active or relict lobes (Fig. 2; Panin, 1989). The prodelta does not cover the whole width of the shelf, but extends to water depths of 50–60 m, ending offshore into coarser-grained palimpsest sediments (Panin, 1989). The alongshore extent of the prodelta is offset to the south, and its southward limit is apparently diffuse (Panin, 1989). Suspended sediments from the Danube plume entrained in the buoyancy-driven coastal current extend at times well to the south of the delta, toward the Bosporus (Yankovski et al., 2004).

The depth of the delta front is significantly greater than the closure depth that defines the nearshore zone, which occurs at ~ 9 m (Giosan et al., 1999). Intense wave reworking keeps the nearshore sandy zone between distributary mouths to the depth of closure (Giosan et al., 1999). However, the texture of bottom sediments beyond the closure depth suggests that the delta-front foresets are a mixture of sand and mud (Panin et al., 1986). Early cores taken on the mouth bar at Sulina showed interstratified muds and sands (Hartley, 1862, 1894–95).

Longshore-Drift System

Giosan et al. (1999) analyzed the longshore sediment-transport pattern along the Romanian sector of the deltaic and associated lagoonal coast. They determined sediment budgets based on shoreline change rates of Vespremeanu and Stefanescu (1988) and on numerical modeling of the potential longshore transport using a wave-energy-flux method based on wave characteristics measured between 1972 and 1981. The calculated longshore transport rates drift were found to be extremely high, with average values of ~ 900,000 m^3/year along sectors of the coast directly facing the dominant ENE waves (i.e., Sulina–Sacalin and the southern part of the Razelm–Sinoe baymouth barrier) and reaching a maximum along the Sacalin barrier, where the nearshore slope is steepest (Fig. 2).

The quantity of sand transported in the nearshore system is of magnitude similar to the amount of bed load delivered by the Sf. Gheorghe or the Sulina (~ 1 million m^3/year; Bondar and Harabagiu, 1992). If the Chilia delivers a similar ratio of bed load to suspended sediment as compared to the other distributaries, we can estimate its bed-load discharge at ~ 3 million m^3/year. This value is significantly higher than for estimates of longshore transport rate along the Chilia coast (~ 700,000 m^3/year to the south; Shuisky, 1984) and has allowed an intense progradation of the delta plain. The mouth of Sulina is heavily engineered and has been dredged periodically since the 1860s (Giosan et al., 1999; Panin, 2003).

A first-order agreement exists between the net transport patterns resulting from the two independent approaches employed by Giosan et al. (1999), suggesting that the coastal dynamics at decadal time scales is controlled largely by the magnitude and direction of the longshore transport. However, the morphodynamics of barrier sectors of the coast is also influenced by overwash, in addition to the longshore transport. Moreover, in the dynamic regions at distributaries mouths, assumptions made by the model (e.g., no feedback between sedimentation and nearshore wave climate and no wave–river plume interactions) do not hold and should be explored further.

Sfantu Gheorghe Mouth

A long series of shoreline and bathymetric surveys exists for the Sf. Gheorghe mouth (Fig 5) performed by the European Danube Commission (EDC), by the Hydrographic Office of the Romanian Navy, and by several research and development groups. Under EDC's management of the lower Danube, the mouth was considered a better alternative for navigation than Sulina (Rossetti and Rey, 1931), but a perennial lack of funding prevented the development of Sf. Gheorghe arm as a shipping channel; consequently, the mouth has evolved under natural conditions. Several important phenomena can be identified in the morphologic evolution of Sf. Gheorghe mouth, although a quantitative analysis of bathymetric changes was precluded because original soundings for many early charts were not available.

The subaqueous lobe built by the Sf. Gheorghe branch is offset almost completely to the south of the mouth (Fig. 6) in the direction of the longshore transport and the preferred orientation of the distributary's plume (Hartley, 1862; Bondar, 1964). In early surveys until the 1900s, the shoreline updrift of the mouth was also offset toward the offshore, probably as a result of the hydraulic groin effect, which promoted its progradation (Giosan, 1998). This geometry of the updrift shore, in turn, sheltered the mouth from dominant waves (Antipa, 1915), favoring deposition on the subaqueous lobe. Alternatively, the intense progradation of the updrift coast could have resulted from the obstruction of the longshore transport system by the subaqueous lobe; both scenarios suggest feedbacks between morphology and hydrodynamics. During the pre-1900s, the subaqueous lobe had built an extensive shallow platform defined by a break in slope at ~ 2 m water depth. A highly depositional, friction-dominated regime for the river plume at the mouth (Wright, 1985) is suggested by a series of islets that probably emerged from middle-ground bars (Fig. 6). Gradually, the original islets increased in size by accretion at both their downstream and upstream sides; others appeared as new channel-margin and middle-ground bars emerged, splitting the distributary in three channels with the updrift, northern secondary channel as the favored path for discharge.

In 1902, a mostly submerged longshore bar, approximately 4 km long, was noticeable on the delta platform (Fig. 6). By 1909 the bar had already emerged as the Sacalin barrier island (Antipa, 1915). Early studies by EDC at the Sulina mouth have shown that during floods the mouth bar becomes more extensive and shallow under intense sedimentation, but is displaced offshore; after the flood, the bar migrates onshore under waves (Hartley, 1894–95). Thus, similarly to other deltas (e.g., Rodriguez et al., 2000), the barrier at Sf. Gheorghe was probably built by waves reworking sediments delivered by extreme river floods at the end of last century (Bratescu, 1922; Vespremeanu, 1983; see discharge peak in Fig. 1 for floods in 1895 and 1897). By 1935, the Sacalin barrier doubled in length while rolling over to the mainland under the influence of overwash and breaching processes (Giosan, 1998). The subaqueous lobe appears to have continued to prograde into the dip direction until 1962, albeit slowly, but it became flatter as the delta platform retreated with the barrier island toward mainland. However, after Sacalin's emergence, the subaqueous lobe elongated to southwest at a dramatic rate of over 200 m/year, compared to less than 100 m/year previously. The hydraulic-groin effect of the river plume continued to obstruct the longshore transport of sand, as indicated by the slow progradation of the coast updrift of the mouth and by the submerged levee-spit consistently flanking the northern channel on the updrift (Giosan, 1998).

The reduction in sediment discharge experienced by the Danube, after dams were built on its lower course in the 1970s and 1980s, was bound to affect the evolution of the Sf Gheorghe lobe, including its subaqueous part. In fact, since Sacalin's northern tip has welded to the mainland in the early 1980s, there have been no discernible signs of a return to the development of a subaqueous lobe with a wide, shallow platform near the mouth as in 1856; on the contrary, the subaqueous lobe shows signs of erosion (Fig. 6).

A multi-phase conceptual model for asymmetric delta development (Fig. 7), based on the evolution of the Sf. Gheorghe mouth (Vespremeanu, 1983; Giosan, 1998), was proposed by Bhattacharya and Giosan (2003). To further explore the development of the Sf. Gheorghe subaqueous lobe, we performed wave transformations using the STWave model (Smith et al., 2001) on two bathymetries, with and without a barrier island. For each bathymetric configuration, an incident wave field representative of average

