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MARINE GEOLOGY

Forecast of Coastal Changes Based on Morphodynamical Modeling

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Abstract—A new version of the prognostic model is suggested, which differs from the previous version by its structure and method for including long-term variables into the process of coastal modeling. A fundamental prerequisite of the model is the assumption of the quasistationary state of the coast at any arbitrary stage of its evolution. The displacement of the coast under the influence of external factors at given moment is predicted on the basis of the obtained equation for the balance of sediments. The model takes into account a number of properties of the coastline. In particular, the distinguished dependence of the coastline shape on the regime of sediment transport is used to calculate the gradient of the alongshore sediment flux. The model also uses the empirical dependence found earlier, which makes it possible to estimate the second most important component of the sediment balance for selected types of coasts – the cross-shore matter flux. Special attention is focused on adjustment of the model to Arctic conditions. The results obtained are illustrated by the example of forecasting the development of accumulative coasts of the Western Yamal. It is shown that the major part of the Sharapovy Koshki sand islands could disappear within the next 700 years due to erosion.

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INTRODUCTION

In a series of previous publications [5, 6, 17–19], the author presented a morphodynamical model that makes it possible to describe the dynamics of the coastline for time periods from a few hours (storm deformations) to a few decades (displacements of the coast related to the influence of long-term processes). In particular, the application of the model to various types of Arctic coasts (accumulative, abrasive, and thermo-abrasive) allowed the author to obtain a number of interesting results related to their possible evolution within the next 100 years. Meanwhile, a number of restrictions used in the model make it difficult to apply the model to longer time scales.

In this article, the suggested approach to modeling the coastal evolution is further developed. The new version of the prognostic model differs from the previous one in its structure and in the method of including long-term variables into the modeling process. It is based on the equation of sediment transport, which relates changes in sea level to fluxes of solid matter within the given coast. The properties of the coast contour influencing the evolution of the morphodynamical system are taken into account. The patterns found, in particular, make it possible to estimate the main component of the sediment transport balance using calculations.

Special attention is focused on the adjustment of the algorithm of calculation to the conditions of Arctic coasts. The capabilities of the suggested model are illustrated by the example of long-term forecast for a portion of the western coast of the Yamal Peninsula in the Kara Sea.

SEDIMENT BALANCE WITHIN THE MORPHODYNAMIC SYSTEM

A coastal morphodynamical system (lithodynamical cell) with inherent energetics and morphological properties is considered [12, 21]. It is assumed that the coastal slope within the system is characterized by a generalized profile which changes its location in the course of evolution under the influence of external factors. Precisely the forecast of these long-term changes is the main goal of this study.

It is known that, in the process of evolution, the seacoast formed by mobile sediments tends to an equilibrium state with respect to external forcing by waves, currents, level changes, and other dynamical factors. We assume that the resources of the material and the geological setting of the region impose no significant restrictions on the development of the coast, while the mean amount of the energy transported to the morphodynamical system is constant. Then, according to Bruun's concept [10], in the dynamical balance, the coast moves forward or retreats (for example, following changes in sea level) conserving the main properties of its morphology. In this case, the seasonal and annual variations seem to be represented by high-frequency fluctuations against the background of long-term tendencies on a decadal scale.

The concept of equilibrium is related to the active zone of the coastal profile, where sediment transport and morphological changes are most perceptible. The active zone is conventionally limited by the depth contours $h = h_*$ and $h = -b$ (Fig. 1). The closure depth h_* is defined as the depth at which the deformations of the bottom measured with an interval of N years do not exceed a given limit related, for example, to the accuracy of measurements [11]. Hanson [14] suggests that the following approximation is possible:

$$h_* = 2H_s, \qquad (1)$$

where H_s is the significant height of the waves during extreme storms with a recurrence of approximately once a year. This estimate is related to the seas of the temperate zone, in which wave impact on the coasts is possible during most of the year. In the Arctic seas, the period of storm activity is significantly smaller due to ice screening. One can expect that here, changes in the bottom are notable in a narrower part of the coastal zone limited by a smaller depth h_*. In the Arctic region, the following estimate is more acceptable [6]:

$$h_* = 2\overline{H}, \qquad (2)$$

where $\overline{H}$ is the mean height of the waves for the storms with a recurrence of once a year ($\overline{H}$ is approximately 1.6 times smaller than H_s).

As to the upper boundary of the active zone $h = -b$, it is traditionally related to the maximal scale mark of the sea level, which is reached once every N years (for example, during superposition of the storm surge and tide). In the morphological aspect, this boundary can correspond to the summit of the coastal ridge (terrace) most remote from the shore formed by the flux of the wave run-up [14, 15]. However, this is mainly related to the conditions of a stable or growing coast. The most characteristic morphological property of the retreating coast is the abrasive cliff, which should obviously be included into the active zone of the profile. Correspondingly, in this case, the edge of the cliff should be considered as the boundary of the active zone, while the height b should be equal to the height of the cliff z_c.

