NEARSHORE MORPHODYNAMICS AND COASTAL EROSION

I.O.Leont'yev

P.P.Shirshov Institute of Oceanology, Russian Academy of Sciences,
Nakhimovskiy Prospect 36, 117997, Moscow, Russian Federation
Tel.: +7-095-124-63-94; fax: +7-095-124-59-83
e-mail: leontev@geo.sio.rssi.ru

Introduction

The sea coast or in more extensive comprehension, coastal zone is a border area between the sea and the land. Sea waves and currents interact here with the surface of coastal slope and accomplish the work forming the shore face and the bed relief.

Result of this work depends on a number of natural factors. Amount of incoming energy (strength and frequency of storms), solidity and composition of rocks constituting the coast, thickness of deposits on fundamental substrate, bed slopes and climate of a region are in particular of great importance. Various combinations of named factors determine the multiplicity of existing types of coasts. Some coasts are composed of solid rocks and actually were not changed by the sea. In the warm tropic seas the coral-built coasts are common, whereas in the Arctic regions where the permafrost is widespread, the thermal-abrasion and ice-built coasts occur.

However further we shall mainly take an interest in sedimentary coasts composed of sands. Sand particles are easily mobilized under the action of waves and transported by the nearshore currents in a certain direction. Sediment fluxes created result in an active morphodynamic processes, i.e. changes in the bed relief and the coastline position. Many sandy coasts are developed by the human for a long time and now used as the objects of economical activity and as the recreation centers. Sandy beaches bordered by a dune belt from land and by a strip of surf from sea side exert beneficial effect on man's emotional health bringing contentment and feeling of harmony (Fig.1).

Coastal zone is a complex system tending to equilibrium with environmental conditions. Rapid changes in those are able to disturb delicate balance and result in coast degradation. At present time the most of coasts manifest the trend to erosion and recession under the rush of the sea (about 70 % by Bird's [1985] estimate). Man's impact plays in this the role of no small importance. Constructing of ports, navigation channels and other structures as well as the sand dredging on the sea bed often lead to perturbations in sediment fluxes provoking material deficit in certain coastal sections. However there are also the global causes of erosion, in particular, the gradual rise in the ocean level (about 15 cm for last 100 years) and tectonic subsidence of some land areas (up to several tens cm per century).

Fig.1. Finnish Bay coast (the Baltic Sea) in the region of Zelenogorsk

Thus many sea coasts are needed in protection of which the strategy should be based on forecast of the future behavior of coast. Reliable forecast implies an appropriate level of knowledge of processes and mechanisms behind the nearshore water circulation, sediment transport and morphological response. Further we shall try to outline existing concepts on some important aspects of nearshore morphodynamics and characterize possible approaches to predicting coastal evolution over the nearest centuries.

Nearshore circulation and sediment transport

The most of energy coming to the open sea coast is delivered by the gravitational surface waves excited by wind. Wave propagation on relatively shallow water (where depths are lesser than half wave length) is accompanied by the water particle motions along elliptic orbits (Fig.2). At the bottom these motions transform into the to-and-fro excursions along the bed. When orbital velocities reach 0.2-0.3 m s^{-1} the sand particles are mobilized and involved into the flow oscillations. These oscillations are asymmetric as the flow velocities under crests of shallow-water waves are greater than backward velocities under the troughs. Besides, in bottom boundary layer the net flow in direction of wave propagation arises resulting in sediment transport toward the coast (Fig.3). This income of material from relatively deep section of bed can play important role in sediment balance of coastal zone.

As the depths decrease toward the shore the wave energy concentrates in a smaller volume of water, what provokes the increase in flow velocities. When the depth becomes comparable with the wave height the waves collapse. Intensive turbulence generated in breaking waves lifts the sediment masses into the water column. Roller arising in front of wave crest dissipates the energy and so the waves decay approaching to coast. At the shoreline the flow takes form of the swash – oscillations of tongue of water along the beach slope.

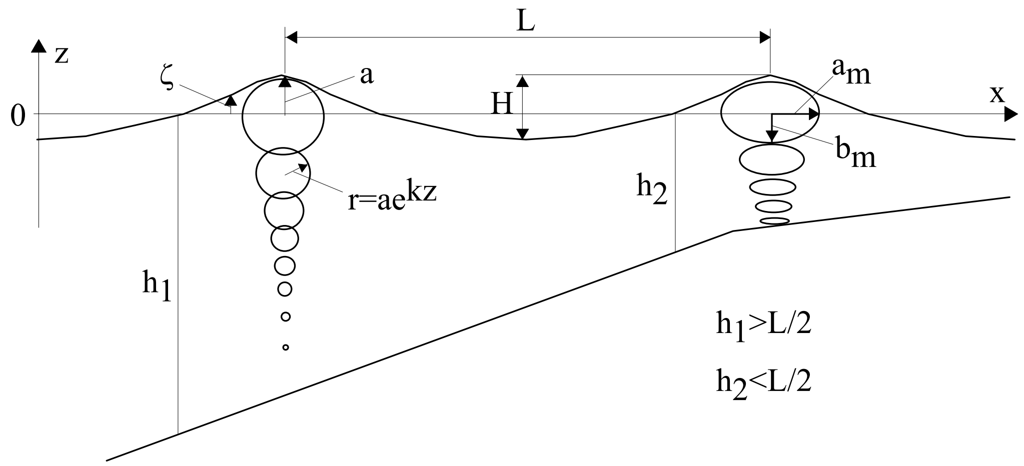


Fig.2. Wave-induced orbital motions at depths greater and lesser than half wave length

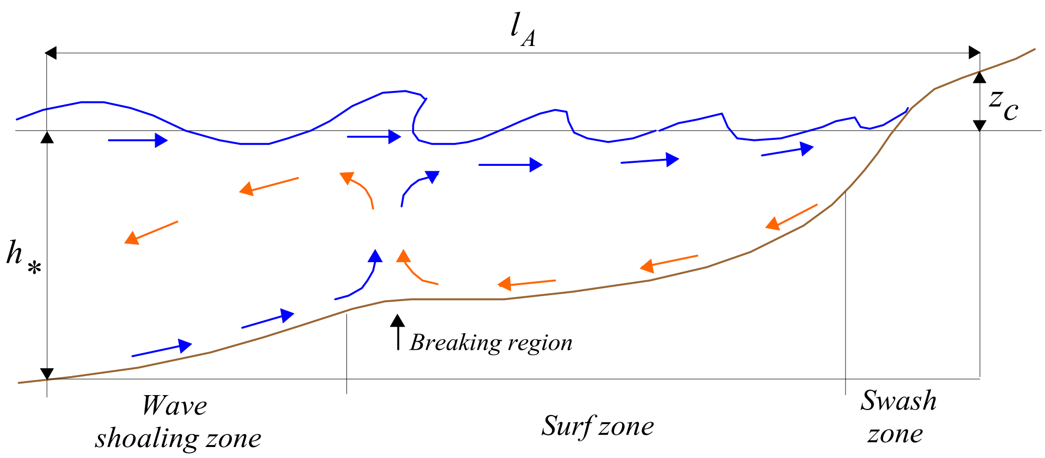


Fig.3. Scheme of wave-induced circulation in the water column of the coastal zone

Certain portion of energy lost by waves in the surf zone transforms into the energy of wave setup (rise of water level) and various wave-induced currents. Especial role of the undertow and the longshore current should be emphasized. Undertow is a seaward flow with 0.2-0.5 m s^{-1} velocities balancing the landward mass transport caused by broken wave crests and carrying suspended material away from coast (Fig.3). Near the break point an intensive ascending motions of water and suspension occur and so a portion of sediments can be involved into the landward movement partly compensating the lost of beach material.

