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Morphological Changes on Russian Coasts under Rapid Sea-Level Changes: Examples from the Holocene History and Implications for the Future[1]

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ABSTRACT

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Examples from the depositional sand coasts of the White Sea and the Sea of Japan provide valuable information regarding the importance of sediment budget and the direction and rates of sea-level change for patterns of coastal evolution. These examples demonstrate a limited applicability of the Bruun Rule and its modifications to the prediction of shoreline movement under the sea-level change. A moderate underwater coastal slope and an excessive or insufficient sediment supply may result in the prevalence of deposition during sea-level rise and erosion during its fall. In general, the faster sea-level rise, the higher the possibility of burial, drowning, or destruction of the coastal depositional body. The faster the sea-level fall, the more probable the preservation of depositional bodies above the retreating sea, e.g., in the form of beach ridges and coastal dunes.

As a first approximation, a model of coastal development under accelerating sea-level rise is established for the conditions of excessive and insufficient sediment supply on sand coasts. Under the former, a moderate acceleration of sea-level rise causes the change from mobilization of sediments at a beach face and formation of a beach ridge to the landward translation of the coastal depositional body and, then, to its transformation. An extreme acceleration causes burial of the coastal depositional body by a transgressive sedimentary sequence. Under the latter, mobilization of existing scarce sediments results in a landward movement of a depositional body, erosion of its seaward slope, drowning, and partial destruction. The extreme acceleration may bring, in some cases, the total grading of the coastal zone profile.

ADDITIONAL INDEX WORDS: *Depositional coasts, coastal responses, Bruun Rule, rate of sea-level rise, sediment supply, White Sea, Sea of Japan.*

INTRODUCTION

The problem of a greenhouse-induced acceleration in global sea-level during the next century and the possible rate of this process still remains to be solved. The Intergovernmental Panel on Climate Change agreed unanimously on the 0.31–1.10 m rise in global sea level between 1990 and 2100 (HOUGHTON *et al.*, 1990). Slightly lower estimates have been derived from the revised emissions scenarios (WIGLEY and RAPER, 1992). Global sea-level rise over one meter is, therefore, not likely, but cannot be totally excluded yet. Nevertheless, recent estimates have suggested that global mean sea-level rise during the last century was generally <0.20 m (PIRAZZOLI, 1986; GORNITZ and LEBEDEFF, 1987; DOUGLAS and HERBRECHSMEIER, 1989; PELTIER and TUSHINGHAM, 1989; KLIGE, 1990; EMERY and AUBREY, 1991). It has been suggested that global mean sea-level rise during the last century was statistically insignificant (DOBROVOLSKII, 1992). Thus, a strong possibility exists that the future rate of global sea-level rise will be unprecedented when compared with those recorded during the period of direct observations and measurements on sea coasts.

It is clear that the use of various indirect data on coastal change under rapid sea-level variations in the past is an urgent necessity. An analysis of morphological and sedimentary sequences in coastal zones under various regimes of relative sea-level change, sediment supply, etc. represents an important source of such information. A broad variety of examples of coastal responses to sea-level changes during the last deglaciation and relatively short-term and low-amplitude sea-level fluctuations of the last millennia may be derived from studies of the Russian coastal zone.

Many Russian and former Soviet writers have analyzed the pattern of sea-level changes following the decay of the last continental ice sheets and examined the coastal morphological and sedimentological responses to these changes. These studies have revealed the spasmodic, site-specific character of the lateglacial and postglacial rise in sea level on the coasts

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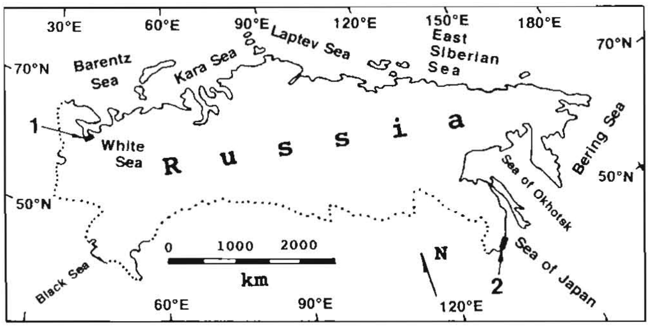

Figure 1. General position of the sites described in the text on the map of Russia: (1) the White Sea, Onega Bay; (2) Middle Primorye, mainland coast of the Sea of Japan. International boundary of Russia is shown by dotted line.

of Russia and the former USSR, and the general climatic dependence of sea-level fluctuations during the Holocene (SELIVANOV, 1995). At times, the rate of the rise in relative sea level was >10 mm/year, *i.e.,* comparable to or exceeding the anticipated rates of the future greenhouse-induced sea-level rise during the next century. Therefore, study of the lateglacial and postglacial shorelines on the shelf and coast of the Russian seas provides valuable evidence regarding coastal responses to rapid changes in relative sea level.