Under the conditions described above, the balance of the sediments in the morphodynamical system is given by equation [7]

$$\frac{\partial \chi}{\partial t} = \frac{1}{z_A}\left(\frac{\partial Q_l}{\partial l} - q_A\right) + \frac{w}{\beta}, \qquad (3)$$

where χ is the position of the coastline; t is the time; l is the distance along the coastline; Q_l and $\partial Q_l/\partial l$ are the alongshore flux and its longitudinal gradient, respectively; q_A is the resulting cross coastal flux; $w = \partial\zeta/\partial t$ is the velocity of sea level changes ζ; $z_A = h_* + b$ is the so-called closure interval [15]; and $\beta = z_A/l_A$ is the mean slope of the bottom within the active profile zone, whose length is l_A. The sense of the notations used is explained in Fig. 1.

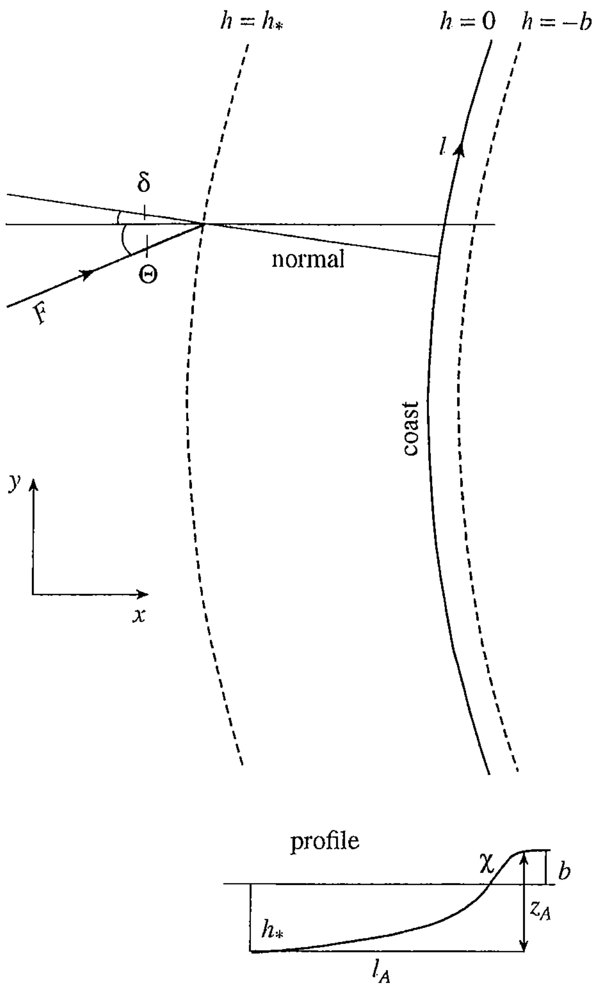

Fig. 1. Schematic of the coastal zone. See the text for notations.

If the sediments are transported through the given system in a transit manner ($\partial Q_l/\partial l = q_A = 0$) or the variations in the cross-shore and alongshore fluxes neutralize each other ($\partial Q_l/\partial l = q_A$), Eq. (3) is transformed to the well-known Bruun rule [10], according to which the variations in the location of the coast $\Delta\chi$ are determined only by the variations in the sea level height $\Delta\chi = w\Delta t/\beta$. However, in many cases, the behavior of the coast depends more on the components of the sediment balance. In particular, this is related to the Arctic coasts, which are eroded and retreat ($\partial\chi/\partial t > 0$) owing to the significant offshore flux of the material ($q_A < 0$) [6, 17, 19].

SELECTED PROPERTIES OF THE COASTLINE

The coastline theory suggested by the author of [7] includes two main prerequisites. First, it is taken into account that, when the coast in an equilibrium state is displaced, its profile and coastline remain unchanged; i.e., each point of the coastline is translated with the same speed. Second, the traditional concept is used that the alongshore sediment flux Q_l is proportional to the longitudinal component of the energy flux F_l:

$$Q_l = kF_l, \quad F_l = F\sin(\Theta + \delta)\cos(\Theta + \delta), \qquad (4)$$

where k is the proportionality coefficient; F is the resulting flux of energy; Θ is its angle with respect to the normal to the general direction of the coastline (to the OY-axis in Fig. 1); and δ is the angle of the coast deviation from the general direction. Under these conditions, the equilibrium contour is described by a parabolic curve of the following form

$$\chi = Py^2, \quad P = \frac{1}{2kF\cos 2\Theta}\frac{\partial Q_l}{\partial l}. \qquad (5)$$

Three cases are possible here: (1) $\partial Q_l/\partial l > 0$ is related to the case of the sediment flux uniformly increasing along the coast, $P > 0$, and the contour is convex (this case is shown in Fig. 1); (2) $\partial Q_l/\partial l < 0$ is related to the case of the sediment flux uniformly decreasing along the coast, $P < 0$, and the contour is concave; (3) $\partial Q_l/\partial l = 0$ is related to the case of a zero or constant sediment flux along the coast, $P = 0$, and $\chi = 0$, i.e., the contour is rectilinear.