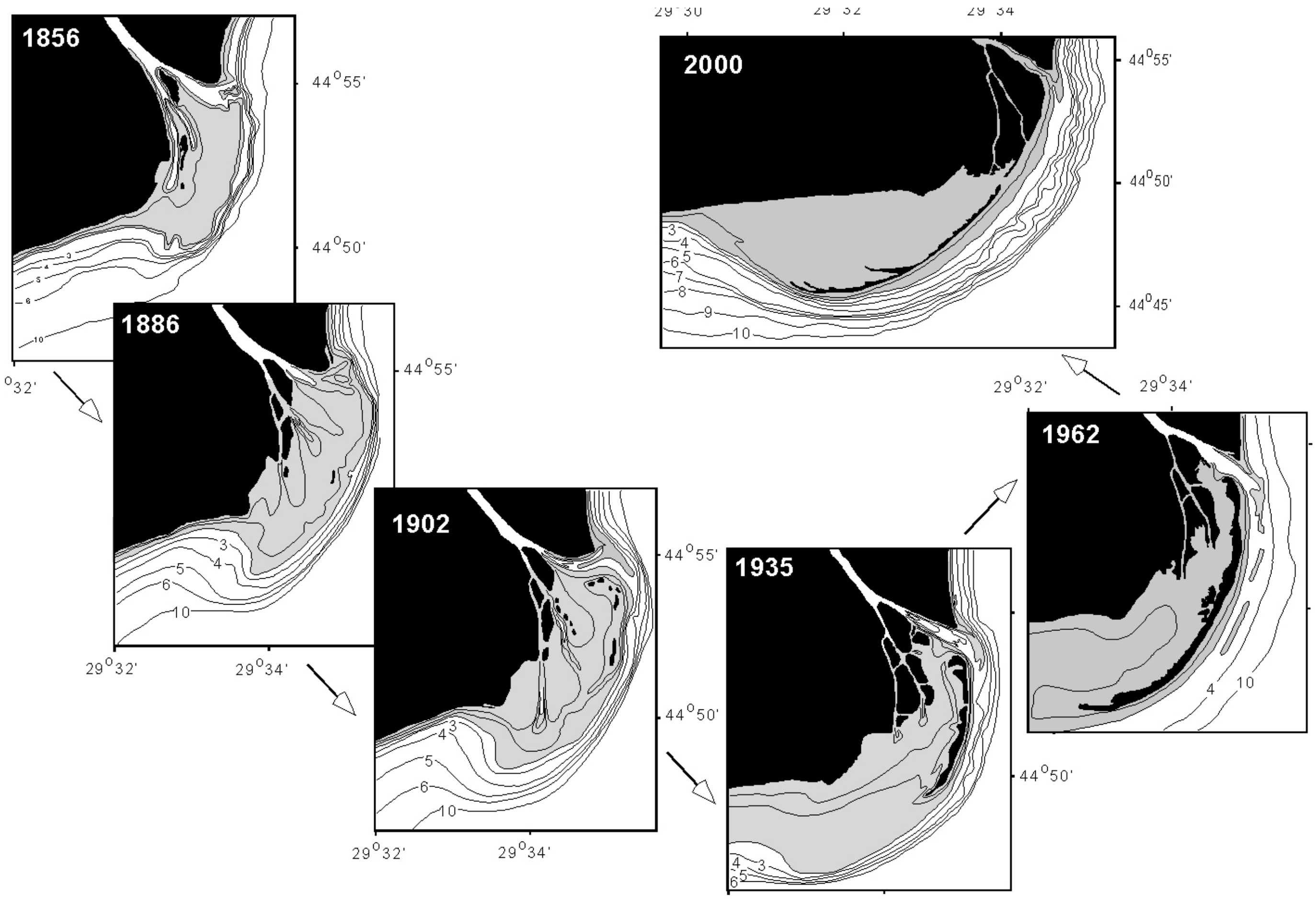

FIG. 6.—Evolution of Sf. Gheorghe mouth between 1856 and 2000. Land is in black and the delta platform (shallow than –2 m) is in gray. Note that intervals for bathymetric contours are not the same on all charts. The subaqueous lobe is strongly offset to south of the mouth whereas the coast north of the mouth is offset toward the offshore relative to the southern mainland coast for 1856–1902 interval. Subaqueous spit-levees are evident north of the mouth from 1902 till 2000. A barrier island (Sacalin) emerged after 1902; the barrier elongated southwestward while rolling over to the mainland coast, where it attached with its northern tip in the 1970s.

conditions in this region (i.e., a height of 1 meter and a period of 4.5 seconds) was propagated across the subaqueous delta from the southeast, east, and northeast directions. Fig. 8 presents the wave field for both bathymetric cases. For simplicity, only waves coming form northeast are shown, but conclusions are similar for other wave directions. Because wave–current interactions were not considered in this simulation, our exercise did not assess the hydraulic-groin effect of the plume.

When no barrier island is present, the shallow delta platform shelters the mainland deltaic coast as it dissipates almost all the wave energy. The resulting longshore transport is therefore reversed toward the mouth. Although at a larger scale, this phenomenon is similar to the reversal of the wave-driven sediment transport on the downdrift side of ebb shoals at inlets, because of the sheltering provided by the shoal itself (e.g., FitzGerald, 1984). The convergence of the sand transport at the mouth is a result of the dynamic interaction between the morphology and wave hydrodynamics and has the potential to act as a positive-feedback mechanism in the development of a subaqueous delta, leading to better entrapment of sands delivered as bed load by the Sf. Gheorghe distributary.

When a barrier is present, the nearshore is much steeper and incoming waves reach the coast of the island less refracted and shoaled than in the no-barrier case, producing an intense longshore transport that is guided and redirected to the south, along the shore of the barrier island (Giosan et al., 1999). This phenomenon, which we term longshore-drift channeling, is another previously undetected feedback loop developed between morphology and wave hydrodynamics, contributing to the dramatic southward extension of the subaqueous lobe after Sacalin's emergence. Frictional dissipation of the distributary's plume over the subaqueous delta should also be expected to decrease, as the bulk of the plume will be guided along the seaward barrier island shore instead of moving over a shallow delta platform.

Both the offset of the subaqueous lobe relative to the river mouth and the presence of a barrier island provide sheltering against dominant waves to the downdrift mainland coast; as a result the coast is presently prograding at Perisor (Fig. 2; Giosan et al., 1999). On the other hand, sheltering of coast updrift of the river mouth is minimal because of a lack of an extensive subaqueous delta; this unequal redistribution of wave energy is different

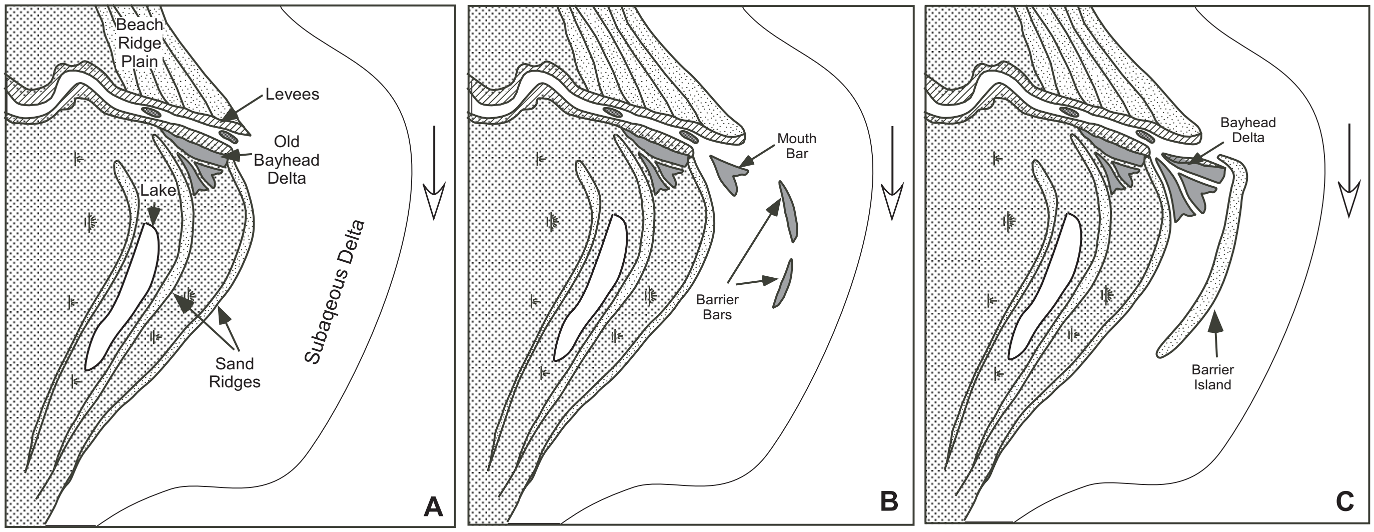

FIG. 7.—Conceptual evolution model for the modern Sf. Gheorghe deltaic lobe (from Bhattacharya and Giosan, 2003): **A)** Subaqueous delta phase: sediment deposition is primarily on the subaqueous part of the delta; the beach ridge plain on the updrift flank is also advancing; **B)** Middle-ground bar phase: a middle-ground bar forms at the mouth, forcing the distributary to bifurcate; linear barrier bars form on the subaqueous delta; **C)** Barrier-island phase: the linear barrier bars coalesce and become emergent to form a barrier island that rolls over to attach to the mainland; a secondary river-dominated bay-head delta may develop in the sheltered lagoon behind the barrier island. Longshore drift (represented by the white arrow) is southward.

than the one envisioned for a subaqueous delta symmetrically developed around a deltaic lobe (e.g., Komar, 1973).

Barriers

Sacalin Island has increased in length at rates over 100 m / year under the influence of high longshore sediment transport rate (Giosan, 1998, Vespremeanu et al., 2004) However, the shoreline progradation predicted for the southern half of the island by the convergence in the modeled sediment transport (Giosan et al., 1999) is not supported by the data on shoreline change, which show a continuous retreat along the entire length of the island. The barrier is actually rolling over toward the mainland (Fig. 6; Giosan, 1998) in an overwash mode (e.g., Kana, 1996). The southern half of the barrier retreats faster because it is narrower and has a lower elevation than the northern part, being more likely to be overtopped and breached during storms (Giosan et al., 1999). In turn, the mobility of the northern tip of the Sacalin shore has decreased considerably since the island joined the mainland in the late 1970s, in response to the reduction in accommodation space available for overwash deposits in the backshore.

Because the wave climate was constantly energetic, driving a large southward drift during the development of the external delta (see considerations above), the evolution of the Sacalin barrier could serve as a model for the development of the older barriers that are segmenting and closing the Razelm–Sinoe lagoon. On the outermost barrier that fronts the open sea, a similar rollover behavior is encountered. The retreat of the shoreline is also greater along narrower stretches that have ample accommodation provided by the lagoon behind (Vespremeanu and Stefanescu, 1988). Each baymouth barrier probably evolved from an elongating spit with sediments delivered from the deltaic coast via the longshore-transport system. Based on Sacalin's rates of elongation, if the bay was initially shallow, the spits could have closed it over several centuries. If the bay was deep, construction of a barrier-island platform would have been necessary; recurves apparent at the southern tips of the two last generations of barriers (i.e., Lupilor and Chituc) might be indicative of such development in deeper waters.

Storm-induced washovers and breachings probably rotated the spit into the lagoon as its tip rolled over faster than its root, which was more stable or even prograding in the wave shadow zone provided by updrift deltaic lobes (Fig. 2). This rotation of the barrier, combined with the accretionary regime at its root, likely promoted the development of a new spit extending from the deltaic coast. Once this second spit closed the bay, the barrier from the first generation was entirely protected from open coast waves and subject only to the considerably less energetic lacustrine wave climate.

Chilia Delta

Detailed surveys of the Chilia delta extend back to 1830. Because sediments from the rapidly prograding Chilia delta have always endangered navigation at the Sulina mouth (Hartley, 1862, 1873, 1894–95; Kühl and Hartley, 1891; Ward, 1929–30), EDC surveyed the delta in 1871, 1883, 1894, 1906, and 1922. In addition, secondary channels of the Chilia have been charted as potential shipping routes by the Russian government (1830) and by Captain Spratt of the British Admiralty (1856). Because we have not yet been able to acquire all original bathymetric data, only a qualitative analysis of the evolution of the subaerial lobe is presented using the EDC's shorelines from Vasilesco (1929), a later shoreline derived from aerial photographs (Slanar, 1945), and an ASTER satellite image from 2003.

