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Small-scale stratigraphy in a large ramp delta: recent and Holocene sedimentation in the Volga delta, Caspian Sea

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Abstract

The Volga delta differs from all other major deltas in the world by its extremely gentle onshore and offshore gradient (~ 5 cm/km) and by being affected by the rapid sea-level changes of the Caspian Sea, at rates up to a hundred times the global sea-level rise. This paper reports (1) the morphological and facies development of part of the lower delta during the last full sea-level cycle between 1929 and 1995, as monitored using remote sensing and field mapping, and (2) the Holocene development of the delta from outcrop data and augered transects.

During a sea-level fall of 3 m between 1929 and 1977, rapid progradation of levees, composed of fine sand, took place along over 800 distributary channels along the delta front. Smaller distributaries became filled with clay and organics. During the 3-m sea-level rise from 1977 to 1995, aggradation occurred, leading to deposition of silt and clay on the levees and minor filling of the flood basins. Sedimentation rates as established with ^{137}Cs dating are up to 2–5 cm/year. Total thickness of Holocene deposits in the lower delta plain is 4–10 m.

A coarsening-upwards sequence in the Damchik sandpit shows freshening-upwards mollusc assemblages dated around 1000 BP, and has been attributed to the Derbent regression at that time. Four transects with a total of 79 augerings down to 7 m depth show rapid lateral facies changes of: (a) lagoonal clays deposited in the palaeo relief between the dunes, (b) channel sands, (c) levee sands and silts, (d) laminated overbank and interdistributary bay deposits, (e) mouthbar deposits and (f) prodelta clays. Holocene depositional patterns are unrelated to the present drainage network, though the spatial variability is similar to that of the present highly segmented network. Seven ^{14}C datings give a range of 6000–800 BP, and several phases of progradation seem to be present, but the lateral variability is too large and the age data too limited to make a solid correlation with known Holocene sea-level fluctuations.

The Volga delta differs essentially from the classic river-dominated Mississippi delta because the offshore gradient is so gentle that no marine reworking takes place at the outlets, and the friction-controlled bifurcation continues basinward until a very fine maze of distributary outlets is produced. The Volga Holocene sequences more resemble those of the Atchafalaya and Saskatchewan lacustrine deltas. However, they differ from them in not being subsystems of a larger delta but the main depositional facies of the delta as a whole. Moreover, the recent Volga delta development shows that progradation is related to

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forced regression, not to avulsion triggered by base-level rise and/or subsidence. Thus, the Volga delta provides an excellent example of the impact of high-frequency sea-level changes on a ramp margin-type fluvio-deltaic system.

Keywords: Volga delta; Fieldwork; Lower delta plain; Sequence stratigraphy

1. Introduction

The great variety in deltas and deltaic sequences has prompted many authors to create elaborate delta classifications, from the simple threefold subdivision of Galloway (1975) to the 12-fold subdivision of Postma (1995). Although the major controls of delta formation are all the same, including drainage area,

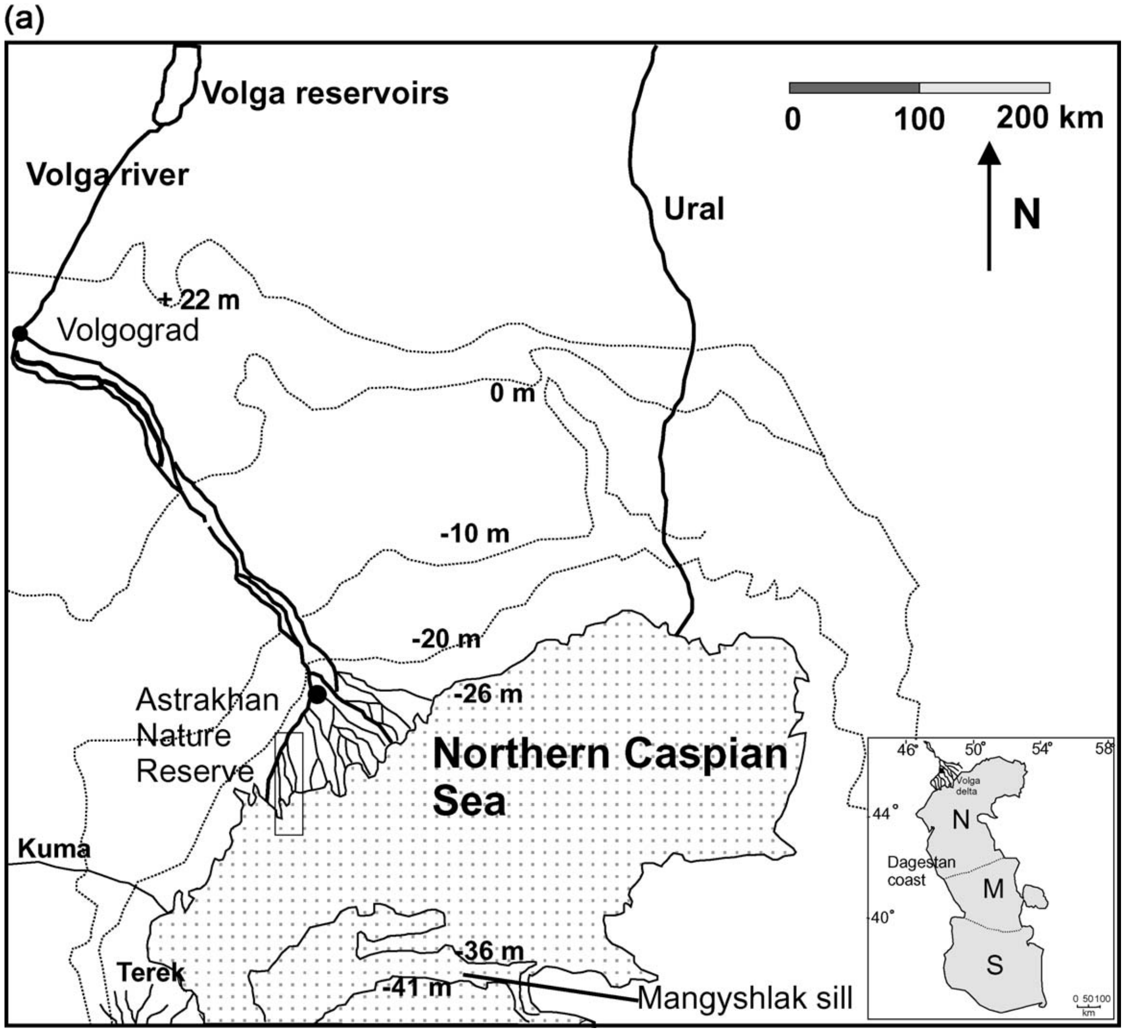


Fig. 1. (a) Northern Caspian basin and Caspian plain, through which the Volga and Ural rivers drain the Russian continent. A rectangle indicates the Astrakhan Nature Reserve where the fieldwork area is located. The inset shows the entire Caspian Sea, divided into Northern (N), Middle (M) and Southern (S) Basins. (b) Bathymetric map of the Northern Caspian Basin, showing the extremely shallow offshore gradient. Dominant wave directions are to the N–NW (36%) and S–SE (32%) (after Kosarev and Yablonskaya, 1994).

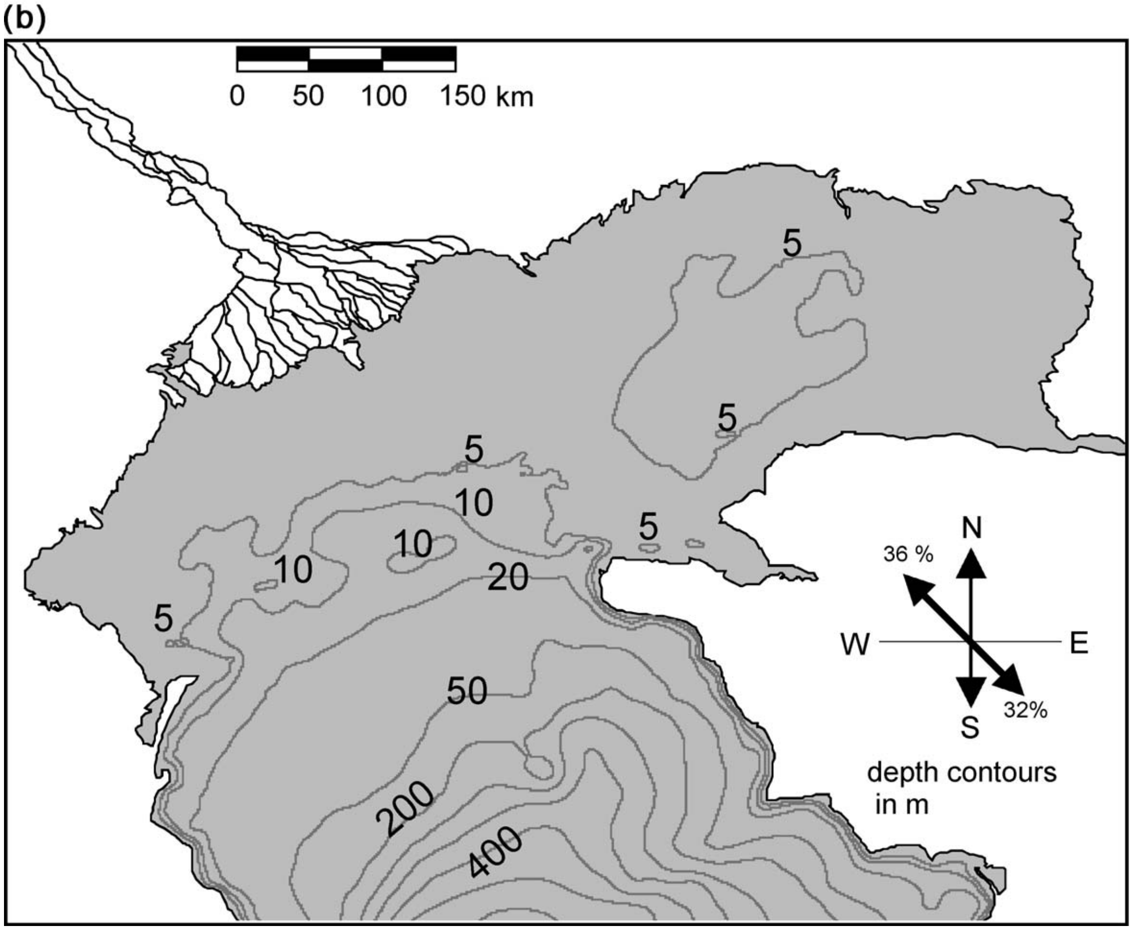


Fig. 1 (*continued*).

water and sediment discharge, sediment grain size, density contrasts in the mouth area, tectonic and compactional subsidence, relative importance of wave and tidal processes, offshore slope, and sea-level history, even modern deltas are so diverse that many of them escape a straightforward classification. Since we know that all modern deltas are unique, we should expect similar variability for ancient deltas.

The Volga delta in the Caspian Sea fits without any doubt into Galloway's fluvial-dominated deltas and in Postma's type 8 (mouthbar-type delta of low-gradient highly stable suspension load river with levees), and yet it is unique in so many aspects that it can be regarded a delta type of its own.