The curvature of the contour characterized by the value of P is proportional to the gradient of the sediment flux and inversely proportional to the energy flux. The greater the energy transported to the morphodynamical system, the smaller the curvature of the coast and the closer the coast to a rectilinear shape.

Eq. (5) can be presented in the following form to compare it with the observed forms of the coastline:

$$\chi = \chi_1 + P[y^2 - (y_1 + y_2)y + y_1 y_2] + \frac{\chi_2 - \chi_1}{y_2 - y_1}(y - y_1). \qquad (6)$$

Using a pair of reference points with coordinates χ_1, y_1 and χ_2, y_2, one can try to approximate the actual contour of the coast by a parabolic curve by selecting an optimal value of the parameter P.

Such comparison performed for a number of morphodynamical systems on the coasts of the seas and large lakes demonstrated the possibility of the existence of the forecasted dependence between the sign of the gradient $\partial Q_l/\partial l$ and the shape of the coast [7]. In addition, calculations of the parameter P confirmed the theoretical conclusion of decreasing curvatures of the coasts with increasing level of energy influence. For example, the values of P for the Presque Isle Spit (Lake Erie), Curonian Spit (Baltic Sea), and northeastern coast of Sakhalin (Sea of Okhotsk) were estimated as 1.0×10^{-4}, -7.0×10^{-6}, and 2.2×10^{-6} m^{-1}, respectively. In the former case, the energy flux to the coast is minimal, while in the latter case it is maximal.

The results considered above are based on the assumption that the morphodynamical system is in equilibrium. Actually, the evolution occurs in a quasiequilibrium regime because the profile and coastline do not always fully adjust to the varying external conditions. However, one can expect that, if the determining factors change sufficiently slowly, the conclusions obtained remain in force.

ESTIMATE OF THE SEDIMENT BALANCE COMPONENTS

Generally speaking, the sediment balance components $\partial Q_l/\partial l$ and q_A in Eq. (3) should be determined on the basis of long-term observations of the coast dynamics. However, the lack of necessary data forces us to apply indirect methods of estimates. The patterns found in the previous section of the article allow us to determine, in particular, the gradient of the alongshore sediment flux $\partial Q_l/\partial l$ on the basis of the data on the shape of the coastline and energy flux to the shore. First, we use Eq. (6) to select the optimal value of parameter P for the coast considered. The possibility of a satisfactory approximation of the actual contour by a parabola would indicate that application of the theoretical model is possible. Then, the value of $\partial Q_l/\partial l$ can be found from relation

$$\frac{\partial Q_l}{\partial l} = 2PkF\cos 2\Theta, \qquad (7)$$

which follows from the definition of P in (5).

The resulting energy flux F is calculated on the basis of the wave regime data. The instantaneous flux of the wave energy per unit coast length (in J m^{-1} s^{-1}) is determined by the relation

$$EC_g = \frac{1}{16}\rho g H^2 \frac{L}{T}, \qquad (8)$$

where E is the wave energy per unit area; C_g is the velocity of its propagation; H, T, and L are the height, period, and length, respectively, of the waves in deep water ($L = (g/2\pi)T^2$). The annual flux in the given direction (in J m^{-1} yr^{-1}) takes into account the duration of wave situations t_w

$$F_a = \sum_i (EC_g t_w)_i, \qquad (9)$$

while the resulting value of F is the vector sum of the values of F_a.

According to the known CERC relation [20], the proportionality coefficient k in relations (4) and (7) is written as

$$k = \frac{0.77}{g(\rho_s - \rho)(1 - \sigma)}\frac{\sin\theta_b\cos\theta_b}{\sin\theta\cos\theta}, \quad \theta = \Theta + \delta. \qquad (10)$$