The Chilia lobe has evolved as a typical river-dominated delta in a frictional regime, which led to repeated bifurcations via formation of middle-ground bars (Fig 9). The striking disparity between the morphology of the Chilia lobe and all other lobes of the external delta that are wave dominated is probably the

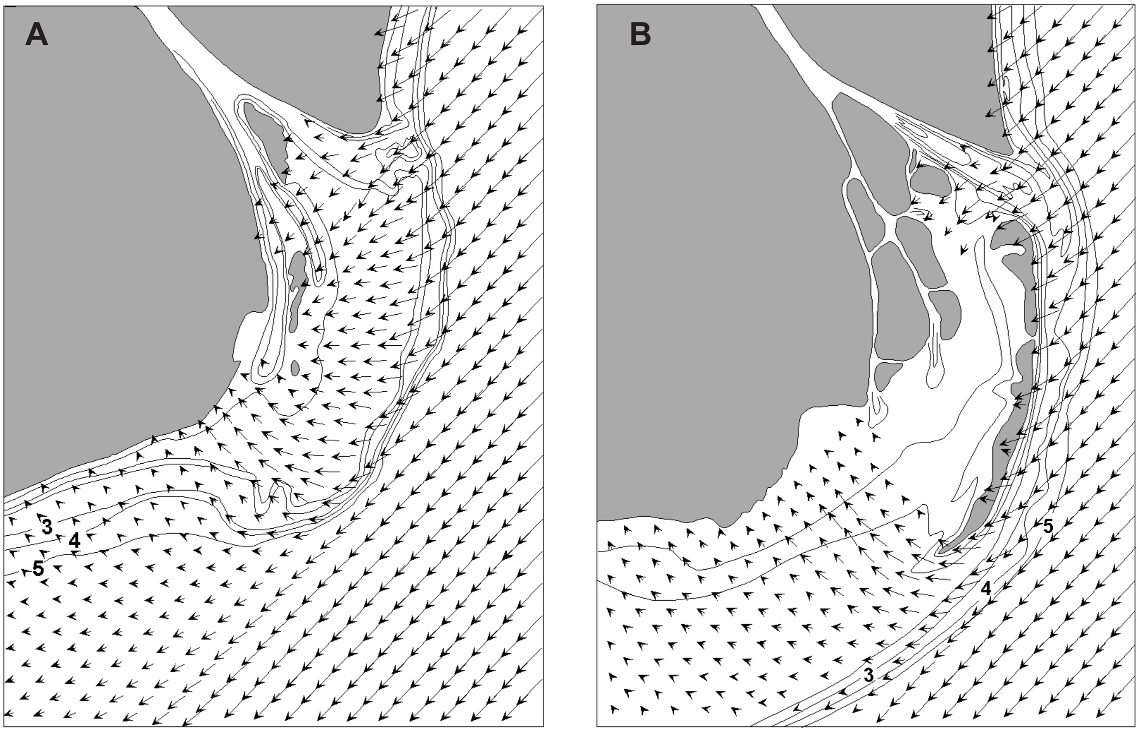

FIG. 8.—Wave transformation of a monochromatic northeasterly wave field of 1 m height (represented by the length of the wave vectors at the eastern side of the chart) and 4.5 s period that was propagated from deep water (> 30 m) on **A)** 1856 and **B)** 1935 bathymetries.

combined result of a greater sediment load and a shallow, ramp-like bathymetry at the initial Chilia mouth (e.g., Hartley, 1862, 1894–95; Diaconu and Mihailov, 1963c). However, the influence of the southward-directed coastal current and longshore transport can be observed in the initial development phase between 1830 and 1883, when the southernmost distributaries were deflected to the south. The roughly isometric shape of the lobe was achieved only after 1883, when a shallow bay left between the deflected part delta plain and the mainland was filled by a secondary bayhead delta.

Another interesting influence attributable to the longshore transport is the preferential progradation of the Ochakov and Old Stambul mouths, at the northern and southern extremities of the Chilia lobe, respectively. Longshore transport has been minimal at Ochakov, where the dominant waves are normal to the coast. Historically, the Old Stambul has discharged the most water and sediment of all secondary branches of Chilia (Diaconu and Mihailov, 1963a). Furthermore, after 1871 the branch has built its river-dominated secondary delta in the wave shadow zone of the updrift part of the Chilia lobe and behind the shallow subaqueous delta that had developed between the Chilia and Sulina jettied channels. In contrast, at mouths where the longshore transport rate is at a maximum because of the regional orientation of the coast relative to dominant waves (i.e., Codina, Monastery's, New Stambul, and Eastern), progradation was slower. The choking of secondary and tertiary distributary mouths by the sands brought in by the longshore transport has been proposed (Vidrasco 1924; Diaconu and Mihailov, 1963c), and this behavior suggests a downstream mechanism for preferential development of some branches and abandonment of others or even for avulsions in wave-dominated deltas.

Wave influence is also recorded in the morphology of the coasts adjacent to the Chilia lobe. Updrift of the lobe, the formation of spits that diverge offshore onto the subaqueous lobe platform are evident in 1856, 1871, 1922, and 2003. They formed from sand that was delivered by the southerly longshore sediment transport system (Barkovskaya, 1948; Zenkovich, 1956; Shuisky, 1984). Through amalgamation of these ridges, the Jebrieni beach-ridge plain has extended to the north, forced by the rapidly forming delta plain that advanced in the same direction. The downdrift coast adjacent to the Chilia lobe is sheltered from the dominant waves, but subordinate waves have also built offshore-diverging beach ridges (see shorelines in 1871 and 1883) that coalesced to form the Musura Cape. Since 1902, beach-ridge formation was no longer possible, because the Musura Cape was already incorporated into the Chilia lobe by the advance of the Old Stambul branch.

The progradation rate of the lobe decreased slowly after 1902, although the sediment discharged by the Chilia branch did not significantly change—this could be attributed to both the advance of the lobe in progressively deeper water and to the decrease in sediment delivered by the distributary per unit shoreline as the lobe perimeter progressively increased. In 1940, the first clear signs of erosion are apparent, especially on the central coast of the lobe between the Ochakov and Eastern secondary branches. By 2003, the Chilia has already become a wave-dominated lobe. The main secondary channels are evolving independently as wave-dominated secondary deltas. Because of a minimal net longshore transport at the mouth, the northern Ochakov branch is building a symmetric secondary delta with flying spits developing on both sides of the mouth. The other two branches that are important in terms of discharge,

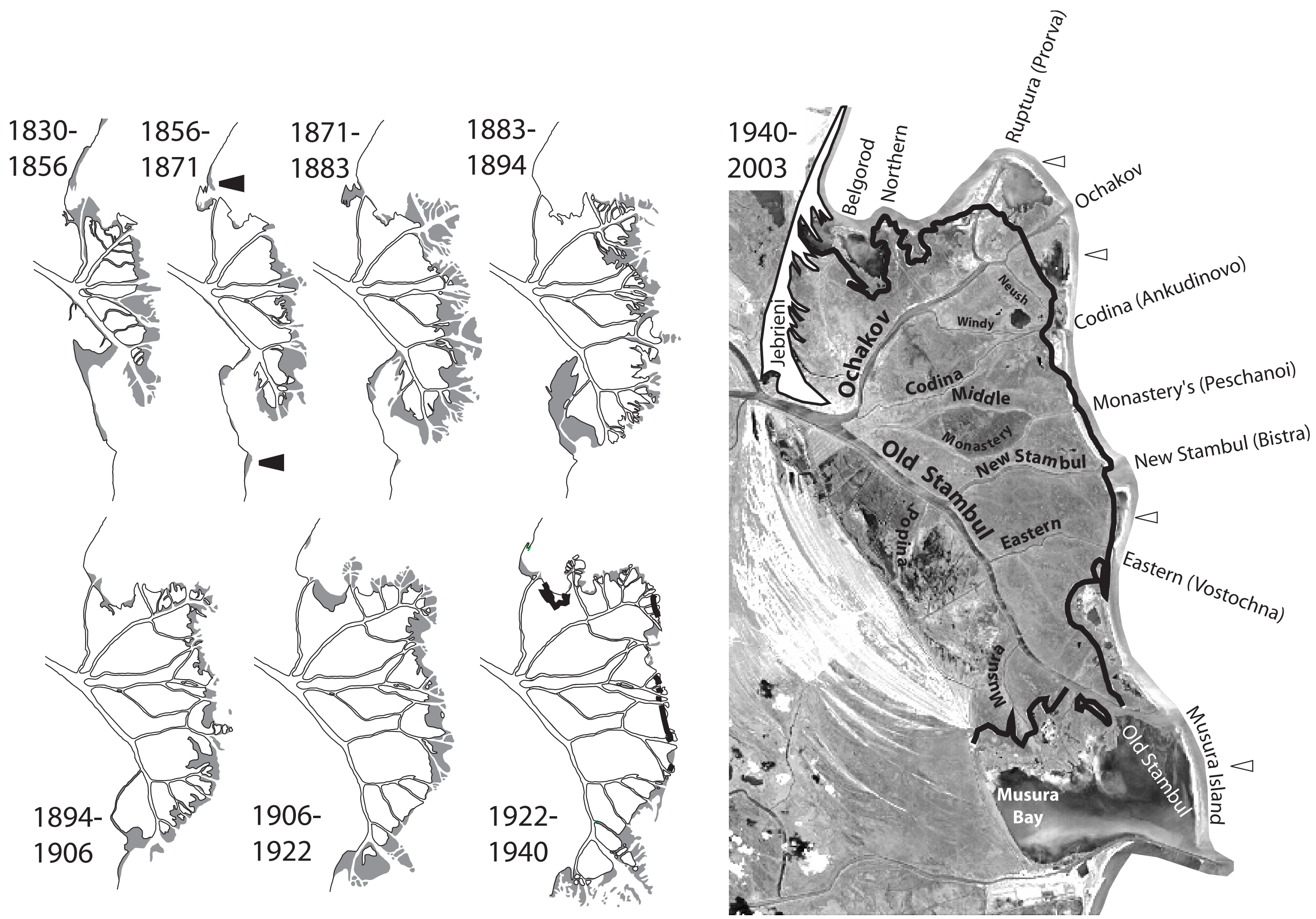

FIG. 9.—Delta-plain evolution of the Chilia III lobe between 1830 and 2003. Prograded sectors are in gray; erosional sectors in 1940 are indicated by the thick black line. The morphology of the lobe suggests a river-dominated regime until 1940. Beach-ridge development of the subaqueous delta platform north and south of the lobe are indicated by black filled arrows. Jebrieni beach-ridge plain is white-filled and marked on the ASTER satellite photo, where the 1940 coast is indicated by the black line. In 2003, the wave influence is felt strongly along the coast; note the flying barrier spits developed at the Ochakov mouth and barrier islands developed south of the New and Old Stambul mouths (indicated by unfilled arrows).

