A method for assessing the coastline recession due to the sea level rise by assuming stationary wind-wave climate

by

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Abstract

The method introduced in this study for future projection of coastline changes hits the vital need of communicating the potential climate change impact on the coast in the 21st century. A quantitative method called the Dynamic Equilibrium Shore Model (DESM) has been developed to hindcast historical sediment mass budgets and to reconstruct a paleo Digital Elevation Model (DEM). The forward mode of the DESM model relies on paleo-scenarios reconstructed by the DESM model assuming stationary wind-wave climate. A linear relationship between the sea level, coastline changes and sediment budget is formulated and proven by the least square regression method. In addition to its forward prediction of coastline changes, this linear relationship can also estimate the sediment budget by using the information on the coastline and relative sea level changes. Wind climate change is examined based on regional climate model data. Our projections for the end of the 21st century suggest that the wind and wave climates in the southern Baltic Sea may not change compared to present conditions and that the investigated coastline along the Pomeranian Bay may retreat from 10 to 100 m depending on the location and on the sea level rise which was assumed to be in the range of 0.12 to 0.24 m.

Key words: modelling, coastline changes, dynamic equilibrium, sediment budget estimation, climate change

Introduction

Coastline and sea level changes become an increasingly important issue for about 40% of the global population who lives along the coast of the world ocean (Burke et al. 2001). An accelerated relative sea level rise can intensify erosion on unprotected coasts. During the 21st century, a continuous and accelerated rate of global mean sea level rise has been projected (IPCC 2013). Accordingly, there is a vital need for projecting coastline responses to the accelerated sea level rise scenarios and estimating the sediment budget for coastal zone management and coastal engineering protection activities.

Coastline recession is usually associated with sediment erosion and re-deposition in sandy/soft-rock cliff coastal areas on the decadal to centennial time scale. Complex process-based morphodynamic models (e.g. Walkden and Dickson 2008, Ranasinghe et al. 2011, Zhang et al. 2012) are able to display the effect of multi-scale hydrodynamic processes and sediment transport on coastline changes. However, these numerical process models usually require a large amount of data and computational capacity. Alternatively, in case of absence of data, a simple method based on the concept of the equilibrium coastal profile and the principle of sediment mass conservation can be used to provide a rapid estimation of coastline responses to sea level rise. Such a method can complement those process-based models for the coastal zone management, planning and engineering protection activities. This simple method has also a potential to be applied to understand a coastline behavior responding to rates of the sea level rise (Asthon et al. 2011) when the spatial scale is extended (Zhang et al. 2004). Bruun (1962, 1988) proposed a linear relationship between the sea level rise and coastline retreat, assuming an equilibrium coastal profile shape and two-dimensional sediment mass conservation. However, the Bruun rule model does not work when there are net external sediment sources and sinks of its two-dimensional model domain (Ranasinghe and Stive 2009). The assumptions behind the Bruun rule model have been debated extensively (e.g. SCOR 1991, Thieler et al. 2000, Cooper and Pilkey 2004, Pilkey and Cooper 2004, Ranasinghe and Stive 2009).

Recent studies tended to adopt the equilibrium

concept to develop a simple method to assess the coastline responses to sea level rise (Bray and Hooke 1997, Davidson-Arnott, 2005, Walkden and Dickson 2008, Asthon et al. 2011, Ranasinghe et al. 2012, Rosati et al. 2013). Furthermore, the unidirectional coastline retreat at the cliff coast facilitates the modelling of coastline responses to sea level rise (Woodroffe and Murray 2012). However, a similar method accounting for a dynamic equilibrium change of coastal profiles to assess the three-dimensional coastline responses to sea level rise, and also sediment budget, remains unexplored. A large-scale modelling approach (Dynamic Equilibrium Shore Model, DESM) based on the generalized Bruun concept of dynamic equilibrium coastal profile changes has been elaborated by Deng et al. (2014) to assess the historical sediment erosion and deposition pattern as well as sediment budget by using the available data sets of relative sea level records, modern DEM and historical coastline changes. By introducing the sediment mass balancing equation and assuming a constant wind-wave induced lateral sediment flux (i.e. total sediment volume change consisting of subaerial cliff and submarine erosion), the DESM model has a potential to be applied for future projection of coastline changes based on relative sea level rise scenarios (Deng et al. 2014).