Here, g is the acceleration due to gravity, ρ and ρ_s are the densities of water and solid particles; σ is the porosity of the sediments; $\theta = \Theta + \delta$ is the angle between the vector of the energy flux (wave beam) and the normal to the coast in deep water (Fig. 1); and θ_b is the similar parameter at the depth of wave breaking. If the value of θ is known, θ_b can be found from the empirical relation [16]

$$\theta_b = \theta[0.25 + 5.5(H/L)]. \tag{11}$$

As to the cross-shore sediment flux q_A, under selected conditions it can also be estimated from the calculations. For example, the following empirical relation was found for the retreating Arctic coasts, where $q_A < 0$ [6, 19]:

$$q_A = -(a\sqrt{z_{ce}}\bar{H}\cos\Theta_w - b), \tag{12}$$

where q_A is expressed in $m^3 m^{-1} yr^{-1}$; $a = 11.2$ and $b = 13.4$ are the constants; $z_{ce} = (1-n)z_c$ is the effective height of the cliff; z_c is its actual height; n is the ice content in the sediments; $\bar{H}$ is the mean wave height during storms with a recurrence of once a year (as above). The angle Θ_w between the dominating direction of the winds and the normal to the coast is actually close to the value of Θ for the resulting energy flux.

The constant a indirectly reflects the characteristic duration of the ice-free season (t^+), during which the coasts are subjected to hydrodynamical and thermodynamic forces. Although the duration of t^+ in the western regions of the Russian Arctic is greater than in its eastern regions, the values of a for both regions do not differ greatly. It is likely that the time factor indirectly existing in relation (12) has a power smaller than unity, and to the first approximation we can accept that $a = \bar{a}\sqrt{\tilde{t}^+}$. Here, the value of $\tilde{t}^+ = t^+/T_1$ is understood as the share of the summer season related to the total duration of the year T_1. Then, relation (12) can be written as

$$q_A = -\bar{a}\sqrt{z_{ce}\bar{H}\tilde{t}^+}\cos(\Theta_w - b). \tag{12a}$$

In the western Arctic, the typical value of t^+ is close to three months ($\tilde{t}^+ = 0.25$) and the value of the coefficient $\bar{a}$ can be estimated as 22.4.

PROCEDURE OF CALCULATIONS

In the new version of the model suggested, the location of the coast at the given moment is determined by integration of Eq. (3). The initial data for the calculations include: the initial profile of the bottom $h(x, t_0)$, the parameters of the storm with a recurrence of once a year (heights H_s or $\bar{H}$), the maximum sea level height η_m, the velocity of the relative mean sea level change w (with account for the vertical tectonic movements of land), the components of the sediment balance $\partial Q_l/\partial l$ and q_A, and the time variations of all these values. If the detailed shape of the profile is known, it can be approximated by the equilibrium Bruun–Dean profile [13]

$$h = AX^{2/3}, \tag{13}$$

where X is the distance from the coast and the parameter A can be estimated, for example, on the basis of the 10-m depth contour.

The time step in the calculations Δt can be of the order of one year because the processes modeled are considered at least on a decadal scale. At each time step, first, the closure interval is determined $z_A = h_* + b$. The value of h_* is calculated from relations (1) and (2) depending on the climatic zone, while the value of b is taken equal to η_m or z_c depending on the type of coast. It is taken into account that the height of the cliff can change in time both owing to the changes in the sea level $\Delta\zeta = w\Delta t$ or due to the inclination of the surface of the eroded marine terrace m:

$$z_c = z_c^0 - \Delta\zeta + m\Delta\chi \tag{14}$$

where z_c^0 is the initial height of the cliff and $\Delta\chi$ is the displacement of the coast. The variations of h_* caused by the fluctuations of the wave activity are also taken into account. In this case, the lower boundary of the profile is displaced to either side and the length of the active zone is either increasing or decreasing, which results in the mean slope of the profile $\beta = z_A/l_A$.

After determination of the parameter z_A, the displacement of the coast $\Delta\chi$ is determined and the cycle of calculations is repeated until the current time t is not be equal to the given duration of evolution T_E.

Examples of the evolution of coastal profiles calculated on the basis of the suggested model are shown in Fig. 2. In each case, the initial parameters of the waves, the duration of evolution, as well as the absolute values of the sediment balance components and the rates of level changes are the same ($H_s = 2.5$ m, $T_E = 200$ years, $(\partial Q_l/\partial l - q_A) = \pm 25\ m^3 m^{-1} yr^{-1}$, $w = \pm 0.005\ m\ yr^{-1}$). The numerals 0, 100, and 200 indicate the location of profiles at the initial moment and after 100 and 200 years. The sequential locations of the sea level are also shown.

Figure 2a depicts the development of the coast in the case of a sea level rise and sediment deficit. In the course of time, the coastal retreat accelerates because the height of the cliff over the sea level gradually decreases.

Figure 2b depicts a similar process for the case when the eroded marine terrace elevates in the land direction. Here, the recession of the coast gradually slows down.