The Volga delta differs in at least four fundamental aspects from other deltas in the world:

(1) Its extremely flat onshore and offshore gradient (5 cm/km; Kroonenberg et al., 1997), comparable only to the Okavango delta (which is not allowed to be called a delta at all by its students; Stanistreet and McCarthy, 1993).
(2) The absence of any marine reworking of sediment at the delta front is in contrast to other examples of river-dominated deltas in the world, such as the Mississippi delta (Fisk, 1961; Gould, 1970).
(3) The absence of subsidence in the delta area with levelings indicating rather a moderate uplift in part of the delta (Lilienberg, 1985, 1995).

(4) The extremely rapid sea-level changes of the Caspianx, up to a hundred times as high as eustatic changes (Kroonenberg et al., 1997, 2000a). In fact, sea-level change alone controls the accommodation space for deposition, since subsidence is negligible.

In this paper, we present data which show that these unique circumstances also have produced a unique type of sedimentary sequences in the Holocene. Classically, the role of fluvial systems and ramp marginal settings has been largely overlooked in the sequence stratigraphical concept (Van Wagoner et al., 1988, 1990; Posamentier and Vail, 1988). More recently, it has been recognized that dynamics in fluvial systems introduces spatial and temporal complexity that has to be incorporated into the sequence stratigraphical concept (Zaitlin et al., 1994; Schumm, 1993; Shanley and McCabe, 1994). This study of the Volga delta may add to the ongoing sequence stratigraphic debate, since it is a totally fluvial-dominated delta subject to rapid sea-level change.

2. Geological setting and physical processes

The Volga River has a drainage area of 1,380,000 km^2 (Rodionov, 1994). The Volga delta (Fig. 1A) has a length of 120 km and a coastline width of 200 km (Kosarev and Yablonskaya, 1994), making it one of the major delta systems in the world. The gradient of the whole delta (both onshore and offshore) is extremely gentle (about 5 cm/km; Fig. 1B).

The climate of the interior Eurasian continent is dry to subtropical with an average precipitation of 527 mm (Coleman and Wright, 1975). During the winter months, December through March, ice partly covers the Volga delta and Northern Caspian Sea (Fig. 2). The average annual discharge is 7835 m^3/s over 1888–1980 (Kosarev and Yablonskaya, 1994). Its distribution is highly influenced by snowmelt, showing peak discharges in May (Polonski et al., 1998; Fig. 2). The highest measured discharge peak under natural conditions, in 1926, is 52,000 m^3/s (Shoubin and Babich, 2001). Suspended sediment load is on average 0.2317 m^3/s (Lisitzin, 1972), whereas bedload has been estimated to be 23% of the average annual sediment load (Shoubin and Babich, 2001).

The Caspian Sea is the largest closed basin on Earth. There are no surface outlets and thus no connections with the world oceans. The basin can be subdivided into the Northern, Middle and Southern Caspian sections (Kosarev and Yablonskaya, 1994). Each part is about the same in surface area, but the volume of the northern part is only 1% of the total volume of the basin (Rodionov, 1994). This area is very shallow, with depths not exceeding 10–15 m (Fig. 1B). About 20% of the Northern Caspian basin is less than 1 m in depth (Baidin and Kosarev, 1986; Rodionov, 1994), so it can be described as a ramp margin (e.g., Van Wagoner et al., 1990). Because of the shallow depth, little wave action exists in the Northern Caspian basin. Storm-strength

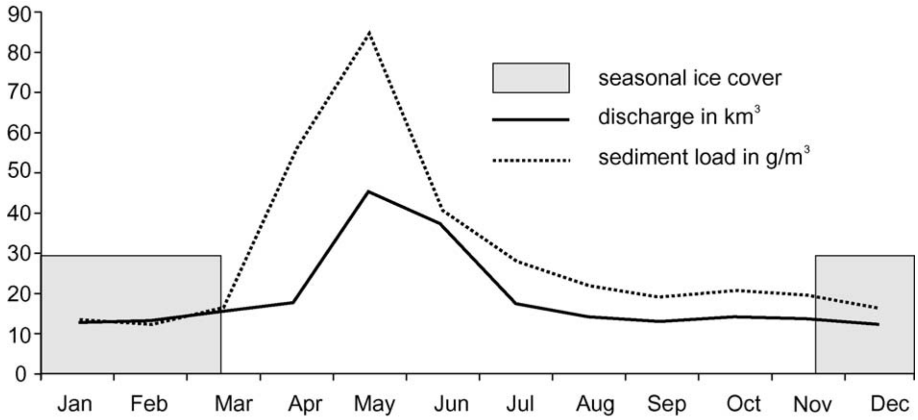


Fig. 2. Annual distribution of discharge (km^3) and sediment load (g/m^3) of the Volga river near the delta apex averaged over 1950–1993 (after Polonski et al., 1998). Shaded area shows the period that the Volga is ice-bound, generally December through March.

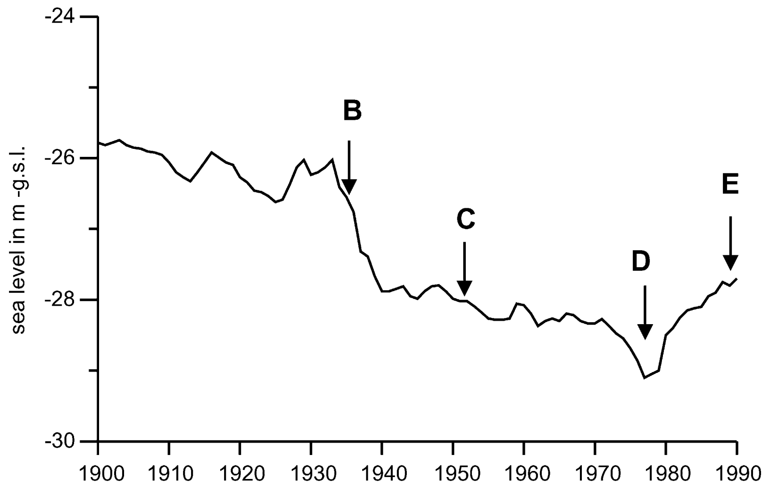


Fig. 3. Caspian sea-level curve between 1900 and 1990. CSL indicated in meters below global sea level (after Kosarev and Yablonskaya, 1994). Arrows indicate the points in time that the aerial photographs were taken, which have been used in Fig. 5.

winds are very rare (1–3 days yearly) in the northwestern part of the Caspian Basin (Kosarev and Yablonskaya, 1994). Even then, the wave heights rarely exceed 1 m (monitored since the 1950s). Prior to the sea-level fall in the 1930s, storm surges periodically inudated the Volga delta, but at lower sea level, the occurrence zone of these surges shifted 40–50 km seaward (Rusakov, 1989). The fiercest storms occur during the cold season and trigger storm surges, although the ice cover attenuates the impact of the waves (Kosarev and Yablonskaya, 1994). The discharge of the Volga (about 80% of the total influx of the Caspian Sea; Kosarev and Yablonskaya, 1994) and the Ural has such important impact on the total water volume due to the shallowness that the salinity is low in the northern part, generally below 3 ppm. The delta front, where fluvial and basinal processes still interact, forms a wide transitional zone. As a consequence, the

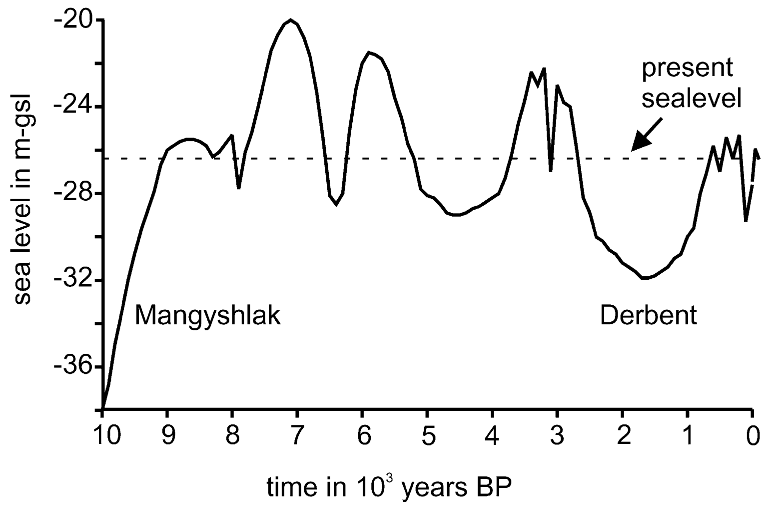


Fig. 4. Holocene Caspian sea-level curve (after Rychagov, 1997), reconstructed based on the heights of terraces along the Dagestan Coast (see inset in Fig. 1).

prodelta facies are located far into the basin, about 200 km south from the coast line. The Caspian Sea has essentially no tides. Based on harmonic analysis of tidal characteristics, the mean range of tide over the entire basin is only 4.5 cm (Spichenko, 1973 in Kosarev and Yablonskaya, 1994). Due to all these factors, the delta system is totally fluvial-dominated.

Sea-level changes within the Caspian Basin are dynamic (Fig. 3). From 1930 to 1977, sea level dropped by 2.7 m, and from 1977 to 1995, it rose at a rate of 15 cm/year (Kaplin and Selivanov, 1995). Numerous transgressions and regressions of the Caspian Sea have occurred in the recent past as well (Svitoch, 1991; Ignatov et al., 1993). The Holocene sea-level history has been reconstructed based on a marine terrace section along the Dagestan coast (Rychagov, 1993a,b, 1997; Fig. 4). Five transgressional phases have been described and dated around 8000, 7000, 6000, 3000 and 200 BP. The lowest documented sea level is estimated at − 50 m below global sea level at the very end of the Pleistocene or very early Holocene (Mangyshlak regression). The Derbent regression, around 1500 BP, reached a probable minimum of at least − 32 m. The maximum level reached by the Caspian Sea during the Holocene is around − 20 m, the elevation of the present delta apex. The frequent Holocene Caspian sea-level oscillations with an amplitude of at least 20 m cause an entirely different forcing process for delta building and erosion than the familiar Holocene sea-level curves for major ocean basins elsewhere in the world.