Sandy shorelines are extremely sensitive to the changes in water level and sedimentary budget and represent an important example of coastal responses to sea-level changes. In this paper, the response patterns of such shorelines are analyzed with reference to the depositional coasts of the White Sea and the Sea of Japan (Figure 1).

EXAMPLES OF COASTAL MORPHOLOGICAL RESPONSES TO RAPID POSTGLACIAL SEA-LEVEL CHANGES IN RUSSIA

The White Sea, Onega Bay

Description of Morphological and Sedimentary Responses

The shallow-water Onega Bay in the south-east of the White Sea is characterized by a moderate underwater coastal slope (0.1–0.5%), relatively low wave influence (wave height reaches 1.0–1.1 m once a year and 1.4–1.5 m once in 10 years), mean storm surges (nearly 1.5 m once a year and 2.0 m once in 10 years in the head of the bay) and mesotidal conditions (spring tidal range of 2.7–3.2 m in the head of the bay).

A series of Late Pleistocene and Holocene limnoglacial, glaciomarine, and marine terraces dominates the coastal morphology. Traditionally, terraces were identified at elevations of up to 60–80 m (LAVROVA, 1969; DEVYATOVA, 1976). However, field observations by the present author, as well as some other investigators during the last decade, do not prove the existence of terraces above 20 m (see KAPLIN and SELIVANOV, 1995A, for a more extensive review of the problem).

In coastal bluffs in the east and south-east corner of the White Sea, where sediment input from rivers and eroding coastal segments was high, boulder tills of the last glaciation are usually covered at 10–18 m above the present sea level by laminated lateglacial grey clays (SELIVANOV, 1984), thus indicating a prevalence of depositional processes during that time. In coastal segments with less intensive sediment supply, an erosion of glacial tills during the lateglacial period is marked by erosional escarpments up to 16–20, rarely 25–30 m, above the present sea level.

In the Onega Bay, a terrace at 10–15 m above the present sea level can be traced along the coast (Figure 2) but has not been dated yet (SELIVANOV, 1984; MATOSHKO *et al.,* 1989). End moraines, dated at 13,500–13,000 BP (ZUBAKOV and BORZENKOVA, 1990), consist of boulder clays that outcrop in bluffs along the 10–15 meter terrace. These sediments are overlain by residual boulder and pebble causeways which indicate an erosion of tills from this site during the drainage of the periglacial lake in the Onega Bay.

The lowest water level recorded by the ancient shoreline at −25 to −27 m on the bottom of the bay was reached nearly 9,800–9,500 yr B.P (Figure 3). The subsequent period of the fluctuating sea-level rise resulted in the formation of submerged shorelines at −2 and −12 to −15 m dated at from 7,980 ± 270 and 7,570 ± 250 yr BP respectively (SELIVANOV, 1984), and an emerged depositional terrace at 6–10 m above present sea level.

In embayments, sections of this terrace usually consist of two transgressive sequences of sand and silt facies of the upper shelf, intertidal flat, beach and lagoon. The tops of both depositional sequences are composed of lagoon silts related to the highest relative sea-level during the corresponding fluctuation. Therefore, the low-amplitude falls in relative sea level at the end of these fluctuations resulted in erosion or non-deposition. Regressive sequences are represented only by peats. The older one is dated at 4,800 ± 180, 4,100 ± 150 yr BP (BOYARSKAYA *et al.,* 1986) and 4,030 ± 70 yr BP (KOSHECHKIN *et al.,* 1977). The sea-level position during the two lowstands in the Middle and Late Holocene have been estimated at −1 to −3 m (KAPLIN and SELIVANOV, 1995A).