Figure 2c characterizes the growth of the coast owing to the excessive supply of sediments during a sea level drop. Simultaneously, wave parameters and maximal storm level decrease ($\bar{H}$ decreases from 2.5 to 1.5 m, and η_m decreases from 2 to 1.5 m). Advancing of

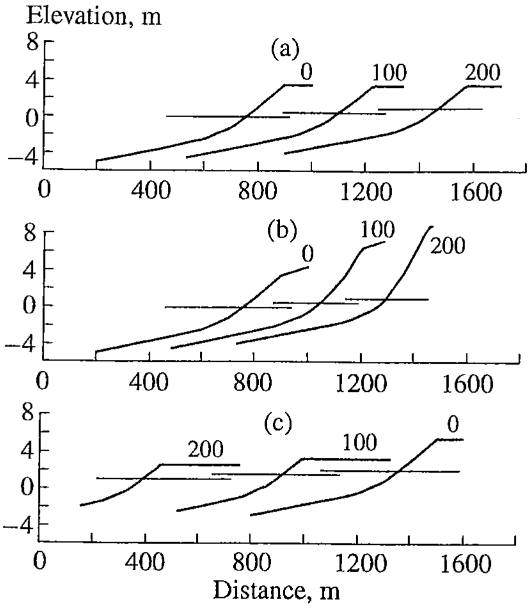

Fig. 2. Examples of the evolution of the coasts based on the model accepted: (a) erosion of a horizontal marine terrace under the conditions of a sea level rise and a deficit of the sediments; (b) erosion of an inclined terrace under the same conditions; (c) seaward motion of the coast under the conditions of a sea level fall, accumulation of the sediments, and a decrease in the storm activity. Sequential locations of the coastal profiles are shown at the initial moment (0) and after 100 and 200 years.

the coast slightly accelerates due to a decrease in the closure interval z_A.

FORECAST OF THE EVOLUTION OF WESTERN YAMAL COASTS

Dynamical Conditions of the Region

Let us illustrate the capabilities of the suggested model by an example of long-term forecast of the evolution of a coastal region of the Kara Sea located in the western part of the Yamal Peninsula (Fig. 3). A particular feature of this region is the existence of sand accumulative coastal systems stretched over tens of kilometers north of Cape Burunnyi and south of Cape Kharasavei. The northern region has a concave coastline. Here, a sandy beach is adjacent to the cliff of a marine terrace. The southern region includes a series of low sandy Sharapovy Koshki Islands, and the arch they form has a convex form. The coastal slope is very flat everywhere. The 10-m depth contour is located at a distance of 2–3 km from the coast.

The calculations based on Eq. (6) show that the contours of both regions are satisfactorily approximated by parabolas (the dashed lines in Fig. 3). This allows us to apply relation (7) to estimate the gradients of the alongshore sediment flux. In this case, the values of parameter P are determined as -5.5×10^{-6} and 5.3×10^{-6} m^{-1} for the northern and southern regions, respectively.

The calculation of the resulting energy flux in the Yamal region considered here is based on the data of the hydrometeorological regime adopted from [3]. The sea basin in this region is ice-free no longer than three months a year (on average, from the second half of July to the first half of October). During this period, the wind fetches for the main directions reach a distance of the order of 100–200 km, while the depths in the fetch region are equal to 50–100 m. The mean parameters of the waves were calculated on the basis of the wind data using the method suggested by Lappo et al., [4]. The wind velocity W, the corresponding heights H and periods of waves T, the annual duration of different wind and wave situations as well as the annual energy fluxes F_a over the main directions (N, NW, W, and SW) calculated from Eqs. (8) and (9) are given in the table. In addition, the graphs of the values of F_a and the resulting vector of the energy flux F are shown in Fig. 3. The value of F is estimated as 7.8×10^9 J m^{-1} yr^{-1}.

It is possible to accept that a value of 3 m is a representative wave height for the storm with a recurrence of once a year. Then the closure depth h_* according to relation (3) would be equal to 6 m. Since the maximal level η_m during storm surges can reach 2 m [1, 9], the closure interval $z_A = h_* + \eta_m$ would be estimated as 8 m.

In the northern coastal region, the vector F is oriented at an angle of $\Theta \cong -10°$ with respect to the normal to the general direction of the coast. The gradient of the sediment transport $\partial Q_l/\partial l$ calculated from relations (7), (10), and (11) at typical characteristics of the sand material ($\rho_s = 2.65 \times 10^3$ kg m^{-3}, $\sigma = 0.4$) is estimated as -2.7 m^3 m^{-1} yr^{-1}; i.e., Q_l decreases along the coast.