Longshore current is generated by obliquely incident waves and reaches the velocities more than 1 m s^{-1}. It is a common case when waves approach at some angle relatively to the shore and so the longshore current actually always exists during a storm. This current creates the sediment flux along the coast. Decrease in flux produces accumulation of sediments while its increase leads to bed erosion. Thus the longshore flux gradient is essential component of sediment balance in the nearshore region.

If the bottom topography and heights of incident waves are non-uniform along the shore then the local horizontal circulations can develop (Fig.4). Those superimpose on the vertical-plane circulation considered before and produce more complex flow patterns. The outflow from shore in some cases takes the form of rip current (with velocities are of the order of 1 m s^{-1}) concentrating in sections of relatively low waves and working out erosion trough – a rip channel.

Together with gravity waves mentioned above also the infragravity waves (with periods exceeding 20-30 s) of different genesis occur in a coastal zone. Those influence the nearshore currents, sediment transport and morphological response. Infragravity waves, in particular, contribute to forming of beach cusps – rhythmic shoreline undulations with spatial step of several meters to hundred meters. Comparatively small beach cusps can be observed on the lower photo in Fig.1.

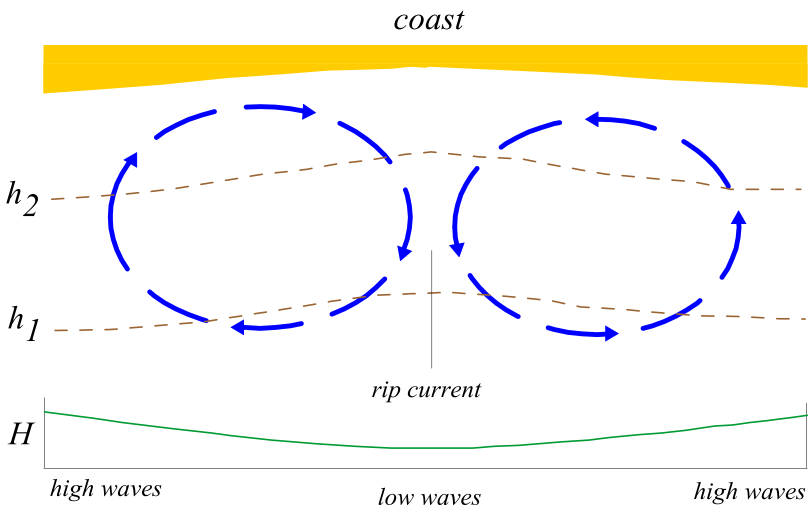


Fig.4. Circulation cells due to longshore variations in the wave field

Under certain conditions behavior of coast can be mainly controlled by tides. This is a typical situation for closed estuarine coasts where the amplitudes of oscillations of tidal level and associated flow velocities are high while the wave activity is weak. Outside the coastal zone tidal currents can form the large-scale sand waves on the sea bed. Slowly moving sand waves may result in bed deformations of magnitude of several meters. Also the undulations in bottom relief influence the properties of incident wave field.

Finally, an important role of wind in coastal dynamics should be emphasized. Combined action of wind and waves creates a storm surge, which is particularly dangerous for gently sloping low coasts where a vast land area can be flooded. Storm surge also results in appearance of outflow transporting sediment particles away from coast. Under conditions of relatively short storm waves this outflow put obstacles in the way of sediment movement toward the shore and thus works as a mechanism maintaining coastal erosion [Leont'yev, 2003a]. Besides, the wind blowing from sea causes Aeolian deflation – the lost of beach and dune material – influencing the sediment balance in coastal zone.

Storm-induced bed changes and equilibrium coastal profile

One of the fundamental statements of coastal geomorphology is that the coastal profile in process of evolution tends to adjust its form to a stable equilibrium shape attributable to influencing dynamical forcing [Zenkovich, 1967, Bruun, 1988]. That is also true for distinct wave situations what is confirmed by numerous field observations and laboratory tests. Deformations of coastal profile are most pronounced in the initial stage of wave attack and then gradually decay and almost cease if the wave parameters remain uniform during sufficiently long period of time (Fig. 5 and 6).

Time lapse needed for profile to achieve a quasi-equilibrium state depends on spatial dimensions of coastal zone and in many cases is comparable with duration of stabilization phase of actual storm.

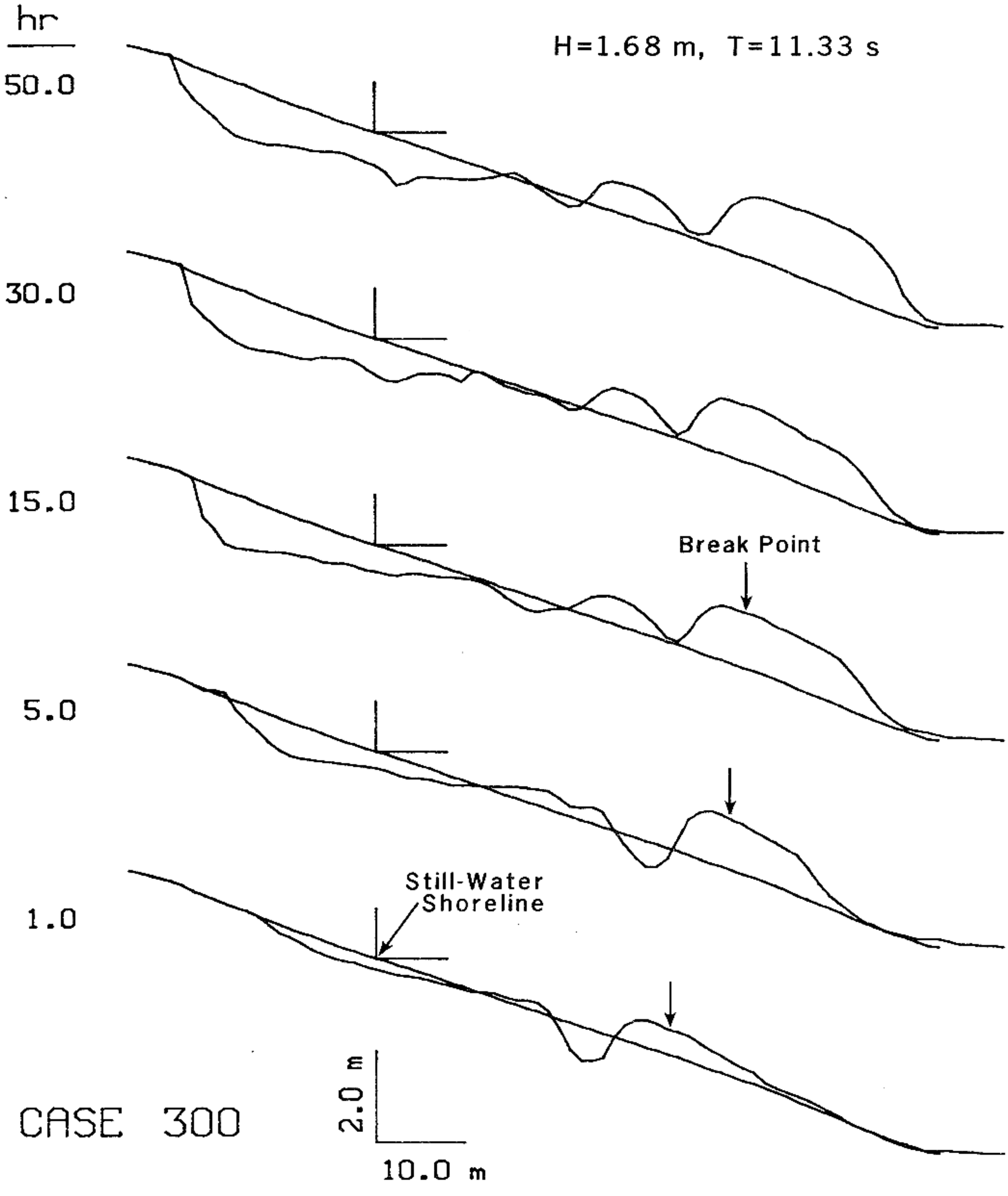


Fig.5. Deformations of sand-composed coastal slope attacked by steep stormy waves. Data from test C300 in large wave flume [Larson and Kraus, 1989]