On the open coast of the bay, the 6–10 m terrace is represented by two series of sand-beach ridges and coastal dunes which overlie the coarse sands and gravels of ancient beaches at altitudes of 8–10 m and 4–5 m (Figure 2). Crest altitudes in both series of beach ridges decrease seaward. The seaward limit of each series is marked by a typical coastal dune which is 2–3 m above the adjacent ridge crests. Behind the higher sequence of beach ridges, lagoon sandy silts and muds outcrop on the surface. They are dated at 6,455 ± 80, 5,940 ± 250, 5,600 ± 250, and 5,240 ± 200 yr BP (BOYARSKAYA *et al.,* 1986). The lower beach ridges (4–5 m) may be tentatively dated at 4,500–4,000 yr BP by the presence of a Late Neolithic archeological site found in its upper sediments (SELIVANOV, 1984). The subsequent period of sea-level fall to −1 m is dated by archeological artifacts at approximately 2,200–2,000 yr BP.

It is worth noting that the thickness of Holocene sediments in the coastal zone of the study area is relatively small and limnoglacial gravel clays are found in the present intertidal

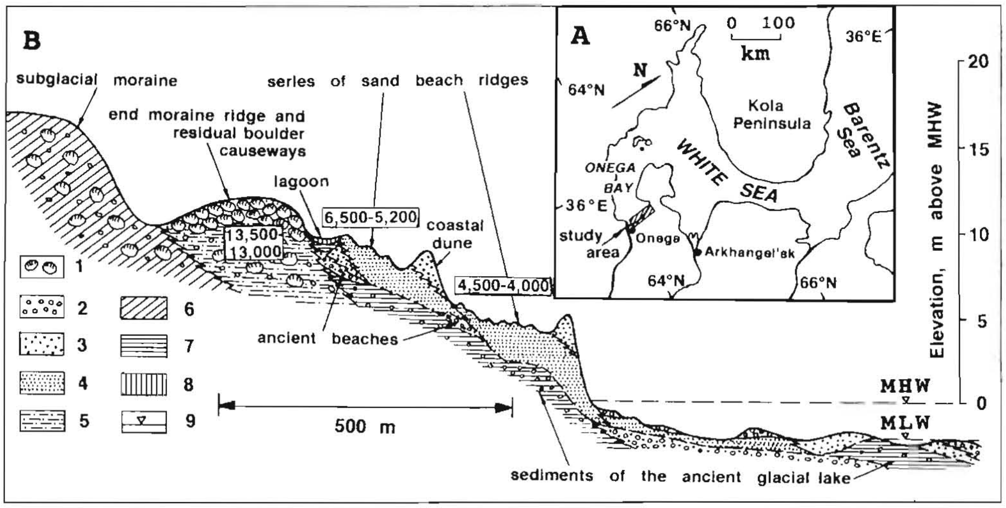

Figure 2. Position of the study area in the Onega Bay, White Sea (A) and a generalized cross section of the coastal zone (B): (1) boulders and pebbles; (2) gravel; (3) coarse and medium sand; (4) fine sand; (5) silty sand and loam; (6) silt and gyttia; (7) clay; (8) peat; (9) levels of mean high water (MHW) and mean low water (MLW). Numbers indicate the ages of depositional sequences in yr BP.

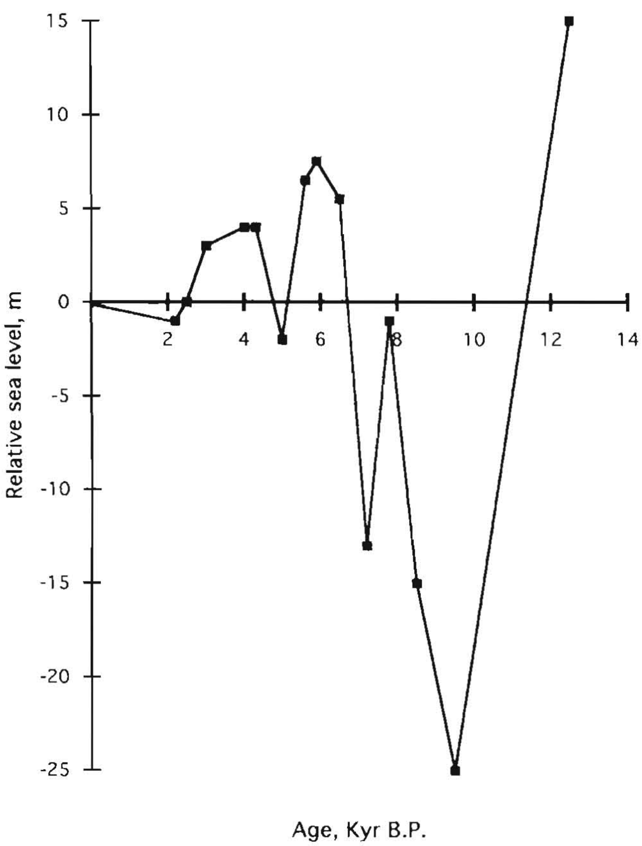

Figure 3. Possible changes in the relative sea level in the Onega Bay, the White Sea during the Lateglacial and the Holocene. Squares mark the position of dated coastal complexes.