Annual durations of typical wave situations (in hours) and energy fluxes for the main directions

W, m/s	H, m	T, s	N	NW	W	SW
6–9	0.63	3.75	104	69.6	71.9	95.7
10–13	1.21	5.07	48.0	30.2	19.0	26.6
14–17	2.09	6.61	19.4	11.4	4.8	8.9
18–20	3.03	7.92	3.2	2.3	1.1	1.2
≥ 1	3.68	8.72	0.2	0.2	–	0.4
F_a, 10^9 J m^{-1} yr^{-1}			4.57	2.92	1.61	2.52

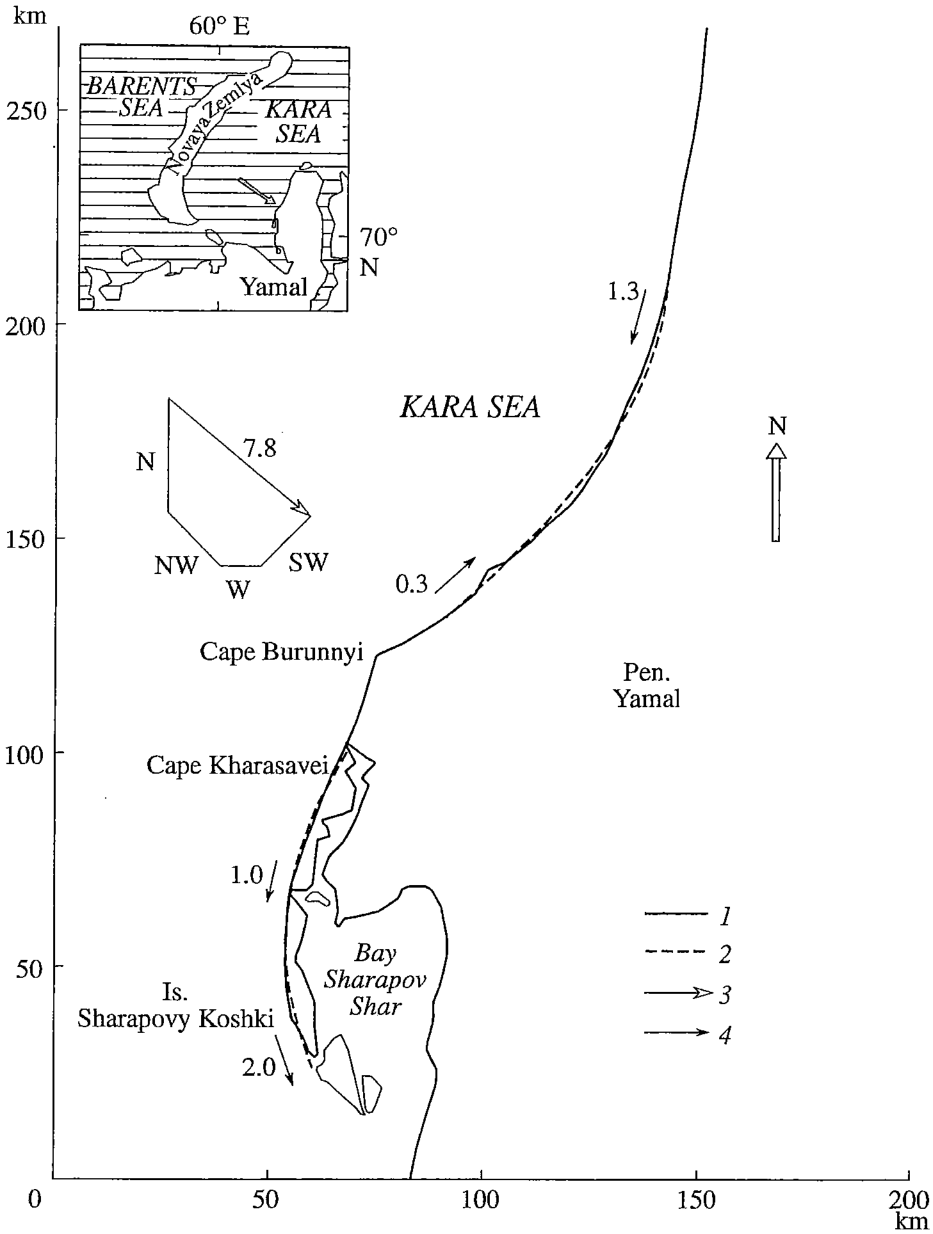

Fig. 3. Schematic map of the western coast of the Yamal Peninsula: *1*—actul coastline; *2*—theoretical coastline in accumulative regions; *3*—energy flux, 10^9 J m^{-1} yr^{-1}; *4*—sediment fluxes, 10^5 m^3 yr^{-1} (*4*).

Moreover, the calculations indicate that two opposite fluxes are possible with initial values approximately equal to -1.3×10^5 and 0.3×10^5 m^3 yr^{-1} (Fig. 3). Thus, the conditions of the alongshore transport favor the accumulation of material and, if the antagonistic factors are absent, the coast could grow here.