the New and Old Stambul, are building asymmetric secondary deltas with barrier islands developing downdrift of their mouths. A shallow subaqueous platform and a clear offshore offset of the updrift coast at the Old Stambul mouth are visible in 2003. Emergence of the Musura barrier island at this mouth after 1988 (Vespremeanu-Stroe, 2003) suggests that increased amounts of sand will be delivered by the longshore transport system toward the navigation channel at the Sulina mouth, in a natural experiment that will further test the drift-channeling hypothesis proposed herein. In 2003, the barrier was over 5 km in length and its southern tip almost touched the northern jetty of the Sulina channel.

EVOLUTION OF THE DANUBE DELTA

Pre-Holocene Geology

The Danube delta occupies a large, structurally controlled embayment along the southern margin of the East European Platform (Fig. 3). It overlies a portion of the North Dobrogean Orogen, the westernmost sector of a Cimmerian (Mesozoic) fold belt extending through Crimea into the Asian Cimmerides, and part of the Pre-Dobrogean Depression, a possible remnant of the former Cimmerian foredeep, characterized by a thick sequence of Middle Jurassic to Lower Cretaceous sediments covering the Caledonian–Hercynian basement of the Scythian Platform (Hippolyte, 2002, and references therein). The North Dobrogean Orogen is separated to the south from the Moesian Platform by the NW-trending Peceneaga–Camena crustal fault and it overthrusts to the north the sediments of the Pre-Dobrogean Depression along a NW-trending reverse fault running roughly along the modern course of the Sf. Gheorghe distributary of the Danube (Hippolyte, 2002, and references therein).

Sediment deposition on the Danube deep-sea fan should be a reliable indicator for the inception of the Danube drainage into the Black Sea. Winguth et al. (2000) correlated fan sequences to a composite oxygen isotope curve, a proxy for glacio-eustatic changes, and suggested that the fan started to develop ~ 900 ky BP. However, during lowstands, when global sea level fell below the depth of the sill connecting the Black Sea to the Mediterranean, the Black Sea became isolated, and its water level oscillated independently (e.g., Major et al., 2002). During isolation periods, the level in the Black Sea was controlled by the magnitude of river discharge, augmented at times by meltwater, as well as by direct precipitation and evaporation modulated by the regional climate. All these controls probably led to high-frequency abrupt lake-level cycles, much as in the Caspian Sea (e.g., Ryan et al., 1997; Major et al., 2002). Therefore, the date proposed by Winguth et al. (2000) should be considered a maximum estimate for the inception of the discharge of the Danube into the Black Sea.

Although hundreds of boreholes have been drilled in the Danube delta since the 1950s, there is little certainty about the stratigraphical architecture of the transgressive and highstand deltaic deposits preserved beneath the delta plain. Published interpretations of these cores (Liteanu and Pricajan, 1963) are based solely on lithologic descriptions combined with some faunal checks. Deltaic deposits are ~ 50 m thick on average and comprise two to three stacked coarsening-upward facies successions, 10 to 30 m thick, capped by levels of peat. This suggests preservation of several deltaic allomembers since the last lowstand. The deltaic facies overlie fluvial deposits, mostly gravels, that we interpret to represent alluvial valley fills of the last lowstand. However, in the pre-sequence stratigraphy era, on the basis of alternation of high-salinity vs. low-salinity fauna, Liteanu and Pricajan (1963) proposed a "layer-cake" stratigraphic model for the deltaic deposits, assigning a mid-Pleistocene age to the oldest preserved alluvial sediments. Their early model contradicts existing data on Pleistocene sea-level variations in the Black Sea (e.g., Chepalyga, 1984; Zubakov, 1988) and does not take in account the fact that relict low-salinity fauna is known to persist during marine highstands in some deltaic sub-environments (Spratt, 1860; Borcea, 1924).

Panin (1972) reinterpreted borehole data along the central axis of the delta plain and proposed that the deltaic lithosome consists entirely of transgressive and highstand deposits of Holocene age. Later, Panin et al. (1983) proposed that older highstand deposits are preserved below the easternmost part of the Holocene delta plain, on the basis of the presence of older reworked mollusks (see below). However, Ghenea and Mihailescu (1991) showed that loess and loessoid deposits, first described by Kühl and Hartley (1891), occur extensively at shallow subsurface depths in most of the northern half of the delta (Fig. 4D). Eolian deposition, including loess, was pervasive on the northern and northwestern shores and shelf of the Black Sea during glacial stages (Conea, 1969; Shcherbakov et al. 1978; Balescu et al., 2003). Deltaic deposits, if present below the loess horizon (Liteanu and Pricajan, 1963), are thus clearly older than Holocene. Preservation of loess in the northern sector of Danube delta suggests that the alluvial valley during the last lowstand was located along the present course of the Sf. Gheorghe branch and that the filling sequence for the Danube Bay as well as the growth of the first open-coast lobe (Sf. Gheorghe I) probably was controlled by the paleorelief developed during the last lowstand. Further work on subsurface stratigraphy and an independent chronology is required to elucidate the subsurface stratigraphic architecture of the deltaic deposits.

Holocene Development

Before the 1950s, research on the evolution of the Danube delta was limited mostly to the interpretation of external delta growth using land surveys (Antipa, 1915; Bratescu, 1922; Valsan, 1934; de Martonne, 1931, Slanar, 1945). The succession of lobe development had been hotly debated among early scholars, with much of the confusion resulting from the fact that some researchers postulated the development of beach-ridge plains downdrift from a fluvial feeder distributary rather than updrift of the mouth of the distributary (e.g., Bratescu, 1922) or considered them strictly marine strandplains (e.g., Valsan, 1934). De Martonne (1931) was the first to sketch a realistic sequence for the development of the external delta: he proposed that the Sf. Gheorghe built the first open-coast lobe, followed by the Sulina, followed by a new Sf. Gheorghe lobe, and later by the youngest Chilia lobe. Mihailescu (1936) presented support for this scenario by proposing that the updrift beach-ridge plains form by the obstruction of the longshore drift, by the river plume. Later, Zenkovich (1956) showed that the progradation of Chilia lobe blocked the longshore drift leading to amalgamation of beach ridges updrift of the lobe (Jebrieni beach ridge plain; Fig. 9), proposing a similar model for the older lobes of the delta. Panin and Panin (1969) and Panin (1974, 1989) used textural, mineralogical, and chemical composition of sediments to confirm the lobe development sequence proposed by de Martonne (1931) and Zenkovich (1956) by showing that most of the Caraorman and the early Letea beach-ridge plain consists of sand transported from the north with little or no contribution from their corresponding feeder distributaries. Morphodynamic considerations require a continuous, and direct, rather than embayed, nearshore zone to have existed between the Bugeac and the first Sf. Gheorghe mouth for allochthonous sediment to be transported from north of the Danube bay to form the updrift wing of the Sf. Gheorghe lobe without being trapped in the former Danube bay. This condition is not fulfilled without a baymouth barrier straddling the former Danube bay; its development was probably favored, and its position controlled, by the predeltaic loess relief, which ramps up toward the continent (Fig. 4D ; Ghenea and Mihailescu, 199).

Much of the early effort in trying to establish a chronology for delta development was spent in trying to reconstruct the ancient geography of the delta as described by Greek and Roman authors. The only modern chronology available is based on the conventional radiocarbon dating of mostly mollusk shells (Fig. 10; Panin et al., 1983). There is a recognized difficulty of dating deltaic formations because age inversions are common (see review in Stanley, 2001). In the case of the Danube delta, although the dating technique used in developing the chronology was valid (Noakes and Herz, 1983), the method of collecting datable material and the subsequent interpretation of dates makes the chronology uncertain. Most shells were collected at depths between 1 and 3 m (Noakes and Herz, 1983; Panin et al., 1983), but no lithostratigraphic and facies description is provided for the collection sites, which makes it difficult to assess what deltaic subenvironments have actually been dated and if reworking or condensed intervals were likely to affect sampling. The large quantity of carbonate material needed for dating probably precluded the use of single shells that could be ascertained to be *in situ*; indeed, all shells dated showed signs of reworking (Panin et al., 1983).