zone beneath only 0.5–1.5 m of recent coastal sands. Coastal facies in the 6–10-meter terrace are as thin as 1.5–2.5 m. Recent erosional processes are obvious on the underwater coastal slope where lateglacial tills are found at the bottom surface (Figure 2). This indicates a generally low sediment supply to the coastal zone of the study area.

Interpretation

This example of coastal response reflects the general slow fall of the relative water level during the Lateglacial and Early Postglacial period as a result of drainage of the periglacial lake and, possibly, isostatic emergence and the subsequent sea-level fluctuations (Figure 3). General erosion of glacial and limnoglacial sediments occurred during the Lateglacial and Early Holocene, especially during water-level fall periods. The highest rate of sea-level rise between 9,500 and 6,500 yr BP resulted in the total grading of the coastal zone.

Since 6,500 yr BP, a mobilization of existing sediments in the coastal zone has caused the formation of the successive coastal depositional series under the fluctuating relative sea level. The earlier period (beach ridges at 8–10 m) was marked by the replacement of an open beach facies by a lagoonal facies due to the development of a coastal barrier. The later period (beach ridges at 4–5 m) resulted only in the formation of beach sands and gravels and, possibly, coastal erosion during the peak of the relative sea-level rise. Episodes of falling sea level caused the formation of beach ridge series with decreasing elevations and, finally, coastal dunes. Thickness and morphological clarity of these beach-ridges decreased from the first sea-level rise period to the second one. It was possibly connected with the lower rate of sea-level fall during the second period (Figure 3). The present beach and intertidal zone bear obvious features of insufficient sediment supply. This pattern reflects the general prevalence of tills as a major

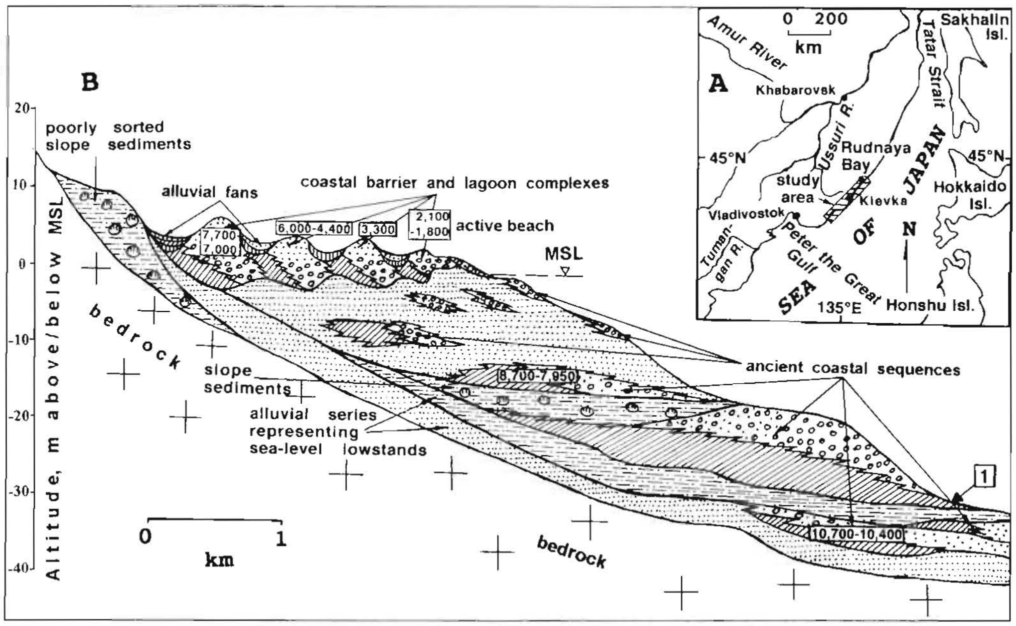

Figure 4. Position of the study area in Middle Primorye, mainland coast of the Sea of Japan (A) and a generalized cross section of the coastal zone (B). Numbers indicate the ages of coastal depositional sequences in yr BP. See Figure 2 for an explanation of the lithological signs. The exposure of lagoon sediments seaward of the respective coastal barrier is shown (1).

sedimentary source for the coastal zone and the repetitive reworking of these tills in the absence of a significant inflow of sediments during the Holocene.