It is noteworthy that the data about the wind regime are available only for the northern part of Yamal and for Baidaratskaya Bay. In this case, we used the data only for the first of the regions named as the most representative for the conditions of the open coast. According to the table, the northerly winds give the main contribution to the energy flux. However, in the southern part of the Kara Sea, the role of the southern winds increases, and the direction of the resulting flux becomes close to the east–west direction. One can reasonably assume that, for the southern accumulative region of the coast, the angle Θ can be approximately equal to $-20°$. Then the gradient $\partial Q_l/\partial l$ can be determined from relation (7) as 2.3 m^3 m^{-1} yr^{-1}. Hence, the sediment flux increases southward and its value is estimated as 1.0×10^5 m^3 yr^{-1} in the middle and as 2.0×10^5 m^3 yr^{-1} at the southern edge of the coast considered here (Fig. 3). Evidently, this kind of regime favors the erosion and retreat of the coasts.

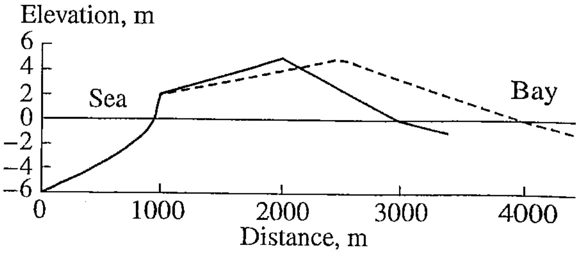

Fig. 4. Schematic profiles of relatively narrow and wide islands used in the calculations.

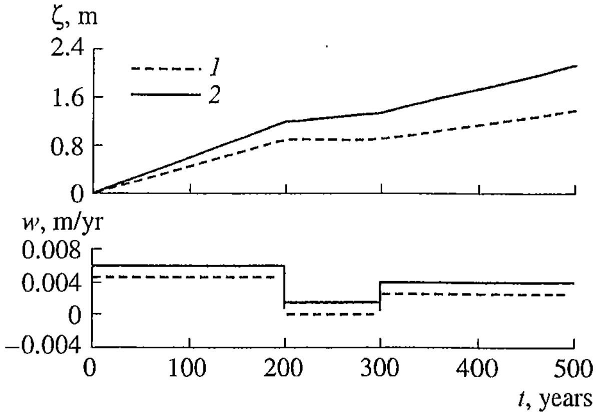

Fig. 5. Supposed variations of the relative sea level within the future 500 years: *1*—the Pavlidis' scenario [8]; *2*—the same scenario with account of the tectonic subsidence of land in the region of Yamal considered in this study (ζ is the position of the level, w is the rate of its variation).

Forecast of the Coast Recession

The Sharapovy Koshki Islands are elevated over the modern sea level only by 3–5 m and their width does not exceed 2–4 km. The region of the Yamal coast considered here is located in the zone of land subsidence with a velocity approximately equal to 0.001–0.002 m yr^{-1} [2]. Against the background of the global level rise, the actual speed of its elevation in the region considered here may soon exceed 5 mm per year. Thus, if the existing tendencies are retained, one can expect erosion and disappearance of the islands in the course of time. Let us make an attempt to forecast the probable duration of this process on the basis of the model suggested here.

Let us, first of all, make a scheme of the transversal profile of the island (Fig. 4) specifying its maximal elevation as 5 m and assuming that its width is either equal to 2 km (relatively narrow island) or 3 km (relatively wide island). The retreating western coast should have a clearly manifested cliff. We shall assume that its height at the initial stage of evolution is equal to $z_c = \eta_m = 2$ m. We assume that the maximal elevation is located in the middle of the island and the surface of the land gradually descends in the direction toward the shallow bay Sharapov Shar located on the eastern side of the island (Fig. 4). We assume that the bottom profile between the coast and the closure depth $h_* = 6$ m corresponds to Eq. (13). At the existing small slope angles of the bottom, such depth should be located at a distance of the order of 1 km from the coast, which gives a value of $A \cong 0.06$.

Estimating the cross-shore sediment flux q_A and using relation (12) with the values $\overline{H} = 3$m, $z_{ce} = z_c = 2$ m (ice content in sediments is considered insignificant), and $\Theta_w \approx \Theta = -20°$, we find that $q_A = -12.4$ m^3 m^{-1} yr^{-1}. Hence, the negative contribution of q_A to the sediment balance is significantly greater than the contribution of the gradient $\partial Q_l/\partial l$. We note that, in the course of the coast evolution, q_A would change following the variations in z_c.

We shall determine the future variations in the ocean level on the basis of the scenario developed by Yu. A. Pavlidis [8]. The evolution of the level ζ and its variations w corresponding to this scenario are shown in Fig. 5 with dashed lines. It is assumed that, in 200 years, the level will increase by 0.9 m. After a 100-year pause, the level increase will continue and, by the end of 500 years, it will reach the 1.4-m mark. The solid lines in Fig. 5 show the same scenario with account of the tectonic subsidence of land in the region considered here with a velocity of 0.0015 m yr^{-1}. After 500 years, due to this process, the maximal elevation of the relative level would exceed 2 m.