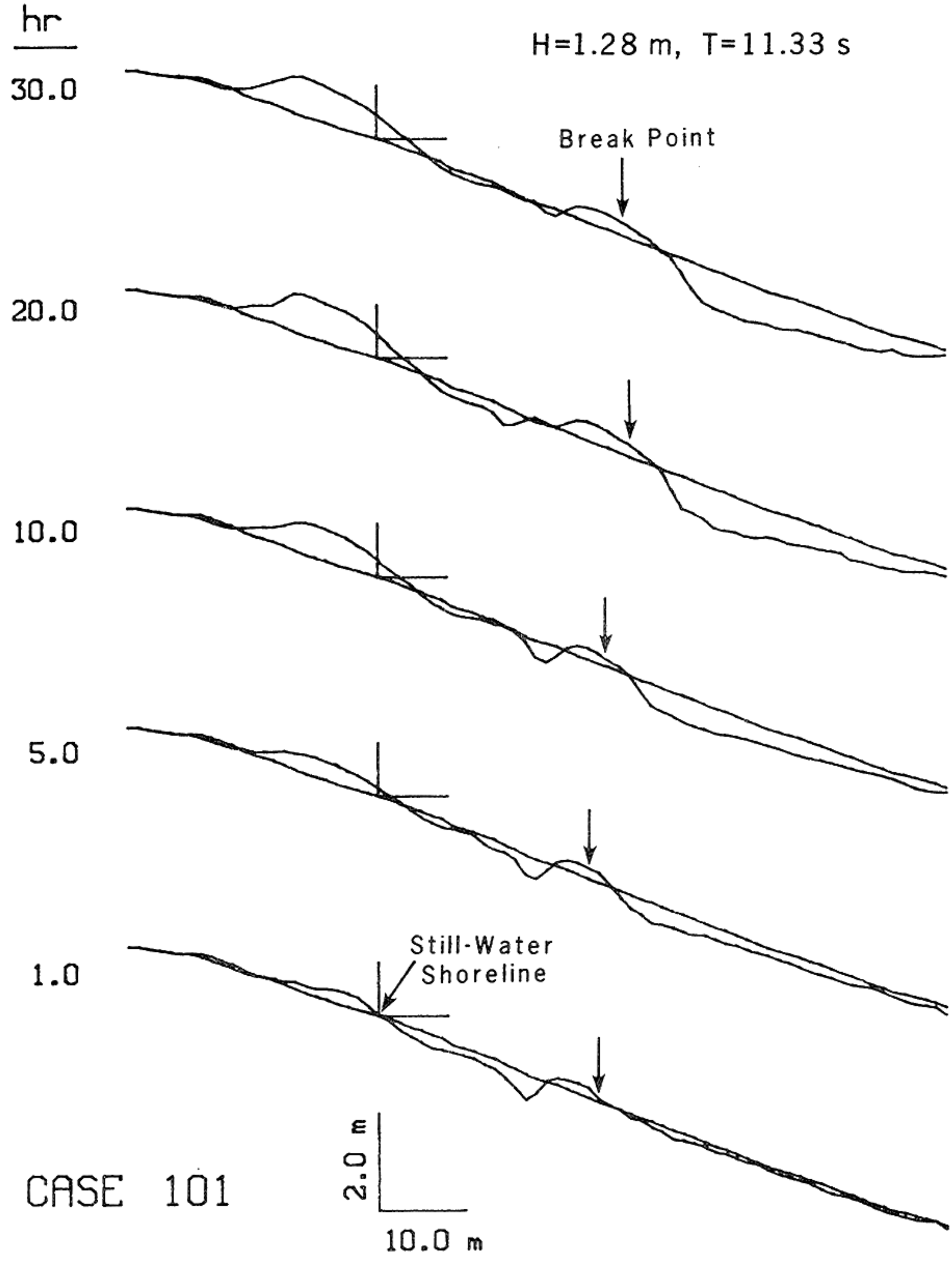


Fig.6. Deformations of sand-composed coastal slope under the action of gentle waves of swell. Data from test C101 in large wave flume [Larson and Kraus, 1989]

In phase of storm decay and during subsequent storm events the profile changes again. However its main features are kept for a long time. This implies the possibility of some virtual mean profile responding to wave climate inherent in a given region. Perturbations due to distinct storm events as well as the seasonal changes look like high-frequency fluctuations on the background of such a quasi-equilibrium profile.

Let's assume that we have determined equilibrium profile corresponding to given dynamical conditions. Then the potential deformations of actual profile might be assessed as its difference from the equilibrium one. Such an approach is employed in some models predicting beach changes [Larson, Wise, 1998]. The problem however is that the equilibrium profile is hard to be determined with desired accuracy and reliability. More effective method to calculate the morphological response is based on mass conservation principle connecting the bed deformations with the spatial gradients of sediment discharges. Increase or decrease in sediment transport rates should lead to corresponding increase or decrease in depths. Although the determination of sediment discharges is rather difficult problem, the recent morphodynamic models are already capable of predicting the storm-induced bed deformations with accuracy acceptable for engineering practice [Leont'yev, 2003b].

Fig.5 demonstrates the typical patterns of changes in coastal profile during a storm. According to scheme of nearshore circulation considered before (Fig.3), the sediments have to move toward the breaking region both from beach and a deeper section of sloping bed. As a result a bar grows in the convergence zone, while the beach and outer part of coastal profile are eroded. During severe storms and high storm surges the waves can erode the dune bordering beach and cause the recession of coast on tens of meters, what is observed, for example, on the Island of Sylt in the North Sea [Newe, Dette, 1995].

When a storm decays and coast is exposed to relatively gentle waves of swell, the bed profile can be partly restored as the waves of this kind tend to move the bed material toward the shore where a coastal bar or berm is built (Fig.6). It is clear that resulting coastal profile would depend on what kind of waves dominates over the year.

Figures 5 and 6 show that with the distance from shore the bed deformations decrease and actually cease at a certain depth. This depth, h_*, bordering an active section of coastal profile and per se marking the outer limit of coastal zone is called closure depth. In accordance with observations, the h_* value for a given coast can be approximately estimated by the doubled height of significant waves during severest storm (acting 12 hours per year) [Hallermeier, 1981, Capobianco et al., 1997]. On the open ocean coasts the closure depth h_* can exceed 15 m, while for the North, the Baltic and the Black seas it falls in range 7-10 m.

The upper limit of coastal zone can be determined by the highest water level or by the height of erosion cliff.

Irreversible perturbations of coast may be caused by anthropogenic intervention in natural processes. In Fig.7 the computed shoreline changes in the vicinity of port with access channel are shown (one year after constructing). Port has been constructed on the open sea coast where the vigorous longshore sediment fluxes in one or another direction are developed during the storms. As a result the sediments are accumulated at jetties, while at some distance from those a sediment deficit arises and the coast retreats over a long extension. In a given case the coast may be protected by means of bypassing – transport of sediments from the jetties to erosion areas.