Continental Coast of the Sea of Japan

Description of Morphological and Sedimentary Responses

The wave-dominated continental coast of the Sea of Japan (traditionally referred to as Soviet Primorye), is characterized by microtidal conditions (spring tidal range is 0.1–0.2 m on the open coast and 0.3–0.5 m in bays), a moderate wave activity under the mean conditions, and an extremely high morphological significance of storm waves during typhoons (wave height once a year is 4.5–5.5 m and 11–13 m once in 10 years). The shoreline is characterized by an alternation of narrow and deep ingressive bays and high erosional headlands. Well-preserved sedimentary and morphological complexes of ancient shorelines are a characteristic feature of the present inner shelf area in the bays. An underwater coastal slope in the bays is moderate (0.3–1.0%). Outside of these the bays, the inner shelf slope is covered by a thin layer of sediments, and the present coastline usually bears an erosional character.

Several coastal depositional features have been identified by bottom coring and side-scan sonar investigations in the bays (VNUCHKOV *et al.*, 1976; KRIVULIN *et al.*, 1978; BADYUKOV and KAPLIN, 1979). These features are represented by transgressive sequences from silts and silty sands, which abound in shell fragments, to lagoonal gyttja and algae peat, and to gravels and well-sorted sands of coastal ridges and dunes (Figure 4). The lower part of the sequences are often missing due to erosion. Moreover, lagoon facies may outcrop at the seaward slope of the existing sediments of the respective coastal barrier (see 1 in Figure 4). This indicates an intensive landward movement of coastal barriers.

Fine alluvial sands underlie the coastal sedimentary complexes and, in places, are intercalated with them. These sands indicate periods of sea-level lowering, as do a series of poorly sorted slope pebbles and loams at the present depth of 16–20 m below mean sea level. The prevailing depths of lagoon sequences, which may serve as the best indicators of former sea-level position, are −90 to −100, −65 to −70, −25 to −40, and −13 to −18 m. Lagoon sediments at the depth of −65 to −70 m have been dated radiocarbon-at 13,300–13,100 yr BP; those at −25 to −40 m at 10,700–10,400; and those at −13 to −18 m at 8,700–7,950 yr BP (KRIVULIN *et al.*, 1978; BADYUKOV and KAPLIN, 1979). A tentative correlation of the deepest coastal complexes at −90 to −100 m with the last glacial maximum was proposed (BADYUKOV and KAPLIN, 1979).

The coastal complexes at 25–40 and 13–18 m below the present sea level are best recorded from the Tumangan River on the Russian-Korean border and the Peter the Great Gulf in the south to the Kievka Bay and Rudnaya Pristan settlement in the northern coastal area (Figure 4). These complexes may be correlated with erosional surfaces in effusive rocks on capes (SELIVANOV and STEPANOV, 1985). These erosional features are possibly the result of several relatively long-lived periods of stable sea level during the Middle Würm.

A series of coastal barrier and lagoon sediments at 2–7 m above the present sea level constitutes another element of the coastal morphology of the region (Figure 4). Several generations of these features may be distinguished by morphological and radiocarbon data (KOROTKY *et al.*, 1980, 1989; SELIVANOV and STEPANOV, 1985). These complexes usually consist of sand ridges and/or dunes resting against peat and lagoon silt and gyttja. In several cases, narrow lagoon depressions are partially covered by overwash sands of younger ridges and filled by alluvial fine sands and loams.

Geoarcheological investigations in several bays in the middle sector of the coast (Rudnaya, Kievka, and Melkovodnaya) demonstrate the existence of three major groups of coastal ridges. The dominant elevations of the ridge crests decrease seaward from 5–6 to 3.5–4 and 2–2.5 m above sea level, whereas lagoon surfaces are situated at 3.5–4, 2–2.5 and 1–1.5 m respectively. Archeological and pollen data indicate that these complexes are related to three separate sea-level fluctuations during the Middle and Late Holocene (SELIVANOV and STEPANOV, 1985). The ages of the sea-level maxima are estimated at 7,700–7,000, 6,000–4,400 and 3,300–3,000 yr BP. Beds of lagoon facies at 0 to −2 m possibly represent the lowest sea-level positions between these phases. In the Kievka Bay and Rudnaya Bay, another coastal complex occurs near the present shoreline at the elevations of 0.5–1.5 m. However, this complex is usually distinguished only by lithological and archeological data. The latter provide an age of 2,100–1,800 yr BP for this complex.