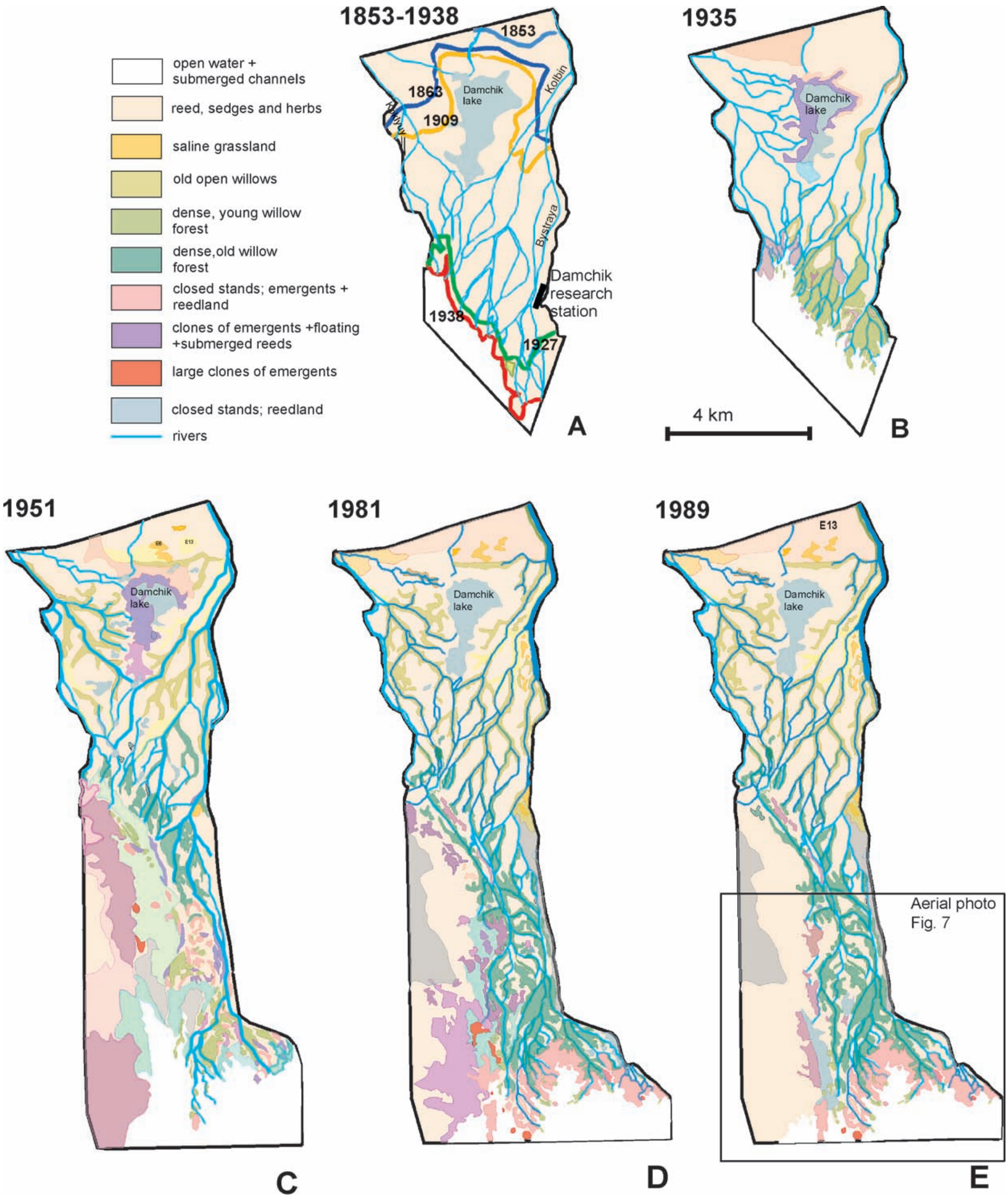
3. Morphology and dynamics during the last sea-level cycle

The Volga delta can be subdivided into an apical part, an upper delta plain and a lower delta plain based on its morphology (Rusakov, 1989; Kroonenberg et al., 1997, their Fig. 3; Polonski et al., 1998). The apical part is characterised by rapid lateral shifts of a maze of sinuous channels. In the upper delta plain, the five major low-sinuosity distributaries that start to fan out from the apex are largely incised into Late Pleistocene and early Holocene eolian longitudinal dunes, the so-called Baer Hills (see aerial photographs in Kroonenberg et al., 1997). Deposition during the annual floods largely takes place in the interdune areas and there are few constructional forms apart from the islands in the middle of the channels. The most dynamic parts of the delta are the lower delta plain and the delta front, the only place where constructional forms are being generated and recent deltaic sequences are being preserved on top of the Pleistocene–Early Holocene palaeo-topography. Their total thickness usually amounts to not more than 4–8 m. This paper only discusses lower delta plain dynamics and stratigraphy.

In the lower delta plain, the distributaries are bifurcating further to the transition with the delta front. There are over 800 outlets along a total coastline length of 200 km (approximately four outlets per kilometer). Most of these are not wider than 10–20 m. The few major outlets such as the Bakhtemir canal are dredged for shipping. The absence of major outlets is in strong contrast with the large hundreds of metres wide passes in the birdfoot delta of other major river-dominated delta, e.g., the Mississippi.

Morphological changes in the lower delta plain in the past 150 years have been studied in detail in the Damchik part of the Astrakhan Biosphere reserve, western part of the lower delta plain (Fig. 1A). The nature reserve was established in 1919 and has been largely untouched by man since then. Detailed maps are available since 1853, and aerial photographs since 1935. The older history has been carefully recon-

Fig. 5. Delta evolution over the last 3-m sea-level cycle from 1930 to 1995 based on aerial photograph analysis and vegetation mapping (Lychagin et al., 1995; Labutina et al., 1995; Baldina et al., 1999; Kroonenberg et al., 2000a,b). Sea-level fall leads to progradation of small-scale sandy levees on offshore clays (coarsening upwards). Progradational sand bodies are thin (< 1.5 m), narrow (metres) and show high lateral variability (tens to hundreds of metres). Distributaries further upstream die off and get clay fills (fining upwards). Sea-level rise leads not to retrogradation but only to aggradation and deposition of clay on top of the sandy levees (FU). (A) Coastlines of long-lasting highstand between 1853 and 1938, reconstructed from old maps (after Belevich, 1956). (B) 1935, just after the highstand. Coast located close to Damchik. Since 1814, only very slow progradation had occurred. (C) 1951, the regressive system tract with evident progradation. A skeleton of prograding levees has extended over 40 km seaward. (D) 1981, just after the lowstand. The contours are similar to 1951, but flooding has filled in the basins. (E) Transgressive system tract; aggradation. The contours seem still intact despite sea-level rise due to stabilizing vegetation. The inset shows the location of the photo used in Fig. 7.


1853-1938
1853
1863
1909
Damchik lake
Kolbin
Bystraya
Damchik research station
1938
1927
A
1935
Damchik lake
4 km
B
open water + submerged channels
reed, sedges and herbs
saline grassland
old open willows
dense, young willow forest
dense,old willow forest
closed stands; emergents + reedland
clones of emergents +floating +submerged reeds
large clones of emergents
closed stands; reedland
rivers
1951
Damchik lake
C
1981
Damchik lake
D
1989
E13
Damchik lake
Aerial photo Fig. 7
E

structed in numerous classic papers by Belevich (1956, 1958, 1963). Our team has studied in detail the changes that occurred during the last 1929–1995 sea-level cycle using maps, aerial photographs and satellite imagery (Baldina et al., 1999; Kasimov et al., 1995; Labutina et al., 1995; Kroonenberg et al., 1997, 2000b). Fig. 5A–E shows the development of the delta in the Damchik area (Baldina et al., 1999).

Fig. 5A shows that the whole area depicted in Fig. 5B–E was still open water in 1853. A large part of this channel network formed as a result of a previous phase of delta progradation between 1909 and 1927, during a short-lived regression of the Caspian Sea (Belevich, 1956). The lake between the two main distributaries, Koklyuy in the west and Bystraya in the east, started to form as an interdistributary bay, but was already almost isolated in 1935. The 1935 map (Fig. 5B) shows a rapidly bifurcating and anastomosing network of channels, which flowed into the sea near the Damchik research station. Sea level had already started to fall in 1929, but its main effects are only visible in the next stage. In 1951 (Fig. 5C), the delta had rapidly prograded across the shallow sea floor for over 25 km. Large parts of the shallow sea bottom emerged passively. Sea level dropped further until 1977, at a lower pace than in the 1930s, however. The delta distributaries advanced another 500 m. Since 1978, sea level has started to rise, and in the 1981 images, the delta contours are largely maintained (Fig. 5D). While sea level had risen another metre in 1989, the contours of the delta front, surprisingly, hardly changed (Fig. 5E). This lack of change is partly because reeds in interdistributary bays and willows on levees both trap sediments, so that the accretion could keep pace with sea-level rise (Baldina et al., 1999). Field observations in 2000 show that as a result of the sea-level fall after 1995, progradation started again. New fine-grained sandy levees are again being formed at the outlets, 1–2 m wide, and not more than 1.5 m thick.

4. Sedimentary facies

Channels in the lower delta plain have a low to moderate sinuosity and show anastomosing patterns, i.e., forming interconnected channels, which enclose islands or floodbasins (cf. Makaske, 1998). Belevich (1956) gave a detailed account of the sedimentation processes and forms at the outlets during the period of sea-level fall in the 1950s. Most large channels, up to 200–300 m wide and 3–4 m deep, bifurcate into smaller ones, 10–20 m wide, when reaching the delta front. Flow velocity there may reach 1 m/s. Because of this high-flow velocity and the fine grain size of the sediment, little sand is being deposited within the channels. Their bottom is mainly eroding into the underlying clays. Small subaqueous bars of fine-grained sand mixed with shell debris are found to move downstream on the bottom of the channels; rarely, they grow into fine-grained sandy midstream islands (Belevich, 1960; Polonski et al., 1998). Repeated bathymetric surveying in the major channels (outside the Damchik area) indicates continuing channel erosion up to 10–25 cm/year in the period of sea-level fall between 1938 and 1977 and still 5–10 cm/year during rising sea level between 1977 and 1990 (Polonski et al., 1998).

Sedimentation occurs largely in subaqueous *levees*, a few tens to hundreds of metres in length, which eventually grow up to 1 m above low water level. The levees consist mainly of very fine-grained sand and silt, and overlie dark clayey prodelta deposits. The strongest vertical growth occurs in the first few years after emergence. Once levees form, they can become several kilometres long, but they are less than 1–2 m thick, and their width is restricted to a few tens of metres, as can be seen from the distribution of willow vegetation on the map (Fig. 5C). Belevich (1960) found up to 25 cm of new sediment on a levee after a large peak flood in 1955. During the sea-level rise after 1977, floods deposited increasing amounts of silts and clays on top of the sands, so that most of the sandy levees now have finer-grained covers.

In the wake of the prograding delta front, a constant reorganisation of the distributary network takes place—some gaining in size, others losing their function and dying. By comparison with old maps, it can be seen that distributaries that formed between 1909 and 1927 (Fig. 5A) have died off since the progradation pulse between 1935 and 1951 (Fig. 5B–E; Belevich, 1956; Baldina et al., 1999). They are filled with a few metres of fine silty–clayey mud overlying fine sandy–silty lag material. An augering in the mud filling of the abandoned Kolbin Creek, a right distributary of the Bystraya that started to form from 1909 onwards, shows a sedimentation rate of 2 cm/

year using ^{137}Cs dating (Winkels et al., 1996). Most distributaries formed during delta progradation of the 1930s are still active.

Between the prograding delta outlets, *interdistributary bays* are formed, 1–1.5 m deep, in which dark-coloured clay and silt are being deposited during floods, usually not more than a few decimetres since their formation in the 1930s. The interdistributary bays are partially overgrown with aquatic vegetation, and brackish to freshwater molluscs dwell on their bottom. Peat is absent. Locally, as in Katyushka Creek, small crevasse splays develop in them, depositing small levees in the same way as at the delta front. In the isolated Lake Damchik (Fig. 5B), millimetre-thin annual layering of organic-poor and organic-rich laminae due to summer floods and winter freezing has been observed.