A number of samples dated by Panin et al. (1983) are older than Holocene: the age of multispecies mollusk samples as well as some monospecific samples of *Paphia* (*Tapes*) *senescens*, *Ostrea edulis*, *Cerithium* (*Thericium) vulgatum* and *Chlamys* (*Flexopecten*) *glabra* was attributed by Panin et al. (1983) to inclusion of specimens from older marine episodes in the dated samples. However, with the exception of *Paphia senescens*, the other species cannot be discounted in the chronology because they also lived in the Black Sea during the Holocene (e.g., Neveskaya, 1965). During the last glacial interval, the Black Sea was isolated and sustained a freshwater Caspian-type fauna rather than a marine one (see Ryan et al., 1997, and references therein). The salinization threshold for establishment of a marine fauna did not occur in the Black Sea before 8,400–8,500 ^{14}C years BP (Scherbakov and Babak, 1979;

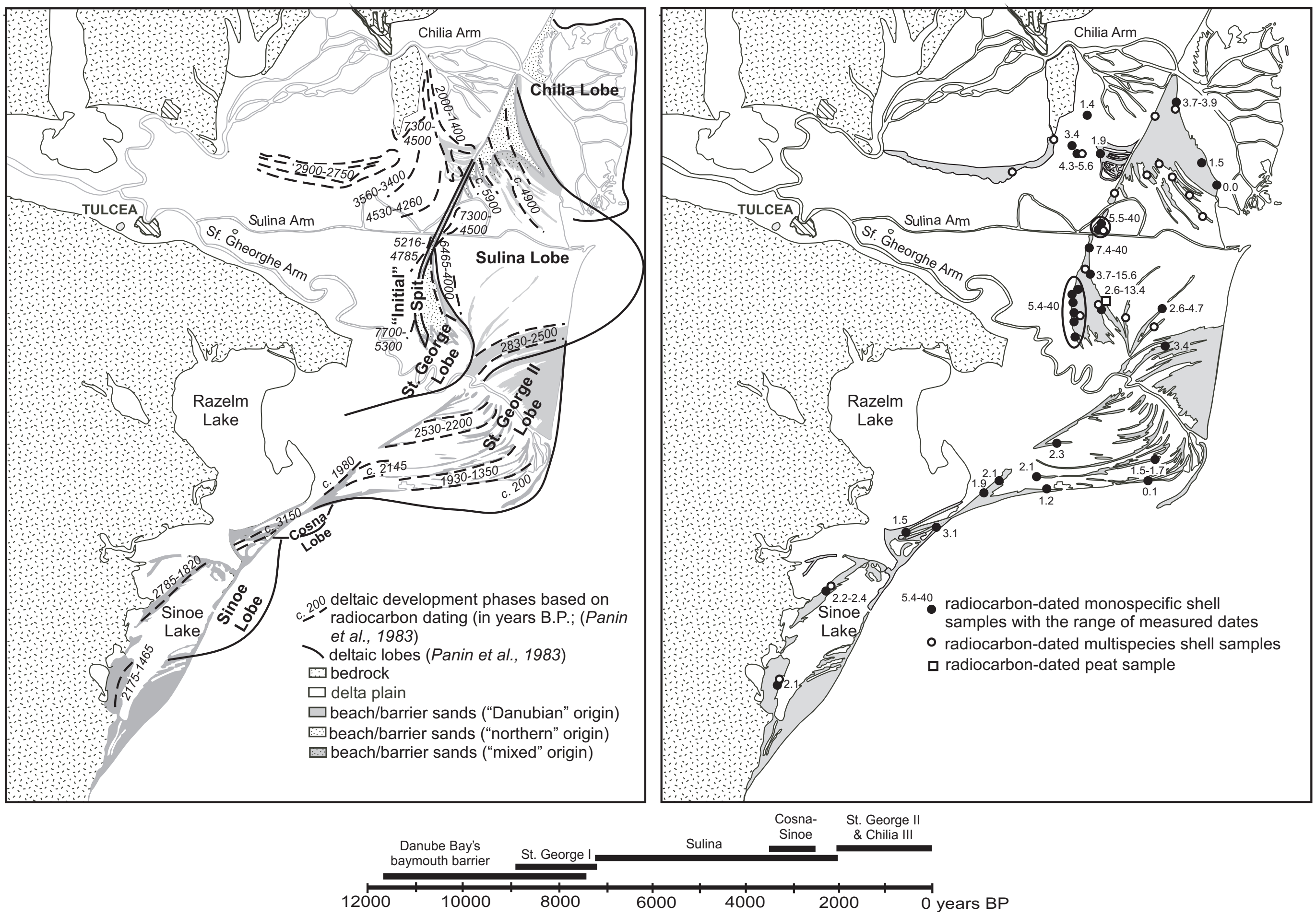

Fig. 10.—Danube delta evolution model (left side; after Panin et al., 1983) and location for radiocarbon dates used in that model (right side; Noakes and Hertz, 1983). Range of dates are specified for monospecific samples, but not for multispecies samples. The chronology of lobe development (lower side) for the external delta is from Panin et al. (1983).

Ross and Degens, 1974; see also discussions in Ryan et al., 1997, and Major et al., 2001); however, the age spread for dated samples of marine mollusks extends much earlier than that. Barring any systematic local effects on the radiocarbon content of the dissolved inorganic carbon, on the metabolic carbon used by the mollusks, or diagenetic postdepositional changes in shell chemistry (see Stanley, 2001, and references therein), the reasonable explanation for this discrepancy is the one proposed by Panin et al. (1983): inclusion of shells of marine species from older marine intervals in the dated samples, altering their age toward values older than their time of deposition. Furthermore, by dating multiple shells in a single sample, there is no reason to believe that Holocene samples showing ages younger than the salinity threshold have not also been affected by a similar "aging" effect.

The northern wing of the Sf. Gheorghe I lobe, which was the first to prograde outside of the Danube bay baymouth barrier, is composed of ridges that are higher than the present Black Sea level, even in places that have not been affected by dune construction. As discussed previously, the allochthonous composition of these ridges requires a baymouth barrier extending from the Bugeac to the Sf. Gheorghe mouth, before the first deltaic lobe began to prograde. The baymouth barrier that closed the former Danube bay was interpreted to have formed between 11,700 and 9,800 ^{14}C years BP in its central part and between 10,700 and 7,500 ^{14}C years BP in its southern part (Fig. 10; Panin et al., 1983). This argues for a sea level close to the present one no later than 9,800 years ago. However, although it has been suggested that the Black Sea level was between –20 and –40 m at that time (see discussions in, e.g., Aksu et al., 2002a, and Aksu et al., 2002b, and Ryan et al., 2003), compared to the ~ –50 m of world ocean (Fairbanks, 1989), to our knowledge no other reliable data exist to support a level as high as today.

Because of the progradational character of a wave-dominated delta lobe, beach-ridge-plain beach ridges become younger in the offshore direction as the mouth of the distributary advances; once a ridge is stranded by the next forming ridge, it can no longer sustain a marine fauna. The youngest sample of the dated marine shell mixtures from a ridge should be the best representation of the stranding time; taking in account the possible "aging effect" when measuring mixtures, that date is also the oldest possible age for the stranding. If we apply this line of reasoning to the western edge of the Caraorman beach ridge plain, the inception of progradation for the first lobe of Sf. Gheorghe, and therefore of the entire external delta plain, could not be much older than ~ 5,500-6,000 ^{14}C years BP, the youngest ages reported for marine assemblage samples collected there (Noakes and Hertz, 1983). This age is consistent with a Black Sea that was already connected to the world ocean, suggesting a sea level similar to the modern one and comparable to the level estimated for the neighboring northern Aegean Sea at that time (Lambeck, 1995).

For the more recent lobes of the external delta, the "aging" effect is evident from the wide range of estimated ages at locations where several samples have been dated, as well as from age reversions between beach ridges (e.g., Sulina lobe and Razelm–Sinoe baymouth barriers; Fig. 10). However, it is apparent that all these arguments remain speculative until a new chronology becomes available. Integration of lithostratigraphic and facies information at collection sites with accurate dating is clearly needed to provide a plausible scenario for the development of such a complex environment as the Holocene Danube delta.

Panin (1983) and Panin et al. (1983) proposed that another lobe, the Cosna, was built by the Dunavatz branch between 3,550 and 2,550 ^{14}C years BP in front of the Razelm–Sinoe lagoon system (Fig 10). Later, the Dunavatz moved (avulsed?) south to build another younger lobe, the Sinoe (Fig 10; Panin, 1983). These lobes were apparently reworked into the present system of baymouth barriers, segmenting and closing the Razelm–Sinoe lagoons (Panin, 1983). Such a hypothesis is hard to envision because there are no connecting channel-levee deposits and other subaerial deltaic deposits between the proposed advanced position of the lobes and the actual inland location of the Dunavatz branch, crossing over the Razelm–Sinoe bay. Erosion of such deposits is improbable because the bay was a sheltered environment, protected by the Cosna and Sinoe lobes or their reworked counterparts as well as by the updrift deltaic plain. Only rapid subsidence, strictly localized to the Razelm–Sinoe area, could have removed the traces of the proposed Dunavatz channels.

Instead, a recent core acquired by us for this study from the Zmeica barrier, the landwardmost baymouth barrier in the lagoon, does not show localized subsidence for the Razelm–Sinoe lagoon. The core consists of sandy beach and overwash deposits intercalated with organic-rich layers that were dated between to 4,700 and 4,100 AMS ^{14}C years BP. (Dates are not corrected for reservoir age and are not calibrated to calendar years to be consistent with the other radiocarbon dates reported from Danube delta. Samples are NOSAMS Lab Number OS-44312 dated at 4,100 ± 40 AMS ^{14}C years and NOSAMS Lab Number OS-44165 dated at 4,760 ± 40 AMS ^{14}C years.) However, there is no need to invoke the presence of initial deltaic deposition at the distal end of the Razelm–Sinoe bay because the abundance of sand from the updrift deltaic lobes and high net longshore transport rate to the south combined with the rollover behavior of barriers have been favorable all along to the development of successive baymouth barriers downdrift of the Holocene delta (see previous discussion).