Another sequence of the middle and late Holocene coastal barriers characterizes the Posyet Bay and the head of the Ussuriysk Bay in South Primorye (IONIN *et al.*, 1971) and the Zerkalnaya Bay in the middle sector of the coast (KRIVULIN *et al.*, 1978). Elevations of coastal ridges, composed of coarser sands, decrease landward from 3–4 to 1–1.5 m and the underlying lagoon facies lie at −0.5 to +0.5 m. This phenomenon may be tentatively explained by tectonic submergence of the bays known from tide-gauge data and intensive aeolian grading of the sand surface of the seaward coastal ridge.

At the open coastline, fluctuations of the relative sea level during the Middle and Late Holocene can be traced by a number of erosional wave-cut features at the prevailing elevations of 3.5–4 and 1.5–2 m.

Interpretation

The rapid rise of relative sea level until 8,000–7,700 yr BP resulted in a landward translation of coastal depositional bodies. Drowning and, finally, burial of successive depositional bodies by transgressive sedimentary sequences were the primary processes in coastal evolution under the fluctuating rise in the relative sea level.

Around 7,700–7,500 yr BP, this pattern of coastal evolution changed. Instead, coastal evolution since that time occurred as a "depositional regression", i.e., the formation of successive barrier bodies and the seaward displacement of the shoreline. Temporal correlation of this change in the coastal morphological regime with the deceleration of sea-level change is obvious (Figure 5).

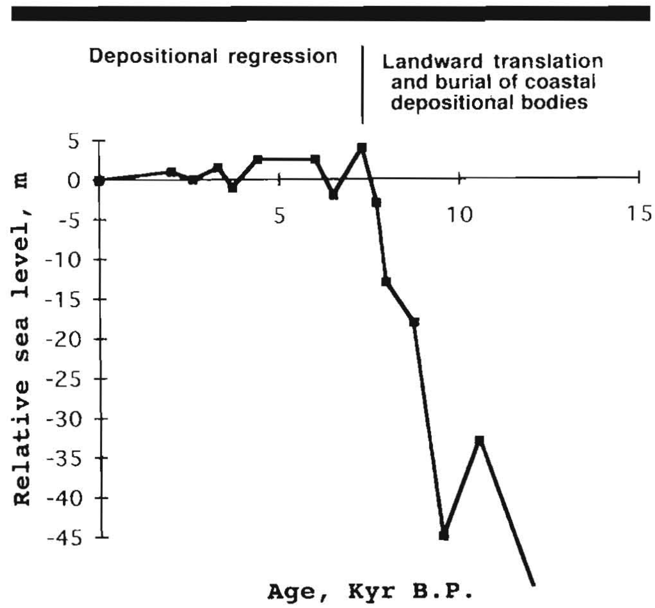

Figure 5. Possible changes in the relative sea level on the continental coast of the Sea of Japan during the Holocene. Squares mark positions of dated coastal complexes. Note that a drastic deceleration in sea-level changes from 7,700–7,500 yr BP coincided with the change in the coastal response pattern from landward translation and burial of successive depositional coastal bodies, to "depositional regression", *i.e.* formation of successive coastal barriers and seaward displacement of the shoreline.

This area represents clear example of the primary importance of the tendency and intensity of changes in the relative sea level in controlling the evolutionary pattern of the coastal zone under sufficient or excessive sediment supply. However, contrary to the Bruun Rule, depositional processes dominated on the coast during the sea-level rise periods. The variability in coastal complexes formed during the successive sea-level fluctuations of the Middle and Late Holocene from one bay to another may reflect temporal variations in sediment supply from rivers outflowing to these bays and not only changes in the rate of sea-level change.

DISCUSSION

The above examples demonstrate the over-riding importance of the tendency and rate of sea-level changes in controlling the pattern of coastal evolution. In general, the faster the sea-level rise, the higher the possibility of burial, drowning or destruction of the coastal depositional body. The faster is the sea-level fall. The preservation of depositional bodies above the retreating sea, i.e., ridges and coastal dunes, is contingent upon the rate the sea level falls.