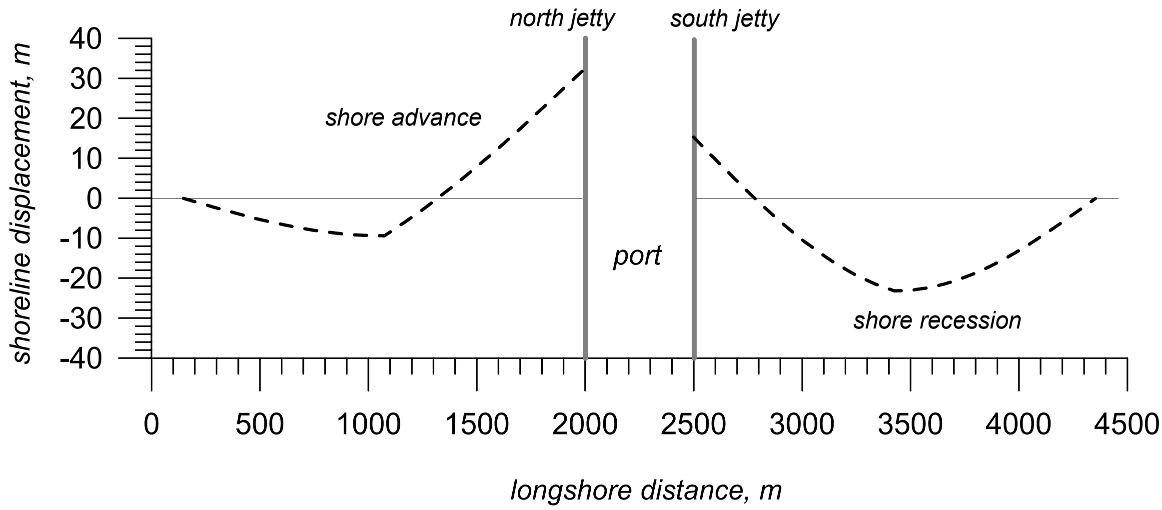


Fig.7. Shoreline changes one year later of constructing a port and access channel at the coast with significant longshore sediment transport

Returning to the equilibrium profile we ascertain that its determination may be based upon two different approaches. One of those employs the principle of mass conservation establishing that net sediment discharge in any point of stable profile tends to zero. In other words, sediment movement toward the shore should be balanced by the opposite motion over the whole profile. An alternative approach tends to interpret the equilibrium profile from the viewpoint of the energy expenditure in the system *waves – coast*. It is implied that the waves propagating over the movable coastal slope dissipate the energy in accordance with some general principle dictating the appropriate bottom profile geometry. A possible version of such a principle formulated by Dean [1991] is that the rate of energy dissipation, D, per unit volume of the water column is uniform.

At present time the solutions of the problem based on both approaches are found. It is surprisingly that these solutions lead to the same convex profile where the depth, h, should monotonically increase with the distance from shore, X, as [Dean, 1991, Larson et al., 1999]

$$h = AX^{2/3}$$

(parameter A depends on size of sediment particles). Theoretical curve is limited by the closure depth h_*.

It is well known however that the most typical profile for a gently sloping sandy coast is the bar-trough profile where the depth oscillates over its extension. In spite of the seasonal changes and migrations, the bar system is a stable morphological structure (on the time scales of decades, at least) and undoubtedly represents the principal feature of the equilibrium profile on the natural coast. Possible mechanism of formation of bar in the wave breaking region has been already discussed before. However it is difficult to give a reasonable explanation of phenomenon of a multiple bar system, and so several hypothesis has been suggested by different researchers [Boczar-Karakiewicz, Davidson-Arnott, 1987, Roelvink, Stive, 1989, O'Hare, Huntley, 1994, Zhang, Sunamura, 1994].

Theory suggested by the author [Leont'yev, 2004] builds on a new version of the energetic principle in which the Dean's formulation is included as a particular case. It is postulated that *the rate of wave energy dissipation in the nearshore zone is directly proportional to the energy itself per unit volume of the water column*. Besides, the presence of long infragravitational waves (IW) is taken into consideration. Those interact with the short gravity waves and generate weak spatial oscillations in energy flux. As it follows from the results obtained, coastal morphology turns out to be highly sensitive to these oscillations. A quite pronounced multiple bar system arises even if the ratio of the IW energy to short-wave energy is only of order of 10^{-2}. Spatial step of the system is controlled by the length of IW. Approaching the shore the bar height increases while the distance between the bars decreases. Theoretical profile satisfactorily reproduces natural profiles observed on different sea coasts (Fig.8).

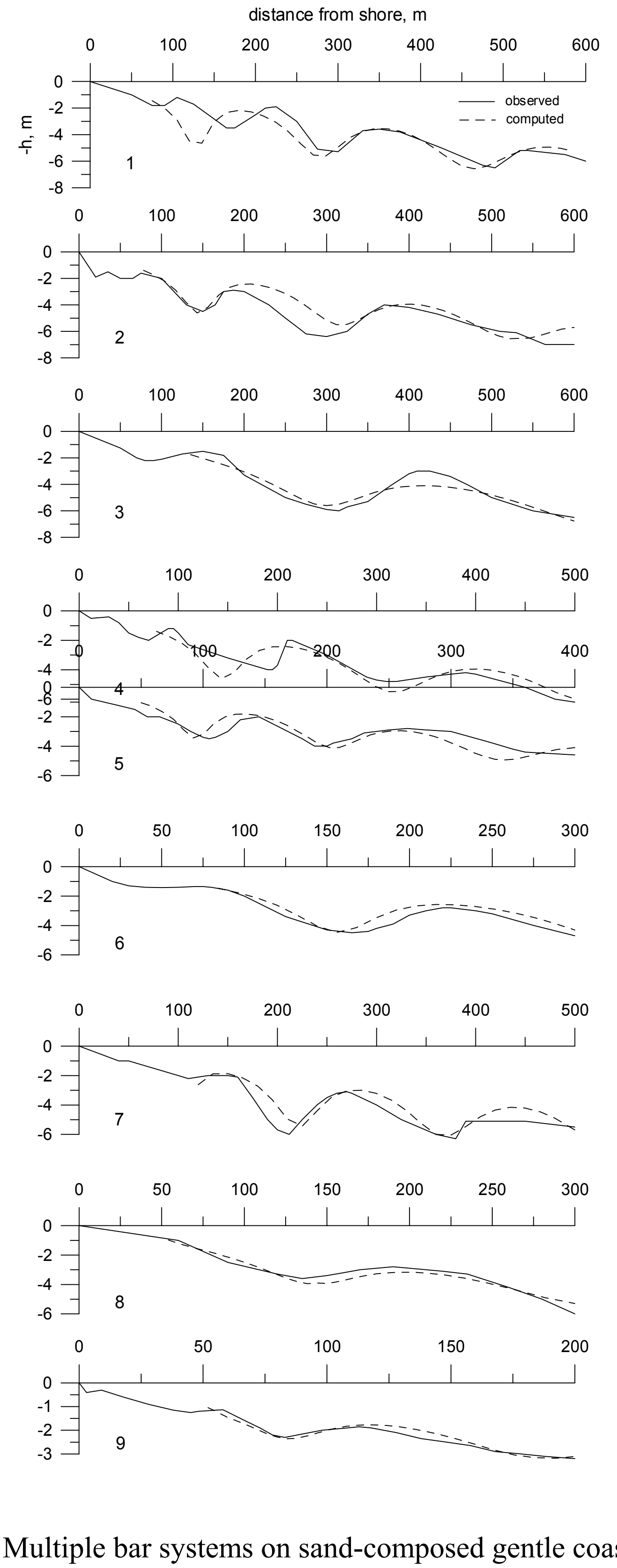


Fig.8. Multiple bar systems on sand-composed gentle coastal slopes:

1 – Lubiatowo (Poland, the Baltic Sea); 2 – Kursiu Nerija spit (Lithuania, the Baltic Sea); 3 – Egmond (The Netherlands, the North Sea); 4 – Le Carent (France, Mediterranean); 5 – Anapa (Russia, the Black Sea); 6 – Yevpatoria (Ukraine, the Black Sea); 7 – Hazaki (Japan, the Japan Sea); 8 – Duck (USA, Atlantic Ocean); 9 – Wendake Beach (Canada, the Lake Huron)

Sediment balance and equilibrium coastline contour

A given coast is usually composed of several morphodynamic systems with inherent energy, sediment transport, morphology and other properties. The limits of systems are determined, in particular, by the coastline prominences or the river mouths. Behavior of coastline is controlled by sediment balance in a given system. In the case of zero balance the coastline is stable, while in other cases it advances or retreats.