CONCLUDING REMARKS

Morphodynamics addresses the coupled adjustments among hydrodynamic processes, sediment transport, sedimentation, and morphology, and has become an established paradigm for studying coastal evolution (e.g., Carter and Woodroofe, 1994; Wright, 1995; Short, 1999). Compared to other clastic coasts, the morphodynamics of deltas is complicated by the continuous and/or episodic delivery of freshwater and sediment to the coast by one or several rivers through one or multiple mouths. The solid load discharged by the river comprises both suspended fine sediment in the river plume and coarse bed-load. Although our understanding of morphodynamic processes related to sedimentation processes related to plume development has made important gains over the last several years with the advent of programs like STRATAFORM, EUROSTRATAFORM, and EURODELTA (see e.g., Nittrouer, 1999, Syvitski and Trincardi, in press, and papers therein), morphodynamic studies of the coarse, bed-load fraction is seriously lagging behind. However, sedimentation of the coarse-grade sediments appears to be a dominant factor in the evolution of river mouths (Wright, 1995) and further, in determining the facies architecture of wave-dominated deltas as well as the distribution of reservoir-quality lithosomes (Bhattacharya and Giosan, 2003).

Our present analysis of the Danube delta development reevaluates the knowledge about a long-studied but still poorly understood delta, while introducing several novel morphodynamic aspects of river-mouth and deltaic deposition in wave-influenced environments that will provide a basis for further quantitative field and modeling studies:

Once a delta progrades to the open coast, it forms a discrete subaqueous protuberance. In many cases, the protuberance is also expressed at the shoreline, which progrades relative to the adjacent coasts. From a hydrodynamical perspective, this deltaic "bulge" can be viewed as a morphological perturbation to the regional circulation system. Morphodynamic adjustments between the deltaic morphology and the fluvial and basinal hydrodynamics are most intense at the river mouths. Morphodynamics of wave-dominated deltaic distributary mouths is highly nonlinear, involving multiple feedbacks between subaerial deltaic progradation, deposition on the subaqueous delta, current and wave hydrodynamics, and wave–current interactions. In spite of this complexity, the morphology at the mouth exhibits a tendency to self-organize that is reflected and preserved in a coherent series of stratigraphic architectural styles (Wright, 1977; Dominguez, 1996; Bhattacharya and Giosan, 2003).

Morphodynamic models (e.g., Komar, 1973; Cowell et al., 2003) identify the rotation of the delta shoreline as a phenomenon responsible for changes in the longshore drift along the deltaic coast. Assuming that the offshore wave regime is not significantly skewed to one side of the regional orientation of the coast, as a delta progrades, the angle of wave attack increases along both sides of the mouth, increasing the longshore transport away from the mouth. Changes in the drift magnitude can be thought as a feedback loop between subaerial morphology of the delta and waves, which ultimately limits the growth of the deltaic protrusion. For purposes of numerical modeling, rotation is generally conceptualized as a two-dimensional problem by assuming that the nearshore profile is invariable for the entire deltaic coast. However, where the offshore wave direction is skewed toward one side of the regional orientation of the coast, the downdrift side is sheltered from dominant waves and longshore transport converges at the mouth.

The river plume interacts with the surface waves to increase the convergence of the longshore sediment transport updrift of the river mouth (i.e., the hydraulic-groin effect of the plume of Todd, 1968 and Komar, 1973). The groin effect occurs regardless of the presence of suspended sediment in the plume. However, the increase in the density of the plumes by addition of suspended sediment leads to an increase in the density contrast between the plume and the coastal waters (e.g., Rodriguez et al., 2000), resulting in wave-energy dissipation when waves encounter the plume. Where the longshore drift reaching the mouth is significant, a feedback loop develops between the relatively rapid progradation of the updrift coast caused by the hydraulic-groin effect and/or by the development of a subaqueous delta, leading to a mouth that is progressively more sheltered from waves, which in turn, results in an increase in sedimentation on the subaqueous delta platform.

On the updrift side, the longshore transport obstructed by the river plume delivers sand to build a deltaic beach-ridge plain comprising mostly allochthonous sediments. However, the subaqueous delta develops mostly on the opposite, downdrift side of the mouth, activating another positive feedback between morphology and wave hydrodynamics. Waves reaching the downdrift coast dissipate their energy more effectively over the subaqueous delta compared to the coast updrift of the mouth where no subaqueous delta platform is present. This results in more quiescent conditions downdrift of the mouth, favoring the expansion of the subaqueous delta. Development of a shallow subaqueous delta platform allows deposition of mouth bars to form offshore at the edge of the platform during floods. Waves can then rework these bars into barrier islands fronting the delta platform.

As the emergent barrier becomes the new shoreline for the downdrift half of the delta, it produces a response of the delta to the wave hydrodynamics. Wave energy is no longer dissipated over the shallow delta platform; moreover, they reach the barrier shore incompletely refracted, resulting in a channeling of the longshore transport (i.e., intensification of the transport guided along the shore of the barrier). This in turn leads to a more rapid development of the subaqueous delta in the alongshore direction.

As much as the river plume, wave-driven longshore transport transmits morphological signals from the deltaic coast to adjacent nondeltaic coastal compartments or from one deltaic lobe to another. Sheltering from waves by the subaerial and subaqueous lobe and development of barrier-spit "wings" are some of the more obvious features occurring downdrift of a delta. Sediment derived from the river could also impose a certain heterogeneity in the nearshore (e.g., forced accumulation of muds in an otherwise energetic environment), leading to a strong coupling between hydrodynamic and morphodynamic processes in regions far from the river mouth, in contrast to a homogeneous systems (Sheremet and Stone, 2003).

In recent 3-D modeling efforts, that include waves as a forcing factor (Overeem et al., this volume), offsets between the subaqueous depocenters and their delta-plain counterparts are clearly developed. However, the simulated response of a river mouth to a strong longshore drift is invariably a deflection downdrift, without updrift–downdrift offsets or the emergence of barrier islands. Three-dimensional sedimentation models that include wave–current interactions, a movable heterogeneous bed, and separate bedload and plume dynamics modules are required to begin to address the complexity of the morphodynamics at wave-dominated river mouths.

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REFERENCES

AKSU, A.E, HISCOTT, R.N., MUDIE, P.J., ROCHON, A., KAMINSKI, M.A., ABRAJANO, T., AND YASAR, D., 2002, Persistent Holocene outflow from the Black Sea to the eastern Mediterranean contradicts Noah's flood hypothesis: GSA Today, v. 12, p. 4–10.

AKSU, A.E., HISCOTT, R.N., YASAR, D., ILER, F.I. AND MARSH, S., 2002, Seismic stratigraphy of Late Quaternary deposits from the southwestern Black Sea shelf: evidence for non-catastrophic variations in sea-level during the last ~ 10000 yr: Marine Geology, v. 190, p. 61–94.

ANTIPA, G., 1915, Wissenschaftliche und wirtschaftliche Probleme des Donaudeltas. Inst. Geol. Rom. An., v. 7, 88 p.

ASLAN, A., WHITE, W.A, WARNE, A.G., AND GUEVARA, E.H., 2003, Holocene evolution of the western Orinoco Delta, Venezuela, Geological Society of America, Bulletin, v. 115, p. 479–498.

BACESCU, M.C., MULLER, G.I., AND GOMOIU, M.-T., 1971, Ecologie Marina, Fourth Edition: Academiei Romane, p. __.

BALESCU, S., LAMOTHE, M., MERCIER, N., HUOT, S., BALTEANU, D., BILLARD, A., AND HUS, J., 2003, Luminescence chronology of Pleistocene loess deposits from Romania: testing methods of age correction for anomalous fading in alkali feldspars: Quaternary Science Reviews, v. 22, p. 967–973.

BANU A.C., AND RUDESCU, L., 1965, Delta Dunãrii: Bucharest, Ed. Stiintifica, 295 p.

BARKOVSKAYA, M.G., 1948, Doklad na okeanograficheskom soveshchanii: Sevastopol, v. _, p. _.

BHATTACHARYA, J.P., AND GIOSAN, L., 2003, Wave-influenced deltas: geomorphologic implications for facies reconstruction, Sedimentology, v. 50, p. 187–210.

BHATTACHARYA, J.P., AND WALKER, R.G., 1992, Deltas, *in* Walker, R.G., and James, N.P., eds., Facies Models: Response to Sea Level Change, Geological Association of Canada, St. Johns, p. 157–177.

BONDAR, C., 1964, Date asupra distributiei spatiale a salinitatii in jurul gurilor Dunarii: Studii de Hidraulica, v. 11, p. 229–245.

BONDAR, _., AND BLENDEA, _., 2000, Water and sediment transport by the Danube into the Black Sea during 1840-1995, IOC/UNESCO, IOC Workshop Report no. 145, Paris, p. 58–63.

BONDAR, C., AND HARABAGIU, E., 1992, Regimul depunerilor de aluviuni la gura canalului Sulina, Studii de Hidraulica, v. 33, p. 131–141.

BONDAR, C., STATE, I., CERNEA, D., AND HARABAGIU, E., 1992, Scurgerea de apa si aluviuni a Dunarii la gurile bratelor Dunarii in anii 1858-1988: Studii de Hidraulica, v. 33, p. 57–71 (in Romanian).

BORCEA, I., 1924, Faune survivante de type caspien dans la limans d'eau douce de Roumanie: Universite de Iassy, Anales scientifique, no. _, p. 207–232.

BRATESCU, C., 1922, Delta Dunãrii. Geneza si evolutia sa morfologicã si cronologicã: Soc. Reg. Rom. Geogr., Bulletin, v. 41, p. 3–39.

BULGAKOV, N.P., GOLUBEV, Y.N., AND KUFTARKOV, A.Y., 1992, Energy active zones of the Black Sea: Rossiskaya Akademiya Nauk, Doklady, v. 326, p. 177–180.