The very shallow *prodelta* (or 'avandelta'; Zenkovich, 1967; Belevich, 1963) extends for over 80 km basinwards from the delta front and is deposited in water not deeper than 20 m. Even at 200 km from the delta, front water depth is not more that 10–15 m. Linear sediment plumes can be followed up to 75 km offshore on satellite imagery. These features demonstrate that much sediment is bypassing the prodelta until somewhat deeper waters are reached. In the prodelta close to the delta front, at water depths of 0.8–1.3 m, a top layer of 15–40 cm of fine-grained dark grey to black sand mixed with organic material and shell fragments is found, which probably is mixed during the annual spring and fall storm surges, in transit towards deeper waters. In the reed fields, the surface sediment has a finer texture. The sedimentation rate over the last 35 years inferred from ^{137}Cs dating from a single core in this area is about 5 cm/year (Winkels et al., 1996). Abundant complete brackish water molluscs and mollusc fragments are present at 0.2–0.3 m depth, apparently living in situ. Below this level, sediments are undisturbed older cohesive clays, reducing the extent of reworking.

5. Holocene stratigraphy

5.1. The Damchik sandpit

A sandpit east of the settlement of Damchik (Fig. 5) in the Astrakhan Biosphere Reserve shows a coarsening-upwards sequence of clays overlain by silty sands and fine sands (Fig. 6). The mollusc assemblages show that the underlying clays have more brackish water species (*Monodacna* spp., *Hypaenis plicatus*, *Adacna vitrea*, indicative of salinity of about 0.7%, the so-called Azov facies; Svitoch, 1991), while the uppermost sandy layers have essentially freshwater species, such as *Unio* spp., *Viviparus* spp., *Sphaerium corneum*, *Dreissena polymorpha polymorpha* and *Dreissena polymorpha caspica*. The diatom assemblages show a similar trend. Assemblage progression is in line with an origin by a prograding delta system in which fluvial sands are deposited on brackish avandelta deposits. Two ^{14}C dates have been obtained: 1320 ± 100 BP (MGU 1550) on shells in the fluvial sands and 1085 ± 100 BP (MGU 1547) on plant remains in the prodelta clays (Fig. 6). In spite of the discrepancy in ages, which is possibly related to the different materials, the coarsening-upwards sequence might either be attributed to the Derbent regression known from historical data to have taken place around 1500 BP (cf. Fig. 3; Rychagov 1997) or to the subsequent lower-amplitude cycle. Further details are given by Svitoch and Badyukova (2001).

5.2. Augering survey

An onshore and offshore survey by shallow augerings up to 7 m deep has been carried out in a subarea (75 km^2) of the Damchik part of the Astrakhan Biosphere Reserve (Fig. 7). Twelve augerings were located in the delta plain and 67 in the delta front zone. Depth, texture, colour, content of organic matter, lime content and presence of shell fragments and complete shells in the sediments were described and organic matter was sampled for ^{14}C dating. Two hundred samples of the most recent deposits (within 1.2 m of the surface) were ultrasonically treated and the grain size distribution has been determined with Coulton laser analysis (Fig. 8).

Selected logs have been compiled into four transects: two along-stream (dip) transects and two perpendicular to the flow direction (strike) transects (Fig. 9A–D). Perusal of the transects shows their extreme variability, both in strike and in dip direction. Contrary to common practice, we have refrained from establishing correlations between the augerings for several reasons.

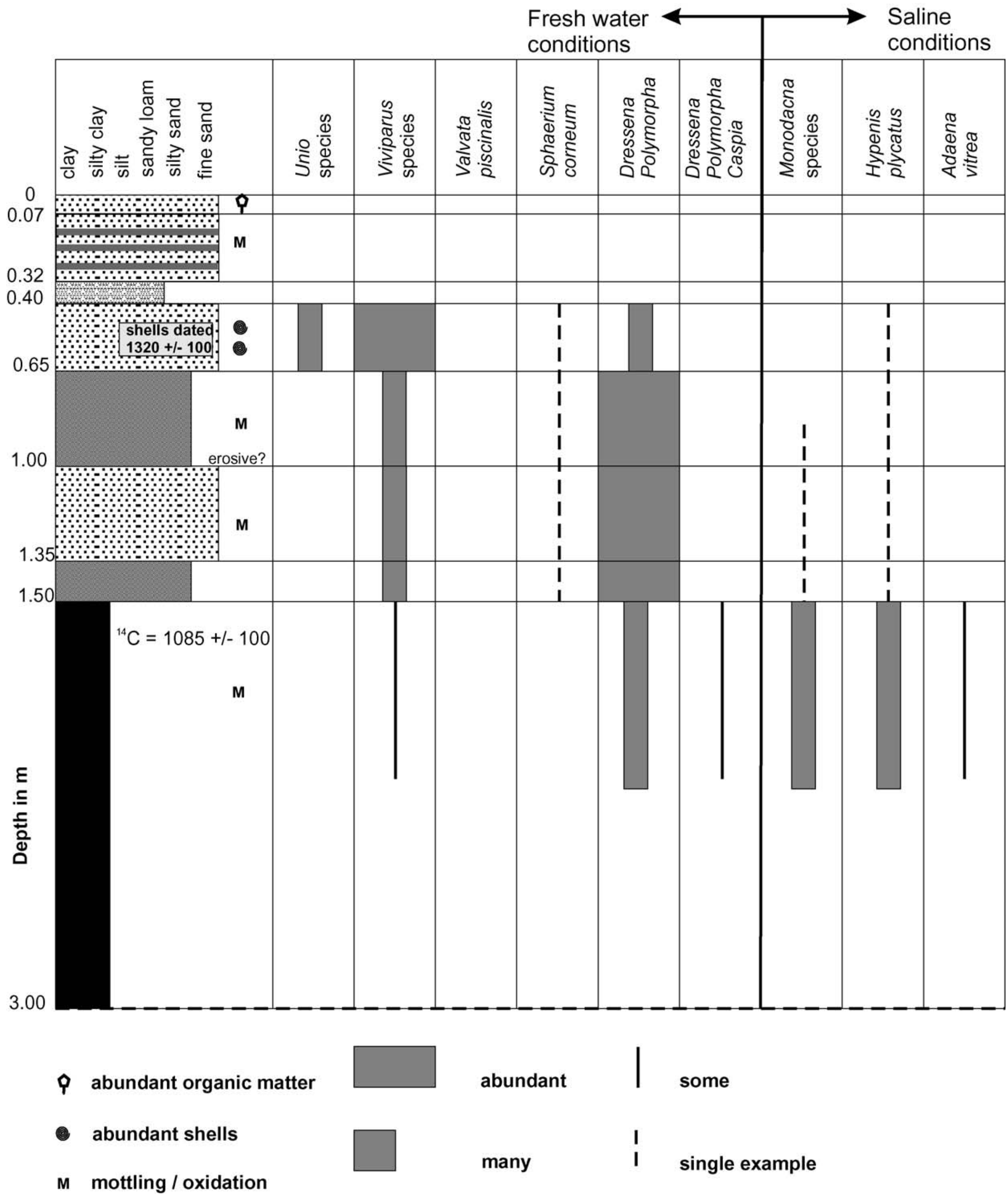


Fig. 6. Damchik sandpit progradation during the Derbent regression (at approximately 1500 BP) deposited a coarsening-upward cycle with freshening-upwards mollusc assemblages.

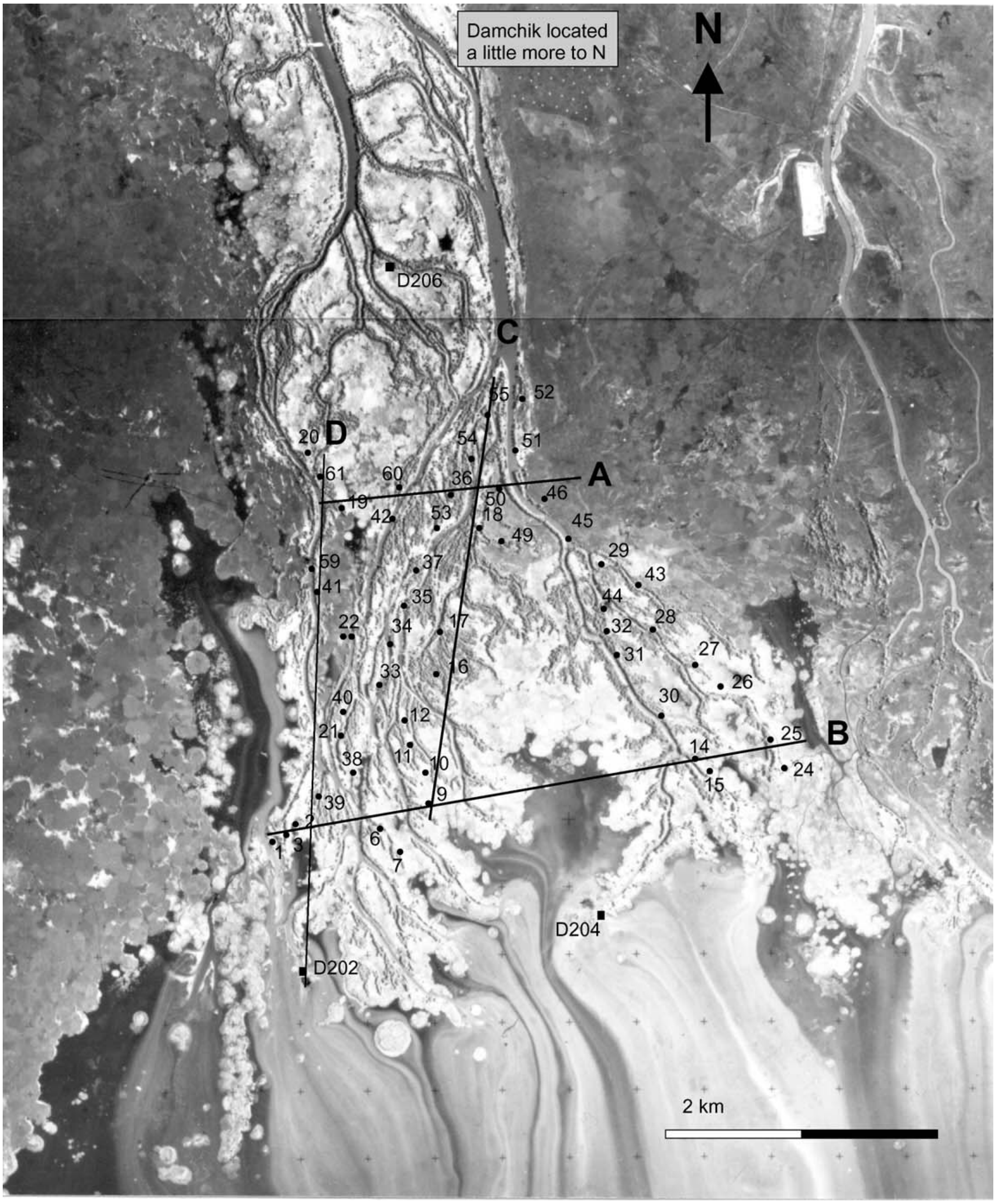


Fig. 7. Aerial photograph (May 1989) with the network of delta channels in the Damchik part of Astrakhan Nature Reserve, showing the anastomosing pattern of distributaries, distributary mouths and marshes with small crevasses. The augerings used in this article and reconstructed transects (Fig. 9) are depicted.