Using the principle of mass conservation and the equilibrium profile concept one can derive the following equation of sediment balance in morphodynamic system:

$$\frac{\partial \chi}{\partial t} = \frac{1}{h_* + z_c}\left(\frac{\partial Q_l}{\partial l} + q_c - q_*\right) + \frac{w}{\beta}$$

The left-hand side expresses the speed of displacement of coastline contour: χ is the coastline position and t is the time (measured here by years at least). Quantities h_* and z_c denote the closure depth and the elevation of the upper limit of coastal zone. Terms in brackets determine income or lost of sediments due to gradient of longshore flux Q_l (l is the longshore distance) and difference in cross-shore fluxes at the upper and lower limits of coastal zone (q_c and q_*). Last term reflects the contribution of sea level changes: $w = \partial \zeta / \partial t$ is the rate of change in sea level ζ and $\beta = (h_* + z_c)/l_A$ is a mean bed slope over the cross-shore length of coastal zone l_A. Meaning of the symbols used is explained by scheme in Fig.3.

When sediment fluxes passing a given system remain unchanged or the changes in cross-shore and longshore fluxes neutralize each other ($\partial Q_l / \partial l + q_c - q_* = 0$), the sediment balance equation transforms into the well known Bruun rule [Bruun, 1988]: $\Delta \chi = \Delta \zeta / \beta$. This states that coastline shifts in direct proportion to displacements of relative sea level $\Delta \zeta = w \Delta t$. Those can be due to both the changes in volume of water in the World Ocean and the tectonic motions of land. Based on the Bruun rule one can calculate that with the recent rate of sea-level rise 1.5 mm y^{-1} and the same rate of land subsidence in some regions the gentle coast with the typical bed slope β=0.006 should retreat on 0.5 m per year.

However mostly the behavior of coast depends to a greater extent on other sediment balance components, and so determination of those is a very actual problem. This problem, in principle, is solved on the base of long-term observations of the behavior of a given coast. But the relevant data required are far from always available and only indirect evaluating methods can help in the majority of cases.

One of these methods is based on the theory of coastline contour developed by the author [Leont'yev, 2005]. It is established that the equilibrium contour is described by the parabolic curve

$\chi = Py^2$ where the parameter P is directly proportional to the sediment flux gradient $\partial Q_l / \partial l$ and inversely proportional to the energy flux into the morphodynamic system. The greater the energy flux, the lesser the curvature of the contour and closer the coast to a straight line. Besides, if the gradient is positive (sediment flux increases) the counter is convex and if $\partial Q_l / \partial l < 0$ (flux decreases) the contour is concave. The theory conclusions are confirmed by the natural data. Several examples of comparison of the theoretical and observed coastline contours are shown in Fig.9.

Thus in order to estimate the gradient $\partial Q_l / \partial l$ for any coastal section one can try to approximate the coastline by the parabolic curve by means of variations in parameter P. Satisfactory result would indicate a relevancy of the above theoretical considerations in respect of a given coast. Then the magnitude of $\partial Q_l / \partial l$ can be found from a simple relationship including product of P and energy flux. The latter one is computed from data on wave regime.

The same data also serve to determine the cross-shore sediment flux at the lower limit of coastal zone q_*. Average annual value of q_* can be found by summation of quantities computed for various wave situations. Study of this kind has been undertaken by the author [Leont'yev, 2007] for different coasts situated in different regions. As was mentioned above, in outer part of the coastal zone one can expect predominantly shoreward movement of sediments due to asymmetry of wave velocities and mass transport in bottom boundary layer. However the study performed shows that the direction of q_* may be also opposite as it is controlled by the parameter $S_2 = 10^5 \beta_* d_s T_{4\%}^2$, where $\beta_* = h_* / l_*$ is the mean bed slope (l_* is the distance from shore to depth h_*), d_s is the mean size of sediment grains and $T_{4\%}$ is the wave period of 4 % regime cumulative accidence. Distribution of q_* values depending on S_2 is shown in Fig.10. When S_2>3 the flux q_* is positive, i.e. directed to the coast, whereas for S_2<3 q_* is negative. Such dependence is explained by the fact that under conditions of relatively short waves propagating over a very gently sloping bed the net cross-shore transport is controlled by the outflow due to storm surge.