The evolution of the coast calculated on the basis of our model under all the conditions described above is shown in Fig. 6. This figure reflects sequential positions of the coastal profile with time intervals of 100 or 200 years, and the initial and final locations of the sea level are also shown. The profiles shown with the dashed lines correspond to the evolution stage, when the maximal elevation of the island becomes smaller than the maximal height of the storm surge (2 m). From this moment, the destructive process should significantly accelerate due to the overflowing and full flooding of the coast during extreme storms. Although after 100–150 years, according to our calculations, these islands will still exist, actually, they may be eroded much earlier. We can expect that, in approximately 500 years, a relatively narrow island (2 km wide, Fig. 6a) will disappear, while the erosion of a wider island (3 km wide, Fig. 6b) will continue over approximately 700 years. We note that the latter result is based on the linear extrapolation of the assumed variations in the sea level.

Thus, in 700 years, only fragments of the islands, whose present width is more than 3 km, will remain. However, the appearance of through scours will lead to an intensification of erosion and, after a short period of time, the entire island system will vanish. After the disappearance of the protecting barrier, the direct influence of the sea on the abrasive cliffs of the western Yamal will start. Their recession, which currently does not exceed 1–1.5 mm/yr [1], may significantly accelerate.

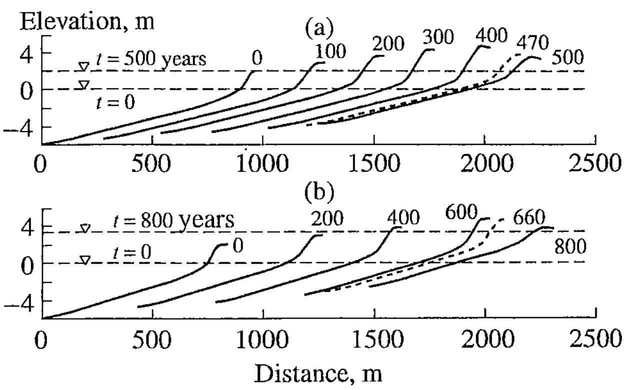

Fig. 6. Forecast of the evolution of (a) a relatively narrow and (b) a wide islands in the coastal system of the Sharapovy Koshki Islands. Sequential positions of the coastline are shown with an interval of 100 and 200 years together with the initial and final position of the sea level. The profiles shown with the dashed lines correspond to the stage of evolution, when the maximal elevation of the island becomes smaller than the maximal height of the storm surge.

CONCLUSIONS

The model presented here is designed for forecasting the morphodynamical evolution of coasts on time scales of the order of a few centuries, although in principle, the time of the forecast is limited only by the reliability of the scenarios describing the variability of the main parameters such as the mean sea level and the fluxes of solid matter in the coastal zone.

A fundamental prerequisite of the model is the assumption of the quasiequilibrium state of the coast at any given stage of evolution. This means that, in the course of evolution, the spatial location of the coast is subjected to the main variation, while its profile within the active zone and its contour conserve relatively constant features (under a constant dynamical forcing). The sediment balance equation for a quasiequilibrium morphodynamical system, which can be used for forecasting the displacement of the coast after a given time period, is significantly simplified. Moreover, the application of this approximation makes it possible to distinguish a number of important properties of the coastline that are reflected in the observed morphodynamical systems of various scales. In particular, establishing the correlation between the geometry of the contour and the gradient of the sediment flux allows us to estimate this important component of the balance by means of calculations, which widens the limits of the forecast. The structure of the calculations using the new model is simplified and their amount decreases as compared to the previous version.

The potentialities of the model were demonstrated by the example of the forecast of the development of the coast of western Yamal. Two large accumulative regions are distinguished here, which are located north of Cape Burunnyi and south of Cape Kharasavei. Application of the model leads to the conclusions that the conditions of the sediment transport in the northern concave region facilitate accumulation of the sediments, while the southern convex region including the chain of the sand islands Sharapovy Koshki is subjected to erosion. The forecast of the evolution of the coasts performed with account of the global ocean level rise and tectonic subsidence of land in the region considered here points to the possibility of the disappearance of the major part of the islands within the next 500–700 years. Soon after this, the island barrier will be completely eroded, and the coasts of the western Yamal will be subjected to direct influence of the sea, which will accelerate their recession.

The results obtained here agree with the previous conclusion that the recession of the Arctic coasts is predominantly related to the mechanisms of the cross-shore sediment transport. Thus, the further application of the prognostic model under the Arctic conditions will be related to the development and improvement of the methods for estimating the volume of the material transported seawards.

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