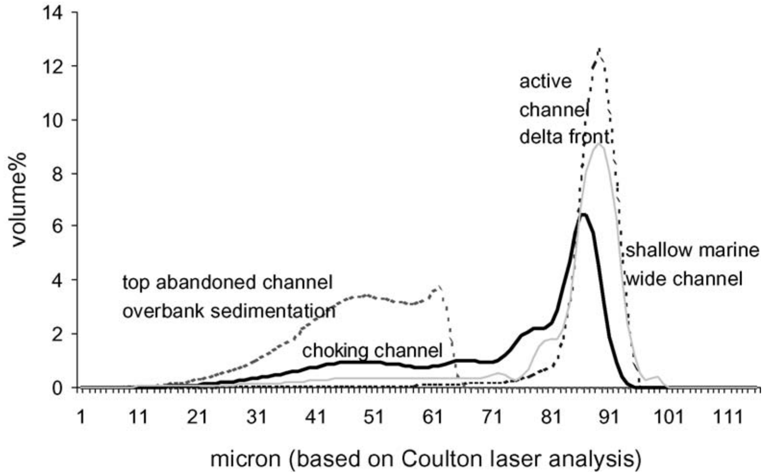


Fig. 8. Grain size histograms of typical Volga delta sediments showing the well-sorted fine sands of the delta and avandelta and the typical interdistributary bay clays.

In the first place, only the upper 1–2 m at most of the sections have been deposited by the present distributary system that developed during sea-level fall in the 1930s. The deeper parts of the sections have been deposited in previous stages of delta progradation with possibly a very different distributary channel configuration. Secondly, ^{14}C dates that could be obtained from the rare organic layers are in the range between 6700 and 800 BP, a period in which various important transgressions and regressions are known to have occurred (Rychagov, 1993a,b; Fig. 4). Attributing a specific sequence to a specific sea-level cycle is extremely hazardous, therefore, especially in view of the fact that the modern regression already left levees of 1–2 m thick. Only in one augering, no. 7 (Transect B), were two datable layers found, leading to an average sedimentation rate of 1.4 mm/year. Most augerings did not contain datable material at all, except for molluscs, which we rarely sampled for dating because of the probability of reworking and resedimentation. Correlation on the basis of the present age data, therefore, is not possible. The extreme variability is in harmony with the present sedimentary system at the delta front, with small channels, small levees and small backswamp areas. Channel and levee sand ribbons are generally not more than a few tens to hundreds of metres wide, and not more than a few metres thick, so the chance to find sand bodies of greater continuity is slight, even though the augerings are generally not further apart than a few hundreds of metres. The resolution of the data set, roughly five augerings per square kilometer, is too low to account for the lateral variability of the sedimentary system. The extreme variability in depth proves, on the other hand, that the delta front has been situated in this area several times during the Holocene. This model is consistent with the conclusion of Rychagov (1997) that sea level has been not far from the -25 to -26 m data during a large part of the Holocene.

In spite of the impossibility to correlate the sections, we have identified a number of typical sequences, which we correlate with typical facies on the basis of the data from the present-day delta front.

5.3. Facies descriptions

5.3.1. Late Pleistocene–Early Holocene marine clay facies

The base of the studied sequence consists of brown clays with admixtures of sand and small shell fragments. These sediments resemble the so-called chocolate clays, attributed to the $+50$ m Early Khvalyn transgression (Belevich 1958, 1960; Kroonenberg et al., 1997). The top of the sediments defines a clear palaeo-topography (Fig. 10) consisting of E–W trending ridges of perhaps 5 m high. At some places, brown clays are already found at less than 3 m depth; in others, they are not even encountered at 7 m depth. They are interpreted as mixture of marine clays and reworked Baer hills (longitudinal dunes). The ridges in the palaeo-topography are transected by N–S valleys, with a locally reconstructed palaeo gradient of 8 cm/km. These valleys probably were incised by streams during the following lowstand, the Early Holocene Mangyshlak regression (Fig. 4). Larger-scale sections through the entire delta (~ 100 km length) show a similar pattern (Aybulatov et al., 2001; Fig. 10 in Kroonenberg et al., 1997).

5.3.2. Lagoonal interdune fill facies

A light grey to white carbonate-rich sediment is occasionally found on top of the brown sediments, especially in the depressions between the ridges. It is a homogeneous mix of clay, sand and silt, texturally similar to the underlying brown clays. They may contain thin organic-rich layers. Shell fragments from augering 52 have been dated at 6760 and 6600 BP. Organic fragments from the same sample (Gr A5442; Table 1) give an age of 4140 ± 50 BP.

In the depressions of the palaeo relief, we find dark-coloured clays, averaging 1.9 m in thickness, commonly with several layers rich in molluscs, e.g., as in transect B in holes 7 and 15 and in transect C, holes 54 and 36. Se-veral horizons of ripened clay indicative of initial soil formation have been found, two of which can be followed at about the same depth throughout the different transects. Lagoonal deposits in augering 22 show an age of 5040 ± 50 BP (Gr A6074). Two dates from lagoonal clays in augering 7 show that deposition continued from at least about 4700 to 3500 BP, indicating a sedimentation rate of 1.4 mm/year. Molluscs sampled from these deposits in augering 34 were dominated by fresh water species *Viviparus viviparus* L.

The fine grain size, the accumulation of organic matter, the apparent in situ presence of molluscs and the ripening horizons suggest deposition in a quiet sedimentary environment between the ridges, with periodical emergence and submergence. Such inter-ridge lagoons are absent in the present lower delta plain in the Damchik area, but they resemble the interdune areas between the Baer hills, which become flooded annually by the Volga west of the present-day Volga delta itself (see Fig. 3 in Kroonenberg et al., 1997). Carbonate enrichment in the light-coloured layers may be due to evaporation of flood-water in interridge depressions, as occurs in present-day isolated interdune lakes that are mostly beyond the reach of the annual Volga floods. Postdepositional caliche-type soil formation may also play a role because calcareous nodules are sometimes found in them.

Sandy deposits in sample Gr N21065 dated at 4150 ± 80 BP (augering D206; Fig. 7) and hole 17 might be correlated to the channels dissecting the Baer hills palaeo-topography. Molluscs from similar sediments in augering 8 (close to Damchik) are riverine water species, *V. viviparus* and *Sphaerium rivicola*.

5.3.3. Channel fill facies (fining upwards)

We infer that the thickest sand bodies (consisting of fine sand, 150–210 μm), with abrupt lower boundaries encountered in the augerings, represent the sandy bedload deposits of stable mature channels. They are characterised by their homogeneous gray colour. The thick sand bodies in augerings 45 and 11 are interpreted as preserved active multi-phase channels deposits. These can form complexes up to 3 m thick. Abandoned channels show a coarse base and a fining upward infill (see Fig. 9). In auger 17, a channel abandonment sequence between approximately 5 and 2.5 m has been preserved.

5.3.4. Levee complex facies (fining upwards)

These are small-scale fining-upwards sequences of fine sand (56–150 μm according to Coulton laser analysis), silt and clay (Fig. 8). In contrast to the channel fills, rooting and mottling are common in these sediments, indicating frequent and prolonged emergence. Thickness of the recent levee deposits, resulting from the last 65 years of sea-level cycle, may be up to about 1.4 m.

5.3.5. Overbank and interdistributary bay facies (laminated)

Crevasse splays are represented by strongly layered silts and fine sands (e.g., augering 36), probably deposited during peak discharges. Thick clay deposits

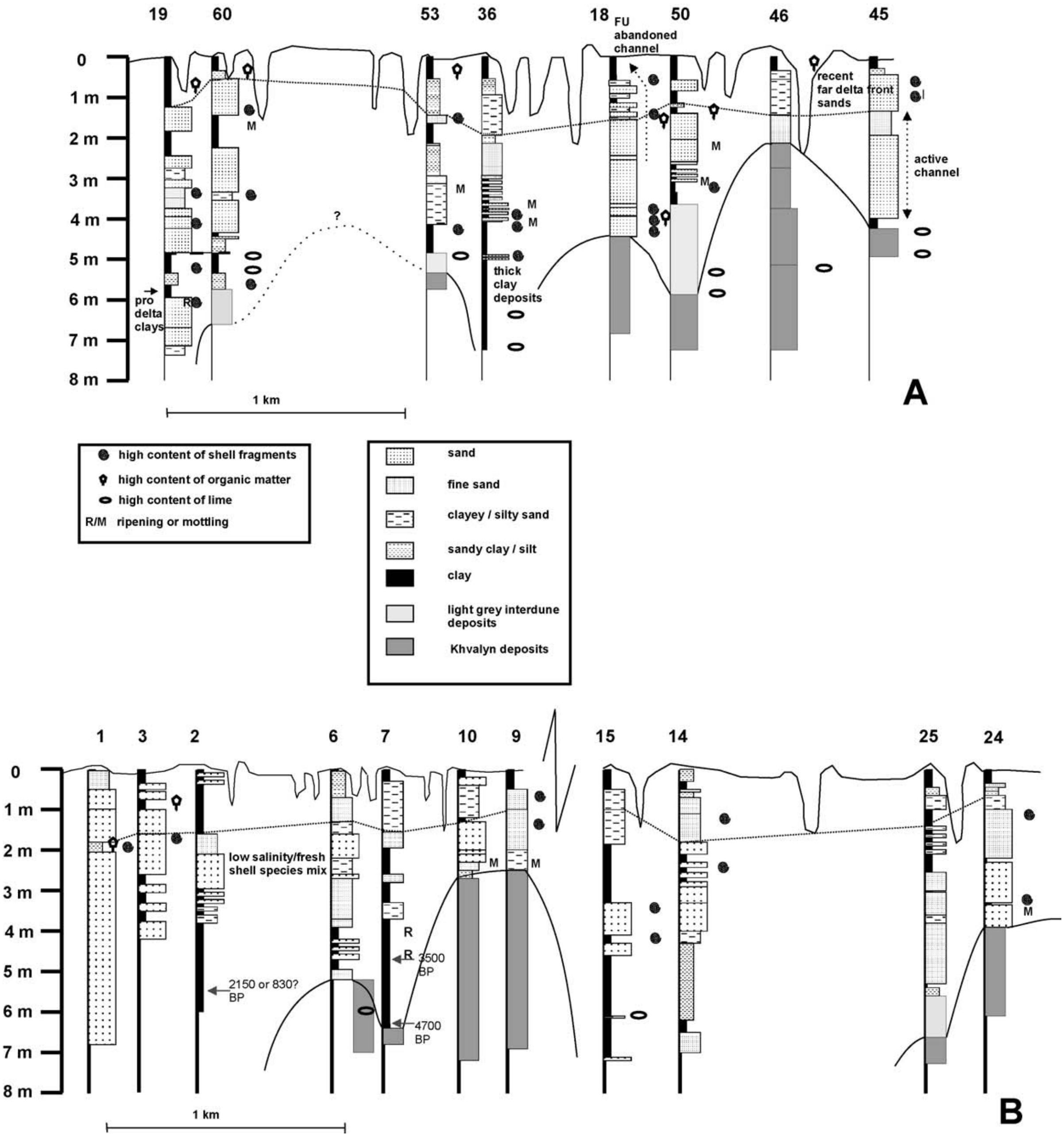


Fig. 9. Transects based on augerings show the stacked sequence in the study area. ^{14}C ages are indicated. The dotted line in the uppermost 2 m indicates the deposits attributed to the most recent delta development. (A) and (B) are oriented W–E, while (C) and (D) are oriented N–S. At the base, a palaeo relief of distinct clay–silt mix of chocolate brown colour has been distinguished; subsequently, lagoonal clays have been deposited. Then rapidly alternating active channels, crevasses, abandoned channel fills, mouth bars and the remains of the recent prodelta platform have been distinguished.