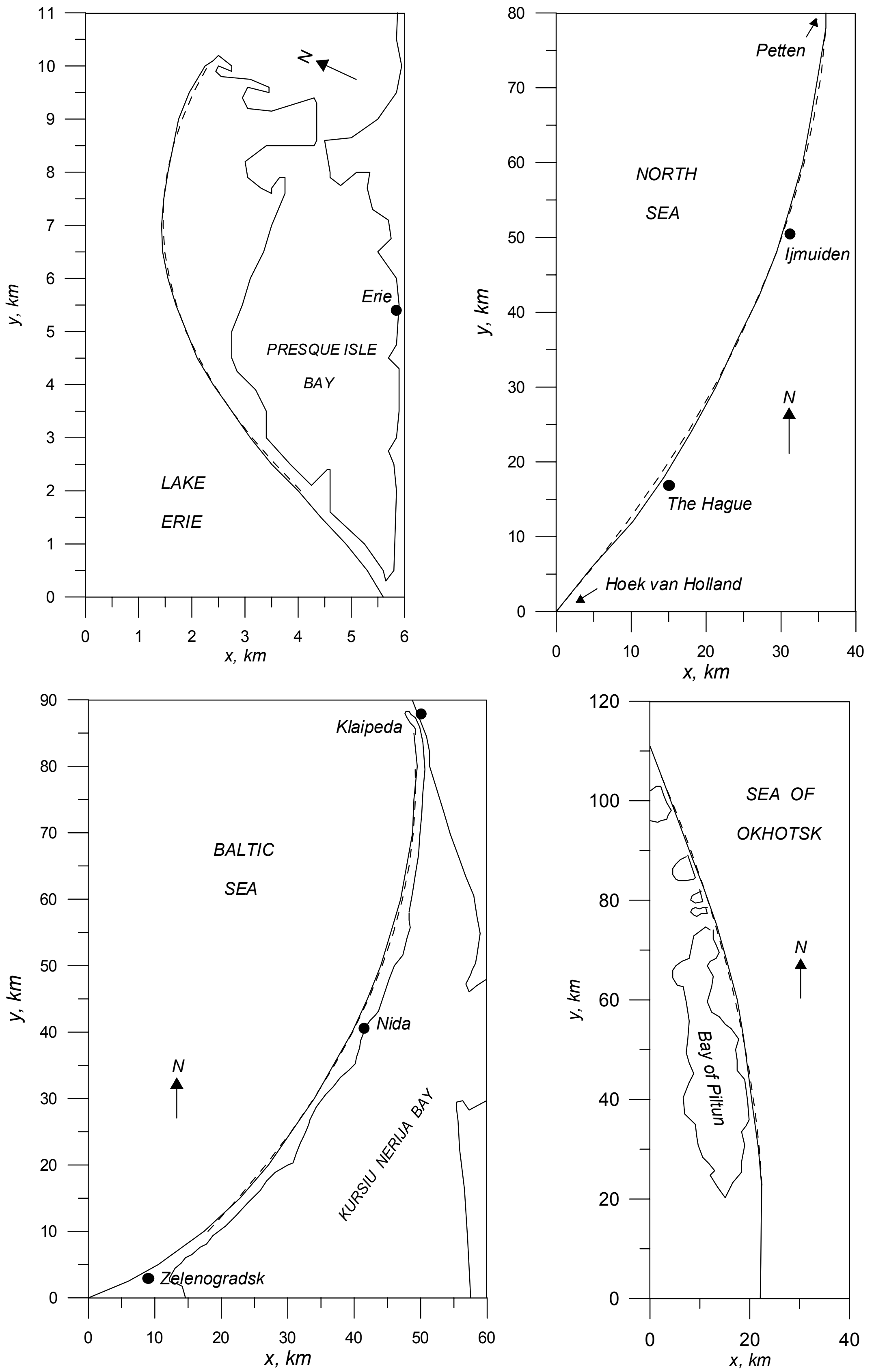


Fig.9. Coastline contours in some morphodynamic systems. Dashed curves show the theoretical parabolic contours

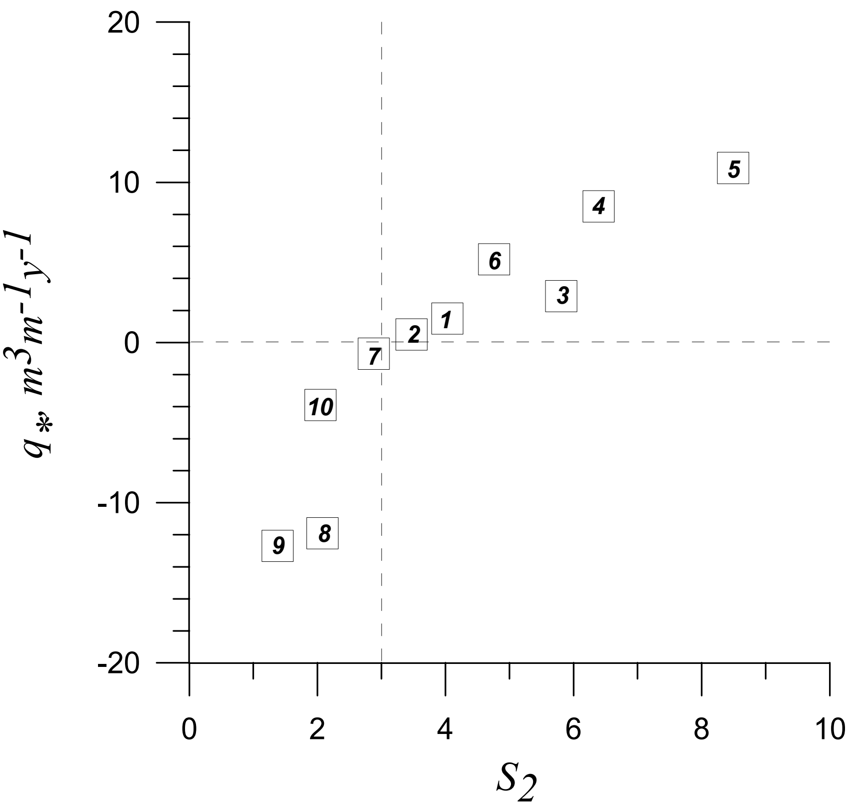


Fig.10. Changes in cross-shore sediment flux at the outer edge of coastal zone as depended on parameter S_2 characterizing dynamical conditions.

1 - Komarovo (Finnish Bay of the Baltic Sea); 2 - Temruk (Sea of Azov); 3 – Presque Isle spit (Lake Erie); 4 - Anapa (Black Sea); 5 – East Sakhalin (Sea of Okhotsk); 6 – Central Holland coast (North Sea); 7 – Pechorskaya Guba (Barents Sea); 8 – West Yamal (Kara Sea); 9 – Baidaratskaya Bay (Kara Sea); 10 – Cape Billings (East Siberian Sea)

Condition $S_2 > 3$ only implies the potential possibility of sediment transport toward the coast. Sediment resources in relatively deep section of coastal slope are also should be taken into account. Existing coasts were formed over the last 5-6 thousand years since the post-glacial transgression stopped and the level of the World Ocean approached to the recent mark. Over this period of time the sources of sand on the bed in some regions had run out and the sand supply to the beach became weaker. Material deficit in many cases is enhanced by the landward Aeolian sand transport from the beach (flux q_c through the upper limit of the coastal zone can reach 5-10 $m^3m^{-1}y^{-1}$ on some coasts [Cowell et al., 2001]). These adverse circumstances together with the gradual sea level rise are often in the bottom of the observed coastal erosion.

Prediction of coastal evolution

Sediment balance equation considered above allows for calculating the shoreline position at the end of a given period of time and so can serve as a quantitative basis to predict the future morphodynamic evolution of coast of interest. In Fig.11 some examples are shown of the computed 200-years evolution of coastal profile for various combinations of balance components. Two cases refer to the sea level rise on the background of sediment deficit. The third case illustrates the conditions of sea level fall accompanied by excess sediment supply and decrease in storm activity (for example, due to changes in climate of the region).