These examples are sufficient to reject the universal applicability of a hypothesis of the exclusive seaward transport of sediments during sea-level rise and landward transport during the sea-level fall. This hypothesis has been put into the basis of the Bruun Rule and its modifications (see SCOR WORKING GROUP, 1991; SELIVANOV, 1993 for the detailed review of the problem).

Another important source of information on this problem

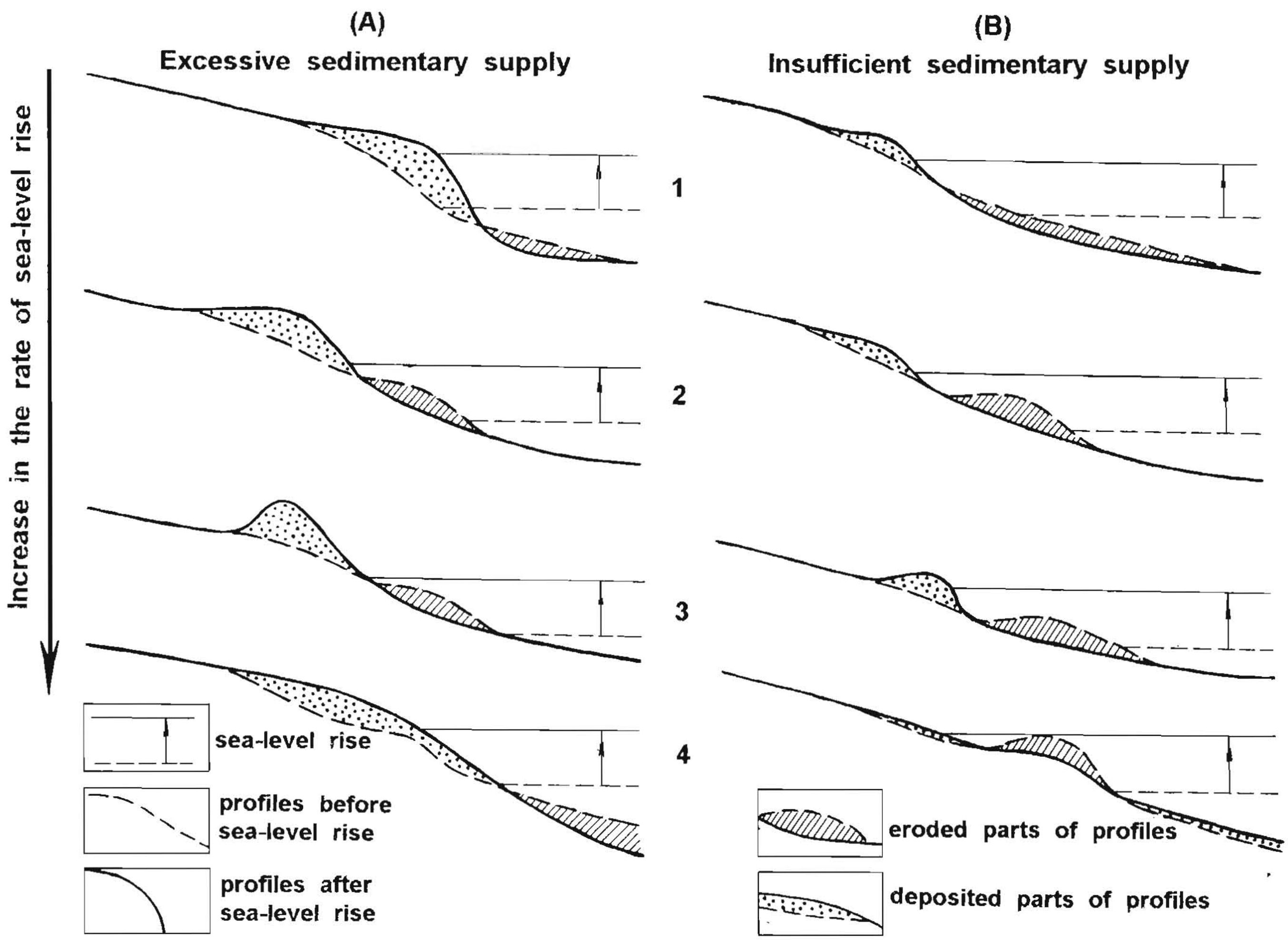

Figure 6. Patterns of coastal response to sea-level rise depending upon the rate of this process (1–4) and sediment supplies in the coastal zone: (A) excessive sediment supply; (B) insufficient sediment supply.

is provided by direct observations of coastal changes in the Caspian Sea and other large enclosed lakes in Central Asia, which have experienced rapid changes in water level of several meters during recent decades (KAPLIN, 1989; IGNATOV *et al.*, 1993; KAPLIN and SELIVANOV, 1995B). These studies revealed a strong link between the pattern of coastal response and the gradient of the underwater coastal slope. Only the steepest coastal slopes (usually over 0.01 for medium sands) followed a pattern of coastal evolution in accord with the Bruun Rule. For other sites, which had ample sediment supply, coastal progradation occurred.