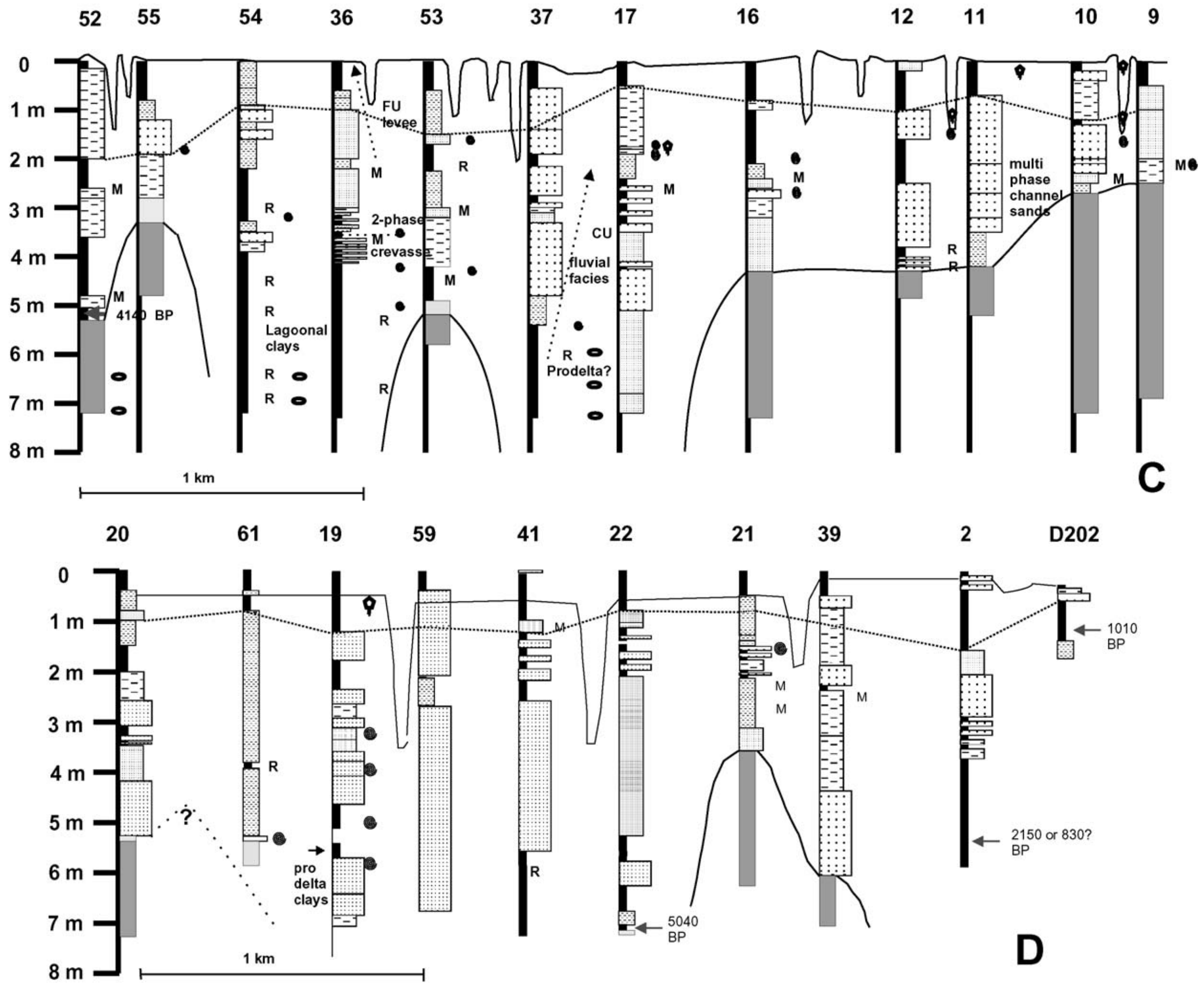


Fig. 9 (*continued*).

in the bays and lakes are rare. Organic matter content of the interdistributary bay sediments is rather low, probably due to high mineralisation rates in the warm climate (Winkels et al., 1996). A crevasse deposit in sample Gr N21064 (augering D204) was dated at 2930 ± 50 BP (Table 1).

5.3.6. Distributary mouthbar complex facies (coarsening upwards)

The Damchik section has been interpreted as a mouthbar complex (Svitoch and Badyukova, 2001). The distributary mouthbar deposits form fine-textured sand bodies in between very shallow bifurcating channels. Augering 1 penetrates an existing mouthbar, which is only about 0.5 m thick. Its general trend is coarsening upwards, although in the present delta, mouthbars are often characterised as having thin fine-textured layers at the very top, which has been interpreted as the response to drowning of the mouthbar during the latest sea-level rise. These deposits are texturally difficult to distinguish from channel sands. Possibly, the presence of a coarse channel lag would be a distinguishing criterion. Sample Gr N21063 (augering D202) with a ^{14}C age of 1010 ± 40 BP was interpreted as a clayey top of a mouthbar or levee. The ^{14}C datings point to deposition during the end the Derbent regression of the Caspian Sea.

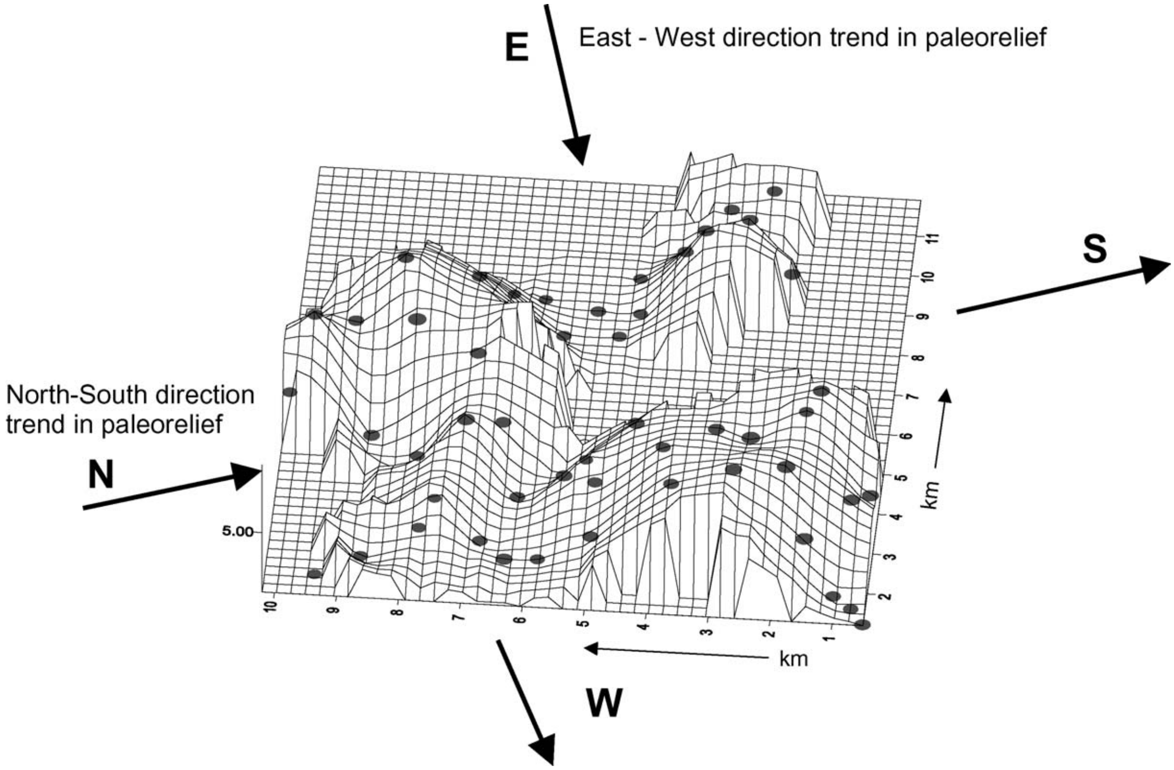


Fig. 10. Palaeo relief as retrieved from depth occurrence of the Pleistocene marine clays, in which two trend directions have been identified; N–S incised valleys of the last large sea-level fall from end-Pleistocene to begin-Holocene and E–W reworked remains of the eolian Baer hills. The dots indicate the augerings.

5.3.7. Delta front; shallow marine delta platform facies

Dark grey to black sands with abundant shells, in average about 0.7 m thick, occur in many sections at shallow depth below the surface (e.g., augerings 9, 14 and 11). The shells appear to be more reworked than in any other deposit. The distinct dark colour and abundant shells are used as indicators for this type of delta front facies in the augerings.

5.3.8. Prodelta clay facies

Another characteristic type of homogeneous clay deposit has been distinguished. The colour is light grey to grey, probably due to low organic matter content. Shell content is rather poor and only fragments are found. These clays show no indications for ripening. They only occur in small layers, quite deep in the augerings (e.g., in augering 19 at 5.7 m depth). These might have been deposited far offshore in the Volga prodelta, resembling prodelta deposits in the classical sense.

6. Evolution of the lower delta plain in the Damchik area

During the Late Glacial or early Holocene, a prominent complex of E–W-oriented longitudinal dunes was formed, probably deflated from earlier marine deposits in the North Caspian Plain. The marine deposits are attributed to the Pleistocene Kvalyn highstand, of which the exact age is still matter of debate (Kroonenberg et al., 1997). The dunes must have formed during a lowstand of the Caspian Sea, which might have been the 9000 BP Mangyshlak regression. Incised valleys, which are N–S-oriented, may have been formed at that time as well. Subsequently, sea-level rise brought the Volga

Table 1
The ^{14}C datings of samples from Volga sediments, taken from Astrakhan Nature Reserve

Sample number	Augering number	Depth (cm)	Layer properties	Fraction (μm)	Age BP ± σ
Gr A5436	2	510	Clay; sandy layers, medium o.m., low lime	org fr. >200	830 ± 40
Gr A5437	2	510	content, noncalcareous layers	wood fr.	850 ± 40
Gr A5443	2	510		org fr. 100–200	2150 ± 40
Gr A6067	7	490	Clay, black (2.5 Y 2/0)	org fr. >200	3540 ± 50
Gr A6077	7	490	High in o.m. in layers, ripened	org fr. 100–200	3440 ± 50
Gr A6073	7	635	Clay, embedded peat, at boundaries abundant lime	org fr. >200	4780 ± 50
Gr A6075	7	635		org fr. 100–200	4620 ± 50
Gr A6074	22	720	Clay, black (2.5 Y 2/0), high in o.m., low lime	org fr. >200	5040 ± 50
Gr A4395	22	720	content, few shell fragments and some whole	org fr. 50–100	4900 ± 70
Gr A4396	22	720		org fr. 100–200	5120 ± 70
Gr A5442	52	555	Sand/clay mix, white to light grey (2.5 Y 2/0),	org fr. >200	4140 ± 50
Gr A6023	52	555	very low lime content	alkali fr. >200	6760 ± 50
Gr A4367	52	555		alkali fr. 100–200	6600 ± 60
Gr N21063	202	80	Clay, grey-greenish brown, o.m.-rich	org fr. 100–200	1010 ± 40
Gr N21064	204	360	Peat embedded in sand, grey with alternating clay layers, shells present	org fr. 100–200	2930 ± 50
Gr N21065	206	350	Clayey sand, dark grey shell fragments, lime and gypsum nodules	org fr. 100–200	4150 ± 80

delta close to the dune area, more upstream and similar to the present position. Interdune areas were flooded annually, while interridge lagoons further removed from the Volga delta were flooded only rarely. Evaporation caused enrichment in carbonate in the lake sediments. The age and correlation with sea level are uncertain because only a single ^{14}C dating is available, for which it is also uncertain whether the molluscs dated at 6000 BP in the carbonate-rich lagoonal deposits are in situ. ^{14}C datings do demonstrate that black lagoonal clays with fresh water molluscs were deposited in the interval between 5050 and 3500 BP. Simultaneously, some Volga distributaries probably flowed through the incised channels, in view of the 4140 BP age for sands in the N–S-oriented depression.