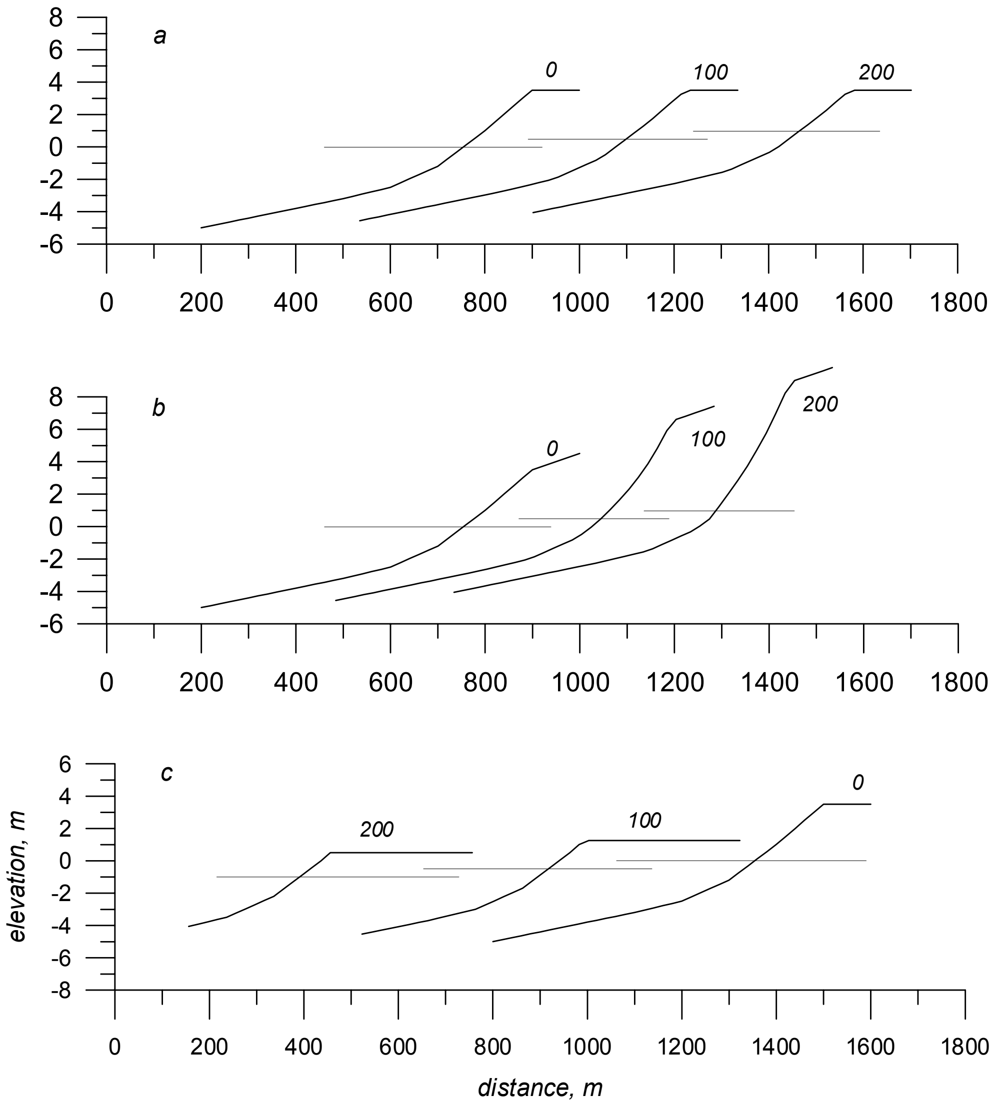


Fig.11. Examples of coastal evolution for different combinations of sediment balance components: a – erosion of horizontal marine terrace under sea-level rise and sediment deficit; b – erosion of the inclined terrace (slope=0.01) under the same conditions; c – advance of coast under sea-level lowering and decreasing storm activity. Numbers indicate the coast positions and sea level elevations at the initial moment and 100 and 200 years later

Serious problem making difficulties for prediction of coast behavior is the developing of reliable scenario of future changes in sea level. Different researchers represent various viewpoints on even the sea level changes over the nearest century. In particular, an opinion is advanced that observed climate warming can alter some decades later by fall of temperature and as a result the rate of sea level rise may remain at the recent value. [Dotsenko et al., 2004]. However the predominate estimation assumes the sea level rise on about 0.5 m to the end of nearest century [Church et al., 2001]. Such level change itself should lead to recession of relatively gentle coasts on distance of order of 100 m.

As an illustration of the morphodynamic forecast a possible evolution of the West Yamal coast in the Kara Sea over the next centuries is discussed further [Leont'yev, 2006]. Two large-scale morphodynamic systems are selected here. The first of those is situated northward of cape Burunniy and characterized by concave coastline (Fig. 12). The second system includes a series of low sandy islands Sharapovy Koshki extending southward of cape Harasavey and constituting general contour of the convex shape.

Although the ice-free season in this region may last only about 3 months per year, the computed energy and sediment fluxes attain significant magnitudes (Fig. 12). Longshore gradients $\partial Q_l / \partial l$ are evaluated as 2.3-2.7 $m^3m^{-1}y^{-1}$ (in the northern system gradients are negative while those in the southern system are positive). Due to very small bed slopes in the region studied the condition S_2<3 is satisfied, i.e. the cross-shore flux through the outer limit of the coastal zone is negative (directed toward the sea) and its magnitude is evaluated as −12 $m^3m^{-1}y^{-1}$. Hence in accordance with the sediment balance equation we have $\partial \chi / \partial t$ >0, what means that coast retreats. In the southern system recession should be more rapid due to contribution of component $\partial Q_l / \partial l$ >0.

Describing future changes in the ocean level we follow scenario developed by Pavlidis [2003] for the next 500 years. Changes in level ζ and the rate of changes w are shown in Fig. 13 by the dashed lines. It is assumed that 200 years later the sea level rises on about 0.9 m. After 100-years pause, the level continues to increase and at the end of 500-years period it attains the mark 1.4 m. The same scenario taking into account the tectonic subsidence of the West Yamal (with the rate 0.0015 m y^{-1}) is depicted in Fig.13 by the solid lines. Due to this additional factor the maximal level elevation can exceed 2 m after 500 years.

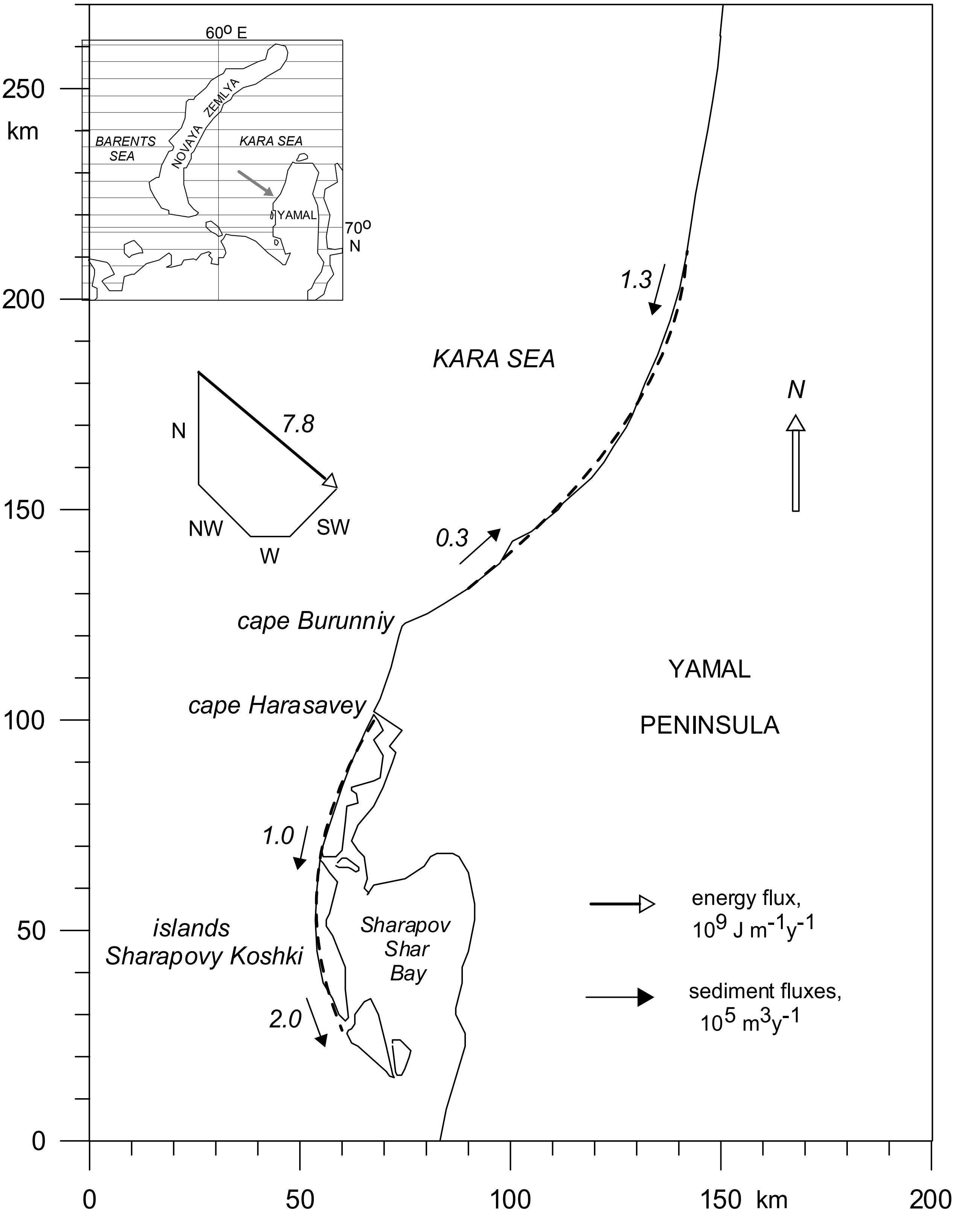


Fig.12. Location map of the West Yamal. Dashed lines indicate the theoretical coastline contours. Also the longshore sediment fluxes and energy flux vector are shown

Forecast of coastal evolution in the island system Sharapovy Koshki is shown in Fig. 14. Herein the successive positions of coastal profile are given (with the time interval 100 or 200 years) and also the initial and final elevations of the sea level are marked. Profiles depicted by the dashed lines respond to a stage of evolution when the maximal elevation of island above the sea level becomes lesser than the maximal surge height (2 m). Since this moment the destructive process is expected to much accelerate due to wave overwashing during the severest storms. Although by calculations the island still exists 100-150 years later, actually it may be entirely eroded much sooner. Therefore the evolution of comparatively narrow island (2 km width, Fig. 14a) is expected to come to the end about 500 years later, whereas the erosion of the wider island (3 km width, Fig. 14b) would continue about 700 years. The latter result assumes extrapolation of the adopted changes in the sea level.