CARTER, R.W.G., AND WOODROFFE, C.D., eds., 1984, Coastal Evolution; Late Quaternary Shoreline Morphodynamics: Cambridge, U.K., Cambridge University Press, 539 p.

CHEPALYGA, A.L., 1984, Inland sea basins, *in* Velichko, A.A., Wright, H.E., and Barnosky, C.W., eds., Late Quaternary Environments of the Soviet Union: University of Minnesota, _ p.

CIULACHE, S., 1993, The wind on the Romanian shore of the Black Sea, University of Bucharest, geografie, Analele, p. 1–10.

COLEMAN, J.M., AND WRIGHT, L.D., 1975, Modern river deltas: variability of processes and sand bodies, *in* Broussard, M.L., ed., Deltas, Models for Exploration: Houston Geological Society, p. 99–149.

CONEA, A., 1969, Profils de loess en Roumanie, *in* Fink, J., ed., La stratigraphie des loess d'Europe: Association française pour l'étude du Quaternaire, Bulletin, Suppl. INQUA, p. 127–134.

COWELL, P.J., STIVE, M.J.F., NIEDORODA, A.W., DE VRIEND, H.J., SWIFT, D.J.P., KAMINSKY, G.-M., AND CAPOBIANCO, M., 2004, The coastal-tract (Part 1): A conceptual approach to aggregated modeling of low-order coastal change: Journal of Coastal Research, v. 19, p. 812–827.

DIACONU, C., AND MIHAILOV, V.N., 1963a, Scurgerea apei, *in* Diaconu, C., and Nichiforov, I.D., eds., Zona de varsare a Dunarii: Bucharest, Editura Tehnica, p. 135–176.

DIACONU, C., AND MIHAILOV, V.N., 1963b, Scurgerea aluviunilor, *in* Diaconu, C., and Nichiforov, I.D., eds., Zona de varsare a Dunarii: Bucharest, Editura Tehnica, p. 177–191.

Diaconu, C., and Mihailov, V.N., 1963c, Procesele formarii deltei in etapa actuala, *in* Diaconu, C., and Nichiforov, I.D., eds., Zona de varsare a Dunarii: Bucharest, Editura Tehnica, p. 192-222.

Dominguez, J.M.L., Bittencourt, A.C.S.P., and Martin, L., 1983, O papel da deriva litoranea de sedimentos arenosos na construcao das planicies costeiras associadas as desembocaduras dos rios Sao Francisco, Jequitinhonha, Doce e Paraiba do Sul: Rev. Brasil. Geocienc., v. 13, p. 98–105.

Dominguez, J.M.L., Martin, L., and Bittencourt, A.C.S.P., 1987, Sea-level history and Quaternary evolution of river mouth-associated beach-ridge plains along the east- southeast Brazilian Coast: a summary, *in* Nummedal, D., Pilkey, O.H., and Howard, J.D., eds., Sea-Level Fluctuations and Coastal Evolution, SEPM, Special Publication 41, p. 115–127.

Dominguez, J.M.L., 1996, The São Francisco strandplain: a paradigm for wave-dominated deltas?, *in* De Baptist, M., and Jacobs, P., eds., Geology of Siliciclastic Shelf Seas: Geological Society of London, Special Publication, 117, p. 217–231.

Fairbanks, R.G., 1989, A 17000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation: Nature, v. 342, p. 637–642.

FitzGerald, D.M., 1984, Interactions between the ebb-tidal delta and landward shoreline: Price Inlet, South Carolina: Journal of Sedimentary Petrology, v. 54, p. 1303–1318.

Geyer, W.R, Hill, P.S., and Kineke, G.C., 2004, The transport, transformation and dispersal of sediment by buoyant coastal flows: Continental Shelf Research, v. 24, p. 927–949.

Ghenea, C., and Mihailescu, N., 1991, Palaeogeography of the lower Danube Valley and Danube Delta during the last 15 000 years, *in* Starkel, L., Gregory, K.J., and Thornes, J.B., eds., Temperate Palaeohydrology; Fluvial Processes in the Temperate Zone During the Last 15 000 Years, New York, Wiley, _ p.

Giosan, L., 1993, Studiul evolutiei dinamice a sedimentelor in zona litorala din fata Deltei Dunarii: unpublished Diploma thesis, Bucharest University, 75 p.

Giosan, L., 1998, Long term sediment dynamics on Danube delta coast, 1998, *in* Dronkers, J., and Scheffers, M.B.A.M., eds., Physics of Estuaries and Coastal Seas: ______, p. 365–376.

Giosan, L., Bokuniewicz, H., Panin, N., and Postolache, I., 1999, Longshore sediment transport pattern along the Romanian Danube delta coast, Journal of Coastal Research, v. 15, p. 859–871.

Hartley, C.A., Sir, 1862, Description of the Delta of the Danube, and of the Works, recently executed, at the Sulina Mouth: Institution of Civil Engineers, Minutes of Proceedings, v. 21, p. 277–308.

Hartley, C.A., Sir, 1873, On the Changes that have recently taken place along the Sea Coast of the Delta of the Danube, and on the Consolidation of the Provisional Works at the Sulina Mouth: Institution of Civil Engineers, Minutes of Proceedings, v. 36, p. 201–253.

Hartley, C.A., Sir, and Harcourt, L.V., 1894–95, The Survey of the Delta of the Danube in 1894: Institution of Civil Engineers, Minutes of Proceedings, v. 122, p. 336–342.

Hippolyte, J.-C., 2002, Geodynamics of Dobrogea (Romania): new constraints on the evolution of the Tornquist–Teisseyre Line, the Black Sea and the Carpathians: Tectonophysics, v. 357, p. 33–53.

Humborg, C., Ittekkot, V., Cociasu, A. von Bodungen, B., and Popa, L., 1997, Effect of Danube River dam on Black Sea biogeochemistry and ecosystem structure: Nature, v. 386, p. 385–388.

Jiménez, J.A., Sanchez-Arcilla, A., Valdemoro, H.I., Gracia, V., and Nieto, F., 1997, Processes reshaping the Ebro delta: Marine Geology, v. 144, p. 59–79.

Kana, T.W., 1996, Signatures of coastal change at mesoscale, *in* Dally, W.R., and Ziedler, R.B., eds., Coastal Dynamics '95: American Society of Civil Engineers, p. 987–997.

Komar, P.D., 1973, Computer models of delta growth due to sediment input from rivers and longshore transport: Geological Society of America, Bulletin, v. 84, p. 2217–2226.

Komar, P.D., 1998, The modeling of processes and morphology in the coastal zone—reflections on the maturity of our science: Shore and Beach, v. 66, p. 10–22.

Korotaev, G., Oguz, T., Nikiforov, A., and Koblinsky, C., 2003, Seasonal, interannual, and mesoscale variability of the Black Sea upper layer circulation derived from altimeter data, Journal of Geophysical Research, 108, 3122, doi:10.1029/2002JC001508.

Kühl, C.H.L., and Hartley, C.A., Sir, 1891, On the Sulina Branch of the Danube: Institution of Civil Engineers, Minutes of Proceedings, v. 106, p. 238–247.

Lambeck, K., 1995, Late Pleistocene and Holocene sea-level change in Greece and south-west Turkey: a separation of eustatic, isostatic and tectonic contributions: Geophysical Journal International, v. 122, p. 1022–1044.

Lancelot, C., Martin, J.-M. Panin, N., and Zaitsev, Y., 2002, The north-western Black Sea: A pilot site to understand the complex interaction between human activities and the coastal environment: Estuarine, Coastal and Shelf Science, v. 54, p. 279–283

Liteanu, E., and Pricãjan, A., 1963, Alcãtuirea geologicã a Deltei Dunãrii: Hidrobiologia, v. 4, p. 57–82.

Major, C., Ryan, W.B.F., Lericolais, G., and Hajdas, I., 2002, Constraints on Black Sea outflow to the Sea of Marmara during the last glacial–interglacial transition: Marine Geology, v. 190, p. 19–34.

Makaske, B., 2001, Anastomosing rivers: a review of their classification, origin and sedimentary products, Earth-Science Reviews, v. 53, p. 149–196.

Malciu, V., and Diaconu, V., 2001, Long-term trend of the sea level at the Romanian littoral: GLOSS national report 2001, *http://www.pol.ac.uk/psmsl/*.

Margesson, R., 1997, Environment and international water management: Dealing with the problems of the Danube delta: Env. Impact Asses. Rev., v. 17, p. 145–162.

Martonne, E., de, 1931, Europe Centrale, vol. II: Paris, v. __, p. 786–787.

Meade, R.H., 1996, River-sediment inputs to major deltas, *in* Milliman, J.D., and Haq, B.U., eds., Sea-Level Rise and Coastal Subsidence—Causes, Consequences, and Strategies, Dordrecht, The Netherlands, Kluwer, p. 63–85.

Mihailescu, V., 1947, Geografia Romaniei: Bucuresti, _____, (in Romanian).

Milliman, J.D., and Meade, R.H., 1983, World-wide delivery of river sediment to the oceans: Journal of Geology, v. 91, p. 1–21.

Munteanu, I., 1996, Soils of the Romanian Danube Delta Biosphere Reserve: RIZA 96-070, 171 p.

Murgoci, G., 1912, Etudes geologiques dans la Dobrogea de Nord: Inst. Geol. Rom., An., v. 6, p. ___.

Nittrouer, C.A., 1999, STRATAFORM: overview of its design and synthesis of its results: Marine Geology, v. 154, p. 3–12.

Noakes, J.E., and Herz, N., 1983, University of Georgia radiocarbon dates VII: Radiocarbon, v. 25, p. 919–929.

Oguz, T., Latun, , Latif, M.A., Vladimirov, V.V., Sur, H.I., Markov, A.A., Özsoy, E., Kotovshchikov, B.B., Eremeev, V.V., and Ünlüata, Ü., 1993, Circulation in the surface and intermediate layers of the Black Sea: Deep-Sea Research I, v. 40, p. 1597–1612.