After 3500 BP, the lagoonal sediments and the ridges themselves became covered with a complex of rapidly alternating facies of about 4 m thickness, including sandy channel deposits and distributary mouthbars, abandoned fining upwards channel fills, levees with mottling and a fine-grained top, crevasse-splay deposits and fine overbank sediments. Several sequences may be stacked upon each other. In some cases, previously abandoned channels are reactivated, like in augering 37. In augerings 24 and 11, several lag deposits have been observed. There are no indications for full-marine sedimentation in between the deposition of the fluvio-deltaic sediments. Shell-rich layers might be the only remnants of transgressive phases. This phase of sedimentation evidently bears much resemblance to the present lower Volga delta. The presence of several mollusc-rich layers, ripening and mottling horizons in a single augering evidences periodic deposition and soil formation. These features might be related to various transgressions and regressions of the Caspian Sea, but the rapid vertical and lateral facies changes and the lack of datable horizons so far preclude the establishment of a precise correlation and chronology. The few available radiocarbon datings show that this phase spans at least the period between 2900 and 800 BP. A full regressive cycle is recorded in the coarsening-upwards and freshening-upwards sequence of around 1000 BP in the Damchik sandpit. This period ended with flooding and deposition of the typical mixed mollusc-rich black sand–silt–clay prodelta sediments, ascribed to reworking of the shallow prodelta bottom. These sediments are attributed to the last highstand, which lasted from the beginning of the 1800s to about 1930.

Delta progradation in the 1930s deposited a few decimetres to 1.5 m of distributary mouthbar and levee sands and interdistributary clays on top of the prodelta sediments, as described above. The last phase of sea-level rise since 1977 covered most profiles with

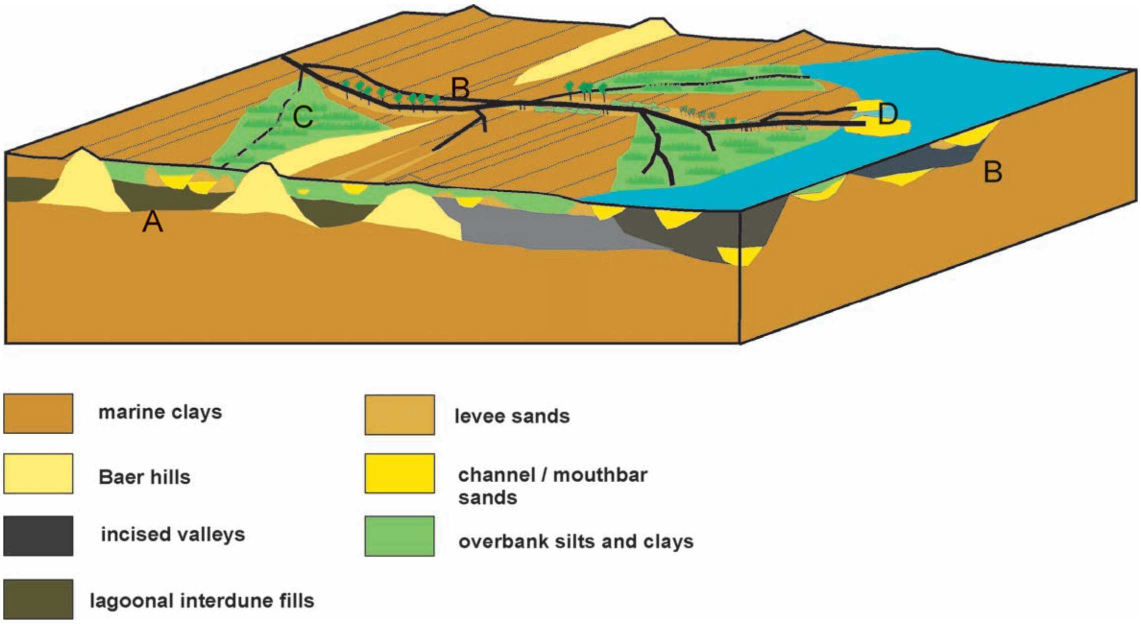


Fig. 11. Simplified representation of the sedimentary architecture of the Holocene Volga deposits. At the base are the marine clays and remains of longitudinal dunes (A). The main channel and mouthbar sandbodies are preferentially located in Early Holocene incised valleys (B). Upstream, previously active channels have now been abandoned (C). Active deposition on vegetated levees occurs at all times (D).

the uppermost clayey sediments. The resulting Holocene sedimentary architecture is schematically summarized in Fig. 11.

Measured sedimentation rates in a levee basin cross-profile in 1955 range from maximal 25 cm/year on top of a levee to less than 1 mm/year at large

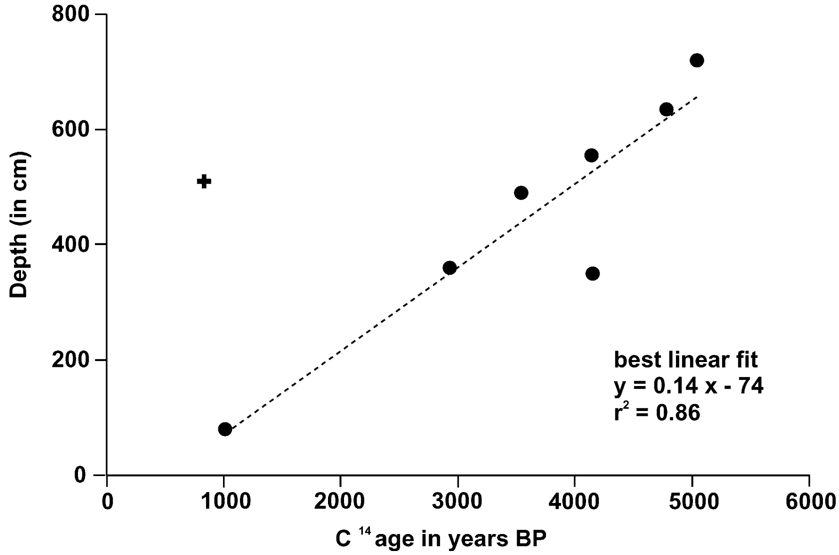


Fig. 12. Plot of the ^{14}C age vs. sample depth shows a linear trend. This allows a rough reconstruction of net sedimentation rate: 1.3 mm/year over the period 1000–6000 BP.

distances from the channel (Belevich, 1960). Deposits of the last cycle of 65 years' duration have an average thickness of 1.1 m, resulting in an average sedimentation rate of 1.7 cm/year. This value may be too high, as augerings on levees are probably overrepresented in our data set, though it is in the same order of magnitude as the sedimentation rates inferred from ^{137}Cs datings over the last 35 years (Winkels et al., 1996).

However, the long-term sedimentation rates are significantly lower. A rough reconstruction of the net sedimentation rate over the Holocene based on the ^{14}C data over 1000–6000 years BP is 1.3 mm/year (Fig. 12). A sedimentation rate of 1.3 mm/year over 10,000 years would yield 13 m of sediment. We found only 5.5 m as an average in the augerings penetrating to the Pleistocene base. Consequently, the sedimentation rate appears to be even lower, averaging only 0.6 mm/year. It is evident that net sedimentation rates in this ramp margin setting are comparatively low.

7. Discussion

Although the Damchik area comprises only a small part of the present lower delta plain, its sedimentary conditions appear fairly representative for the Volga delta front as a whole. Holocene deposits in most of the lower delta plain area do not surpass 5–10 m in thickness, except for two deeply incised valley fills in the central part of the delta (Rachkovskaya, 1951; Kroonenberg et al., 1997; Aybulatov et al., 2001). The dense network of delta distributaries is typical of this delta. Prominent features of the sedimentary architecture of Holocene Volga lower delta plain are (Figs. 11 and 13):

(1) rapid lateral facies changes (mostly metres to tens of metres in plan view);
(2) rapid vertical facies changes (mostly decimetres to metres);
(3) many small radial (along-stream) sand bodies with low connectivity, either in coarsening-upward distributary mouthbar and levee sequences over clayey prodelta deposits, or in fining-upwards channel fills (Fig. 13);
(4) evidence of frequent emergence and submergence; and
(5) low average sedimentation rates.

7.1. Recent analogues

For a better understanding of the significance of the small-scale stratigraphy of the modern Volga delta deposits, it is useful to compare it with other well-documented river-dominated deltas, especially the Mississippi delta (Fisk, 1961; Frazier, 1967; Gould, 1970).

The present-day birdfoot delta differs in essential aspects from the Volga. In the first place, there are only four large passes in the Mississippi birdfoot, each up to hundreds of metres in width. Bar finger sands, disposed radially in the delta through channel progradation, are up to 40 km long, up to 10 km wide and up to 70 m thick. At the outlet itself, the top of the distributary mouthbar sand body is being slightly reworked.

This distribution of sand-rich facies in the Mississippi birdfoot delta results from the fact that all sediments are concentrated in a single channel with four main distributaries, while in the Volga delta the sediment is spread evenly among over 800 small outlets. In addition, the Mississippi delta builds over its own highly unconsolidated prodelta sediments, leading to considerable compaction, and dewatering and sediment deformation, resulting in rapid creation of new accommodation space. The Volga delta, however, spreads a thin layer of sediment over well-consolidated Pleistocene deposits that do not allow for much additional compaction.

In the Volga delta, the offshore gradient is so shallow that waves action is minimal and little redistribution of sediments takes place, so that even the thinnest levees are being preserved. In this way, a large delta can be characterised by small-scale sedimentary architecture.

If recent analogues are to be found of the present-day Volga lower delta plain, it is rather in small subsystems of other deltas than in whole deltas.