Of crucial importance for the coastal evolution under sea-level changes is a sediment budget of the coastal zone. This fact has been stressed by various authors (CURRAY, 1964; THOM, 1984; SWIFT *et al.*, 1985; CARTER *et al.*, 1987). The present author believes that both excessive and insufficient sediment supply may result in a dominantly landward movement of sediment in the coastal zone under sea-level rise. A moderate inclination of the underwater coastal slope is possibly another precondition for such a response.

As a first approximation, the following two coastal responses to an accelerating sea-level rise may be proposed for the conditions of substantially excessive and insufficient sediment supply respectively (Figure 6). Under excessive sediment supply on a graded coastal profile (Figure 6A), slow sea-level rise causes the mobilization of sediments at a beach face and the formation of a beach ridge (1). A moderately accelerating sea-level rise results usually in a landward translation of a coastal depositional body by overwash processes. No significant transformation of coastal morphology occurs (2). Faster sea-level rise results in a transformation of a depositional body, namely an increase of its elevation and steepening of a landward slope (3). An extreme acceleration causes burial of the coastal depositional body by a transgressive sedimentary sequence (4).

Under insufficient sediment supply (Figure 6B), mobilization of existing scarce sediments on a primary graded coastal slope results in the formation of a poorly expressed depositional body (1). With an accelerating sea-level rise, this body moves landward in a translational manner (2) and undergoes erosion of its seaward slope (3). The last process causes partial or total destruction of a coastal body under the very rapid sea-level rise (4). The extreme acceleration may bring, in some cases, the total grading of the coastal zone profile.

The pattern distinguished for the conditions of the excessive sediment supply (Figure 6A) may be interpreted in the terms of quasi-equilibrium evolution under a relatively slow sea-level rise (2). With the accelerating sea-level rise, possible intensity of coastal reformation becomes insufficient to keep pace with the rising sea level (3). Disequilibrium becomes total under the extremely fast sea-level rise (4). Following the terminology proposed for evolution of coral reefs under the sea-level rise at various rates (NEUMANN and MACINTYRE, 1985, SPENCER, 1993), the response patterns (2–4) for the conditions of excessive sediment supply (Figure 6A) may be denominated as keep-up, catch-up and give-up respectively.

CONCLUSIONS

An analysis of the morphological and sedimentary sequences in coastal zones under various regimes of relative sea-level change and sediment supply may provide valuable information for the prediction of coastal evolution under the anticipated future acceleration of sea-level rise.

The examples from two Russian seas demonstrate the primary importance of the tendency and rates of sea-level changes on the pattern of coastal response. In general, an acceleration of sea-level rise results in the burial, drowning, or destruction of a coastal depositional body. An acceleration of sea-level fall usually results in the preservation of depositional bodies above the retreating sea as beach ridges, coastal dunes, etc.

These examples demonstrate the limited applicability of the Bruun Rule and its modifications, based upon the assumptions of the equilibrium development of coastal morphology during sea-level changes, and the exclusively seaward transport of sediments during sea-level rise and landward transport during the sea-level fall.

The sediment budget of a coastal segment may be the dominant factor in coastal evolution under sea-level changes, notably in cases where there is either heavily excessive or insufficient supply of sediment. Both situations may result in a dominantly landward movement of sediments. Moderate inclination of the underwater coastal slope is possibly another precondition for such a response.

A series of evolutionary patterns under various rates of sea-level rise has been established for the conditions of excessive and insufficient sediment supply on sand coasts (Figure 6). An acceleration of sea-level rise causes the change from the mobilization of sediments at a beach face and the formation of a beach ridge to the landward translation of a coastal depositional body and, then, to its burial by a transgressive sedimentary sequence under excessive sediment supply or erosion and destruction under insufficient supply. These patterns may be interpreted in terms of a turn from quasi-equilibrium to disequilibrium evolutionary patterns.

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