Özsoy, E., and Ünlüata, Ü., 1997, Oceanography of the Black Sea: A review of some recent results, Earth-Science Reviews, v. 42, p. 231–272.

Panin, N., 1972, Histoire Quaternaire du Delta du Danube. Essai d'interpretation des facies des depots deltaiques: Cercetari Marine, v. 4, p. 5–15.

Panin, N., 1974, Evolutia Deltei Dunãrii în timpul Holocenului: Institutul Geologic-Studii Tehnnice si Economice, v. 5, p. 107–121.

Panin, N., 1983, Black Sea coast line changes in the last 10,000 years. A new attempt at identifying the Danube mouth as described by the ancients: Dacia, v. 27, p. 175–184.

Panin, N., 1989, Danube delta. Genesis, evolution and sedimentology: Rev. Roum. Géol. Géophys. Géogr., v. 33, p. 25–36.

Panin, N., 2003, The Danube Delta Geomorphology and Holocene Evolution: a Synthesis: Géomorphologie : Relief, Processus, Environnement, v. 4, p. 247–262.

Panin, N., and Jipa, D., 2002, Danube River sediment input and its interaction with the north-western Black Sea: Estuarine, Coastal and Shelf Science, v. 54, p. 551–562.

Panin, N., and Panin, S., 1969, Sur la genese des accumulations des mineraux lourds dans le delta du Danube, Revue de Geographie Physique et de Geologie Dynamique, v. 11, p. 511–522.

Panin, N., Panin, S., Herz, N., and Noakes, J.E., 1983, Radiocarbon dating of Danube delta deposits: Quaternary Research, v. 19, p. 249–255.

Panin, N. et al., 1986, Harta sedimentologica a platoului continental al Marii Negre de la Sulina la Tulcea: Romanian Institute of Geology and Geophysics, p. __.

Penland, S., and Suter, J.R., 1989, The geomorphology of the Mississippi River chenier plain: Marine Geology, v. 90, p. 231–240.

Popescu, I., Lericolais, G., Panin, N., Normand, A., Dinu, C., and Le Drezen, E., 2004, The Danube subaqueous canyon (Black Sea): morphology and sedimentary processes: Marine Geology, v. 206, p. 249–265.

Popp, N., 1961, Caracterizarea litologica a paminturilor Deltei Dunarii pe baza datelor de foraj: Meteorologia, hidrologia si gospodarirea apelor, v. 4, p. 126–139.

Pranzini, E., 2001, Updrift river mouth migration on cuspate deltas: two examples from the coast of Tuscany (Italy): Geomorphology, v. 38, p. 125–132.

RCMGG (Romanian Center for Marine Geology and Geoecology), 1994, Monitoringul geo-ecologic al sistemului Dunare-Delta-Litoral-Marea Neagra: Centrul Roman de Geologie and Geoecologie Marina, Bucharest, 46 p.

Reschke, S., Ittekkot, V., and Panin, N., 2002, The nature of organic matter in the Danube River particles and north-western Black Sea sediments: Estuarine, Coastal and Shelf Science, v. 54, p. 563–574.

Rimbu, N., Boroneant, C., Buta, C., and Dima, M., 2002, Decadal variability of the Danube river streamflow in the lower basin and its relation with the North Atlantic Oscillation: International Journal of Climatology, v. 22, p. 1169–1179.

Rodriguez, A.B., Hamilton, M.D., and Anderson, J.B., 2000, Facies and evolution of the modern Brazos Delta, Texas; wave versus flood influence: Journal of Sedimentary Research, v. 70, p. 283–295.

Ross, D.A., and Degens, E.T., 1974. Recent sediments of the Black Sea, *in* Degens, E.T., and Ross, D.A., eds., The Black Sea: Geology, Chemistry, and Biology: American Association of Petroleum Geologists, p. 183–199.

Rossetti, C., and Rey, F., 1931, La Commission Européenne du Danube et son oeuvre de 1856 a 1,931: Paris, Imprimerie Nationale, ___ p.

Ryan, W.B.F., Pitman, W.C., Major, C.O., Shimkus, K., Moskalenko, V., Jones, G.A., Dimitrov, P., Görür, N., Sakinç, M., and Yüce, H., 1997, An abrupt drowning of the Black Sea: Marine Geology, v. 138, p. 119–126.

Shcherbakov, F.A., and Babak, Y.V., 1979, Stratigraphic subdivision of the Neoeuxinian deposits in the Black Sea: Oceanology, v. 19, p. 298–300.

Shcherbakov, F.A., et al., 1978, Sedimentation on the Continental Shelf of the Black Sea: Moscow, Nauka, 211 p.

Sheremet, A., and Stone, G.W., 2003, Observations of nearshore wave dissipation over muddy sea beds, Journal of Geophysical Research, 108, 3357, doi:10.1029/2003JC001885.

Short, A.D., 1999, Handbook of Beach and Shoreface Morphodynamics: New York, John Wiley & Sons, 392 p.

Shuisky, Y.D., 1984, Dynamics of Kiliya delta coastline of the Danube river: Soviet Oceanographic Institute, Transactions, v. 172, p. 50–60.

Slanar, H., 1945, Zur Kartographie und Morphologie des Donaudeltas, Geogr. Gesellschaft, Mitteilungen, p. 1–12.

Smith, J.M., Sherlock, A.R., and Resio, D.T., 2001, STWAVE: Steady-State Spectral Wave Model, User guide for STWAVE Version 3.0, ERDC/CHL SR-01-01: U.S. Army Engineer Research and Development Center, Vicksburg, Mississippi.

Spratt, T.R.N., 1860, On the freshwater deposits of Bessarabia, Geological Society of London, Quarterly Journal, v. 16, p. 281–294.

Stanev, E.V., Beckers, J.M., Lancelot, C., Staneva, J.V., Le Traon, P.Y., Peneva, E.L., and Gregoire, M., 2002, Coastal–open ocean exchange in the Black Sea: Observations and modeling: Estuarine, Coastal and Shelf Science, v. 54, p. 601–620.

Stanley, D.J., 2001, Dating modern deltas: Progress, problems, and prognostics, Annual Review of Earth and Planetary Sciences, v. 29, p. 257–294.

Stanley, D.J., Mart, Y., and Nir, Y., 1997, Clay mineral distributions to interpret Nile cell provenance and dispersal: II. Coastal plain from Nile delta to northern Israel, Journal of Coastal Research, v. 13, p. 506–533.

Syvitski, J.P.M., and Bahr, D.B., 2001, Numerical models of marine sediment transport and deposition: Computers & Geosciences, v. 27, p. 617–618.

Syvitski, J.P.M., and Trincardi, ___, in press, Marine Geology, Comdelta special issue.

Todd, T.W., 1968, Dynamic diversion: influence of longshore current-tidal flow interaction on chenier and barrier island plains: Journal of Sedimentary Petrology, v. 38, p. 734–746.

Valsan, G., 1934, Nouvelle hypothèse sur le Delta du Danube, Congr. Int. Géogr., Comptes rendus, II, p. 342–355, Warsaw.

Vasilesco, G., 1929, Debit solide du Danube et evolution recente du son delta: Bucharest, p. __.

Vespremeanu, E., 1983, Geomorphological evolution of Sfintu Gheorghe arm mouth (Danube delta, North-West of the Black Sea) in the last 200 years: Revue Roumaine de Geologie, Geographie et Geophysique, v. 27, p. 61–68.

Vespremeanu, E., and Stefanescu, D., 1988, Present-day geomorphological processes on the Romanian delta and lagoon littoral of the Black Sea: Universitatii Bucuresti, Analele, Geografie, v. 37, p. 85–91.

Vespremeanu, E., Vespremeanu-Stroe, A., and Constantinescu, S. 2004, The Black Sea level oscillations in the last 150 years: Universitatii Bucuresti, Analele, Geografie, v. 52, in press.

Vespremeanu-Stroe, A., 2004, Tarmul deltaic danubian al Marii Negre—studiu de geomorfologie: Unpublished Ph.D. thesis, 199 p.

Vidrasco, I., 1924, La voie navigable maritime du Danube: Bucharest, v. __, 45 p.

Vörösmarty, C.J., Fekete, B., and Tucker, B.A., 1998, River Discharge Database: The Institute for the Study of Earth, Oceans, and Space, University of New Hampshire, *http://www.rivdis.sr.unh.edu/*.

Ward, E.T., 1929-30, The Navigability of Lower Danube, Institution of Civil Engineers, Minutes of Proceedings, v. 230, p. 280–332.

Winguth, C., Wong, H.K., Panin, N., Dinu, C., Georgescu, P., Ungureanu, G., Krugliakov, V.V., and Podshuveit, V., 2000, Upper Quaternary water level history and sedimentation in the northwestern Black Sea: Marine Geology, v. 167, p. 127–146.

Wright, L.D., 1977, Sediment transport and deposition at river mouths: a synthesis: Geological Society of America, Bulletin, v. 88, p. 857–868.

Wright, L.D., 1985, River Deltas, *in* Davis, A.R. ed., Coastal Sedimentary Environments, New York, Springer-Verlag, p. 1–76.

Yankovsky, A.E., Lemeshko, E.M., and Ilyin, Y.P., 2004, The influence of shelfbreak forcing on the alongshelf penetration of the Danube buoyant water, Black sea: Continental Shelf Research, in press.

Zenkovich, V.P., 1956, Zagadka Dunaiskoi Delty: Priroda, v. 45, p. 86–90.

Zubakov, V.A., 1988, Climatostratigraphic scheme of the Black Sea Pleistocene and its correlation with the oxygen-isotope scale and glacial events: Quaternary Research, v. 29, p. 1–24.