7.1.1. Mississippi bay fill cycles (Coleman et al., 1998; Coleman and Roberts, 1989)

Between the major distributaries of the modern Mississippi birdfoot delta, bay fills such as the Cubits Gap subdelta form as a break in a major distributary bank during flood conditions. Increasing flow through this break during successive floods reaches a peak of maximum deposition and then wanes and becomes inactive within a period of 100–150 years. At its

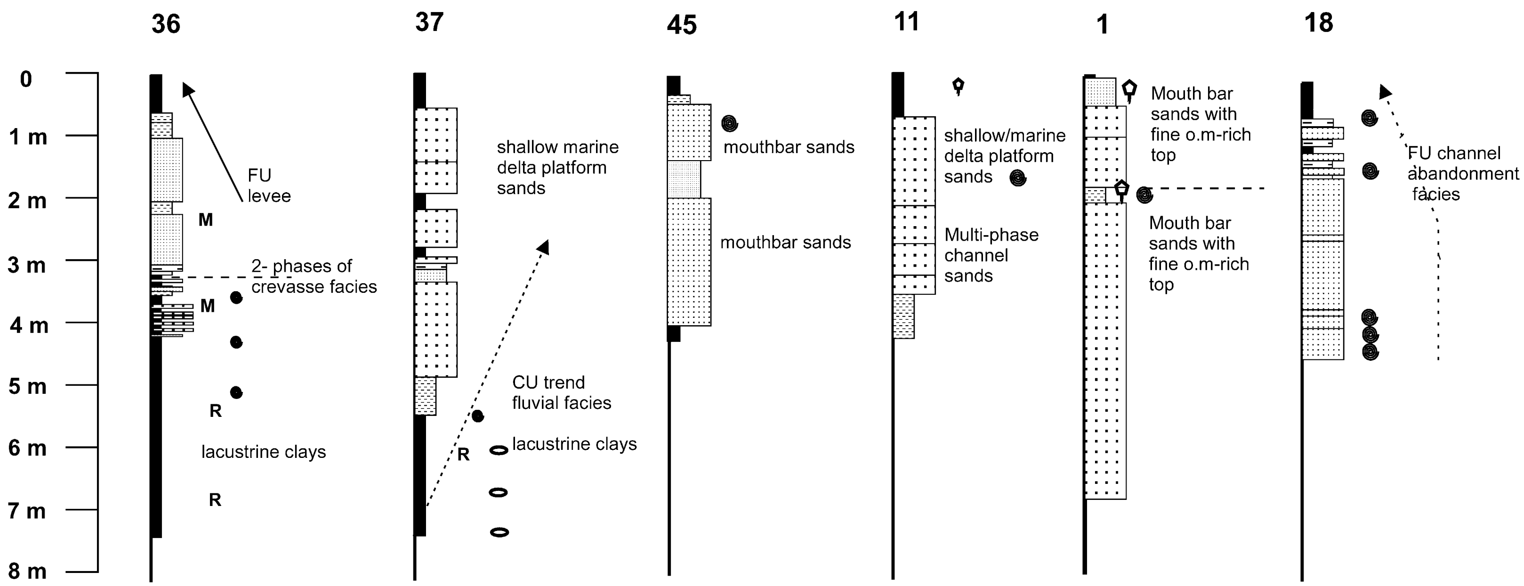


Fig. 13. Typical facies in the Holocene Volga deposits as occurred in the augerings:

—FU levees (36) and channel fills (18);
—CU mouth bars (45) and multiphase channel sands (11) forming thicker sand bodies;
—interbedded crevasse deposits (36); and
—lacustrine clays with shell and ripening horizons (36 and 37).

The symbols used as in Fig. 9.

maximum stage of development, the subdelta has an area of several hundreds of square kilometers and shows a maze of small-scale rapidly bifurcating channels present in the subdelta at the highest stage. Degradation starts upon closure of the breach due to sediment starvation and the combined action of compaction and waves. In contrast to the Volga delta, however, these bay fills build out in comparatively deep water, and they are a better small-scale analogue of deltas with a shelf break than of ramp deltas such as the Volga delta. Moreover, the role of compaction in the Volga is insignificant.

7.1.2. Mississippi shoal water deltas

Except for the modern birdfoot delta, all the deltas exposed on the Mississippi delta plain were formed by prograding into shallow inner shelf areas, the so-called shoal water deltas, such as the Lafourche delta (Fisk, 1955; Gould, 1970). Also here, a great number of bifurcating tributaries are present, but marine sediment reworking at the delta front is strong enough to form a continuous delta front sand sheet commonly 6–15 m in thickness, and progressively overlain by delta marsh deposits as the shoal water delta progrades (Fisk, 1955; Coleman et al., 1998). Associated channel fill sands are up to 30 m in thickness, all an order of magnitude greater than the Volga delta. Marine reworking, e.g., during hurricanes (Coleman et al., 1998), precludes the formation of the finely layered stratigraphy we see in the Volga delta.

7.1.3. Mississippi lacustrine deltas

A third Mississippi subenvironment of interest is the Atchafalaya lacustrine delta system further upstream (Tye and Coleman, 1989a,b), which is the result of the initial delta development after a river diversion (Roberts, 1997, 1998). Short-lived shallow lakes and basins between the distributaries in the lower floodplain and upper deltaplain are being filled up rapidly by flat-bottomed prograding deltas as soon as part of a larger distributary is diverted into them by avulsion. The average water depth is only 2–3 m, there is no tidal influence and the basins are essentially fresh water lakes. Channels, levees and bifurcation dynamics are in the same order of magnitude as in the Volga delta. The coarsening-upwards deltaic sequences recorded by Tye and Coleman (1989a,b) have been deposited since 1917 with a progradation rate of 1.1–2.0 km/year, reaching an average thickness of 3 m. These sequences greatly resemble the depositional patterns in the Volga delta, though accommodation space in the Atchafalaya basin is created by subsidence rather than sea-level fluctuations. It seems that the Volga delta is so frequently disrupted that its deposits mainly reflect this initial phase of the delta cycle.

7.1.4. Saskatchewan inland deltas (Morozova and Smith, 2000)

Lacustrine deltas are also an important component of the 8000-km^2 Cumberland Marshes, which form where the Saskatchewan river experiences a sudden gradient reduction upon entering the former glacial Lake Agassiz. Progradational coarsening-upwards sequences 2–3 m thick are an essential component of these ‘avulsion belt deposits.’ During the Holocene, nine avulsions took place, the last one of which occurred in 1870 (Smith and Pérez-Arlucea, 1994; Smith et al., 1998; Pérez-Arlucea and Smith, 1999). The facies patterns show strong resemblance to those of the Volga delta, although in the latter, peat is generally absent.

7.1.5. Okavango fan (or delta) (McCarthy et al., 1991, 1992; Stanistreet et al., 1993)

The Okavango fan in Botswana is subdivided into an entry corridor, an upper fan characterized by meander belts diverging from the apex and peats, the middle fan with highly confined single and anastomosing low-sinuosity channels and peats and a lower fan in which annual floods form relatively unconfined channels (Stanistreet and McCarthy, 1993). It resembles the Volga delta in size, in its extremely low gradient (0.00036) and in the important role of vegetation in stabilizing the positions of the channels. However, the sedimentary successions are hardly comparable as the Okavango sediment consists mostly of reworked eolian sand, while the only fines are macerated organic matter derived from the peats in the fan. Furthermore, there is no sea-level influence.

From the given examples, it appears that the depositional patterns in the Volga delta most closely resemble those of lacustrine deltas in shallow lakes of larger deltas and river systems, especially the progradational delta systems in the Atchafalaya basin and in the Saskatchewan Cumberland Marshes. Both analogues are characterized by shallow gradients, rapid progradation, high bifurcation rates and deposition of progradational coarsening-upwards sequences, with little continuity along depositional strike and more continuity along depositional dip. Both systems lack marine or lacustrine reworking.

However, there is a major difference in the sequence of events leading to the analogues mentioned above and the Volga delta. The Atchafalaya basin and Saskatchewan lacustrine deltas start to develop after a major avulsion has occurred. Sea-level rise, rise of floodplain lake levels and subsidence all promote avulsions by diminishing the gradient of the streams, and hence render a shortcut to the base level more probable (Jones and Schumm, 1999). This avulsion process has also been demonstrated in larger delta systems during Holocene sea-level rise such as the Rhine-Meuse delta (Törnqvist, 1993) and in the Mississippi itself (Törnqvist et al., 1996; Aslan and Autin, 1999).

The sequence of events in the modern Volga delta is different: progradation started here as a result of sea-level *fall* (forced regression) since the 1930s (Fig. 5B–E), whereas sea-level rise only led to consolidation of the depositional framework. The specific conditions that enabled progradation during forced regression in the Volga delta are an offshore gradient identical to the onshore gradient and sufficient sediment supply, so that no downcutting of channel mouths will occur. As long as sea level does not drop beyond the Mangyshlak sill at the −34 m isobath, the delta will behave as a ramp delta. The ensuing 3-m sea-level rise since 1977 was apparently not large enough and too short-lived to cause sufficient ponding and gradient reduction to lead to avulsion further upstream.

The sequence of events described does not invalidate the possibility that during a phase of more continuous and larger-amplitude Caspian sea-level rise, avulsion processes like those described in other environments occur. After all, remains of older Volga deltas further upstream in the North Caspian Plain (Kroonenberg et al., 1997) suggest that the present delta might be completely overstepped during a later major transgression. It remains a challenge to distinguish progradational sequences related to avulsion from those originated by forced regression as in the Volga delta.

8. Conclusions

During the last sea-level cycle (1929–1995), the Volga delta formed a progradational coarsening-upwards cycle by deposition of along-stream levees during forced regression, and filling in of the basins between them during sea-level rise. The absence of any marine reworking due to the extremely gentle gradient determines the radial distribution and preservation of the levees and channel sand bodies and their small scale in comparison with that of other river-dominated deltas such as the Mississippi delta.

Similar sequences occur in small-scale lacustrine deltas of other fluvial subsystems such as the Atchafalaya and Saskatchewan basins, but differ from them in extent and timing. While the latter deltas are relatively small subsystems of larger rivers, the small-scale stratigraphy is characteristic of the Volga delta as a whole. The Volga delta seems to remain in the initial stage of the ‘delta cycle’ (in the sense of Roberts, 1997, 1998). Moreover, in the lacustrine deltas, progradation is related to avulsion generally associated with sea-level rise, lake-level rise and/or subsidence, while in the modern Volga delta, progradation occurred during forced regression.

The small-scale Holocene stratigraphy of the present-day Volga delta shows that a similar network of fine distributaries was present during several episodes within the last 6000 years. This corroborates the conclusions of Rychagov (1993a,b, 1997) that Caspian Sea level oscillated largely around −25 m during the whole Holocene. Nevertheless, the Holocene stratigraphy of the Volga delta shows no direct relation with the present configuration of distributaries. Apparently, sea-level changes within the Holocene were important enough to temporarily disrupt the deltaic drainage pattern during extreme highstands or lowstands, leading to a complete reorganisation each time the −25 m-datum level was again approached. The details of these events have to be elucidated in the future.

Summarising, the small-scale stratigraphy as encountered in the Volga delta might be characteristic for ramp deltas on very gentle offshore slopes. This means that large ramp deltas of this type might be easily overlooked in the geological record, especially if they have been deposited in closed basins with high-frequency (fifth order) sea-level regimes such as the Caspian basin.

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