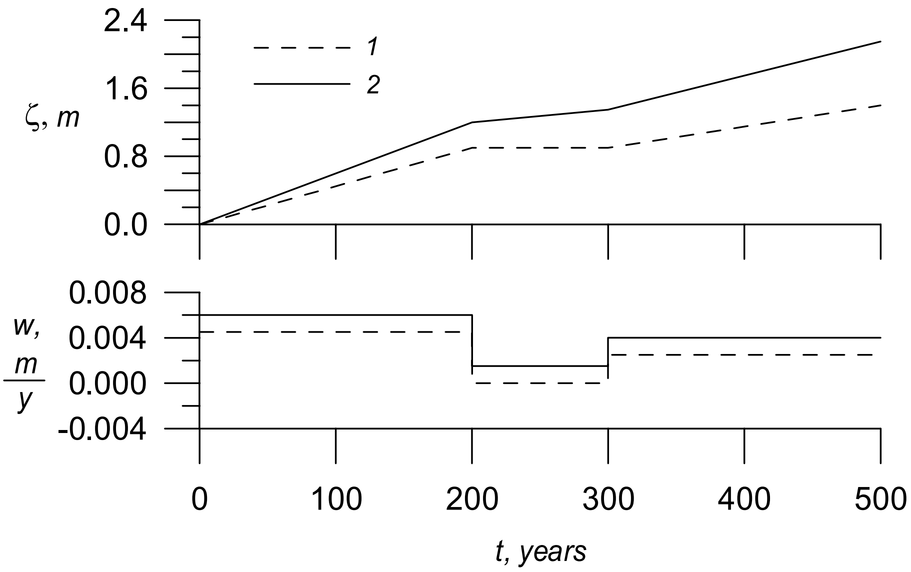
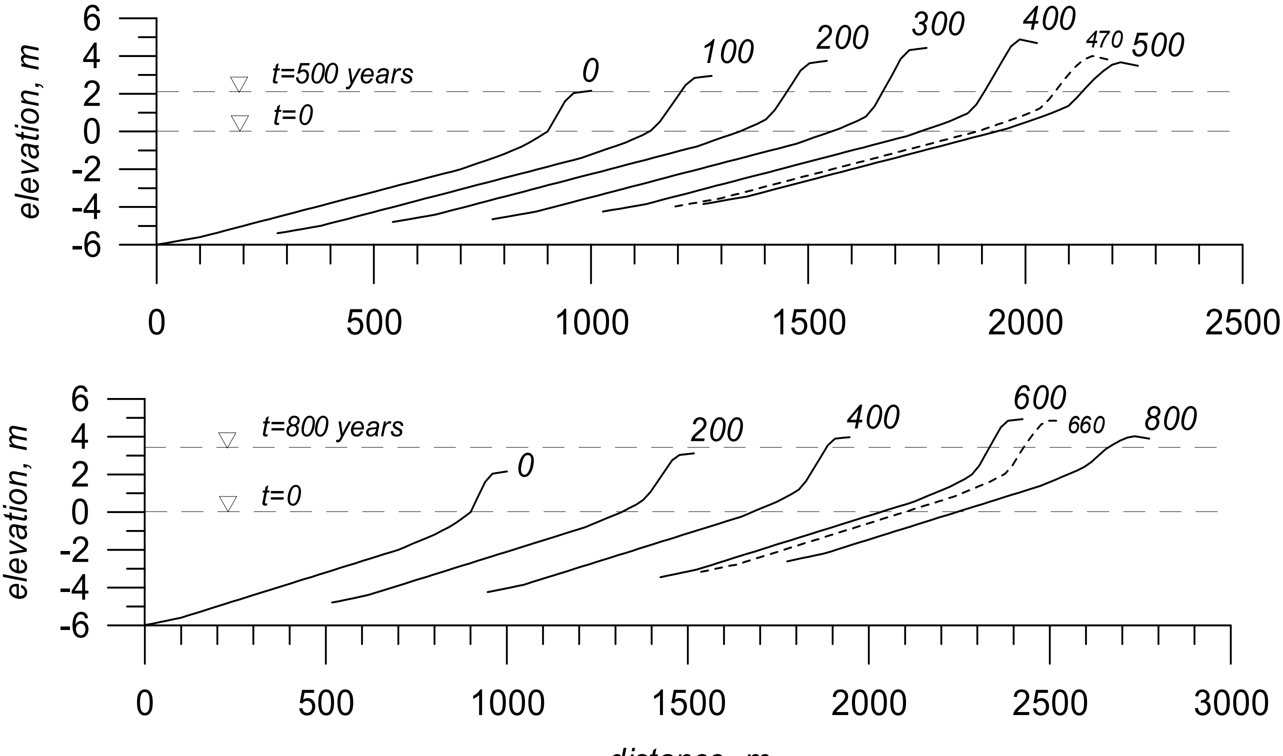


Fig.13. Expected changes in relative sea level over the following 500 years:
1 – scenario by Pavlidis [2003], 2 – the same scenario taking into account the tectonic subsidence of coastal region (ζ is the level elevation above the recent position and w is the rate of level changes)

Thus about 700 years later only the land fragments with the initial cross-shore length greater 3 km would still exist. However a number of developing scours and breaches would intensify the coastal erosion and 1-2 centuries later the whole system of islands Sharapovy Koshki can disappear. When the protective island barrier will be destroyed the abrasion cliffs of the West Yamal will be exposed to direct impact of the Kara Sea and recession of coasts can considerably accelerate.

a)



b

)

Fig.14. Predicted evolution of comparatively narrow (a) and wider (b) islands in the system of Sharapovy Koshki. Successive positions of coastal profiles are shown (with the time lapse 100 or 200 years) and also initial and final sea level elevations are marked. Dashed-line profiles respond to evolution stage when land elevation above the sea level becomes lesser than maximal height of storm surge

Conclusions

In this brief review we concerned with only a few problems of the coastal-zone morphodynamics. The sea coast is developed under influence of numerous processes characterized by various spatial and temporal scales and resulting in different trends in coast behavior. Properties of individual storms and average annual wave regime are important as well as the environmental conditions including the climate, sea level oscillations, flow regime, ice conditions, tectonic situation and other factors. At present time the quantitative methods using the mathematical modeling are involved into description of coastal dynamics and in many cases these methods help in selecting the right way to protect the coasts from erosion. However the coast is a complex system characterized by numerous feedbacks and tending to equilibrium with natural environment. Comprehensive understanding of many elements in this system is not yet reached. Recent level of knowledge not always allows for explanation of the observed trend in coastal changes and this offers the challenges for future researches.

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