The aim of this paper is to present a method to convert the inverse modelling technique adopted by the DESM model to the forward mode. This method assumes a linear relationship between the sediment budget, coastline and relative sea level changes. In this study, historical sediment budget is estimated by the DESM model for the time span of ~100 years, using the available data sets of modern DEM, historical coastline changes from AD ~1900 to AD 2000 and the relative sea level records by Deng et al. (2014). The idea of this study comes from the Bruun Rule derivation, but differs in the fact that it represents only the first-order approximation of dynamic changes of coastal profiles and its three dimensional feature. Interestingly, similarly to the Bruun Rule relationship, this linear relationship allowing for dynamic coastal profile changes can also assess the sediment budget by using the information on the coastline and relative sea level changes, which is useful for the coastal zone management and planning activities.

When this linear relationship is applied for the

future projection of coastline changes, a constant wind-wave induced lateral sediment flux has to be assumed to model the coastline recession in response to relative sea level changes. The first assessment of possible future changes of a lateral sediment flux can be made by examining the future changes of probability distribution of wind speed and direction. To test this assumption, we use regional climate modelling projections for our application area located in the southern Baltic Sea (Meier et al. 2011). In these simulations, model wind is bias-corrected by using a gustiness parameterization.

The paper is structured as follows. In section 2, we provide a description of the terms used in the model, the main concepts of the DESM model and the method to derive the forward mode of the DESM model. The analysis method of wind climate change is also given in Section 2. Section 3 describes the study area, data sources used for the DESM model for the past and future projections. In Section 3, we also present statistical analysis of climate-change modelling data of wind speed and direction, and a forward function as a result of a linear regression on the past reconstruction. Section 4 presents future scenarios of coastline changes. Section 5 discusses

how the forward mode presented in this paper differs from the model by Bray and Hooke (1997), and fits into the general framework of potential cliff coastline responses to rates of the sea level rise established by Asthon et al. (2011). The limitations of the method are also discussed in Section 5. Section 6 summarizes our conclusions.

Materials and methods

Brief description of the DESM model for the past reconstruction

Sediment erosion and deposition pattern and sediment budget estimated by the DESM model requires the input data of modern DEM, relative sea level changes and historical coastline information. The DESM model presented in Deng et al. (2014) consists of four parts: the generalized Bruun concept, modelling of coastal retreat, modelling of coastal accretion and the inverse modelling technique adopted in this approach. Here, we recall several key terms used in the DESM model for budgeting the sediment mass volume (Fig. 1) and

Figure 1

Terminology for the DESM model. Note that tidal effect is not included in the DESM model

a brief description of this numerical approach. The three-dimensional generalized Bruun concept is also illustrated in Fig. 2. For a given time interval, only net sediment transport behaviors and average driving forces are considered in this model. At the retreating coast, the eroded sediment of a sub-aerial cliff or dune is transported cross-shore to the deeper sea, where part or all of it may be further transported by longshore sediment transport processes. At the advancing dune coast, the onshore sediment transport process translocates the sediments from the sea to the land for the dune formation process.

represented by one cross-shore profile (Fig. 2). Each zone starts from a terrestrial location which remains untouched by hydrodynamics during a given time span, and extends seaward until the so-called “closure depth” (i.e. for a given or characteristic time interval, it is the most landward isobath of which there is no significant change in bottom elevation and no significant net cross-shore sediment exchange is measurable, according to the definition by Krauss et al. (1999)). By this definition, each zone can be characterized by a cross-shore profile and along-shore width. The information on the cliff height

Figure 2

Sketch to show the three-dimensional concept of the DESM model

The model domain is a semi-enclosed area in terms of sediment budget, when the lateral sediment flux $Q_{lst,1}$ (m^3) and $Q_{lst,m}$ (m^3) at the domain's lateral boundary is zero (i.e. $Q_{lst,1} = Q_{lst,m} = 0$) in Fig. 2. However, it may have a minor coast-to-basin cross shore sediment flux. The model domain of the DESM model consists of eroding and accreting coast segments. The whole model domain is thus discretized alongshore into $n+m$ zones, each zone

H_{cliff} (m), dune height H_{dune} (m), and the coastline change c (m) are regarded as key parameters in the assessment of the sub-aerial sediment mass budget. For the submarine part, the closure depth H_* and the associated closure distance L (m) are defining the offshore boundary of the DESM model domain. The submarine profile is approximated by the exponential function where the paleo-curvature parameter is regarded unknown. By defining the

uniform ratio between paleo- and present-curvature coefficients of all cross-shore profiles, the paleo-coastal morphology can be simulated by applying the principle of sediment mass conservation. This ratio is the only unknown parameter after obtaining the information from the modern DEM, the relative sea level and historical coastline changes. Therefore, the mathematical expression of sediment mass conservation can be described by Eq. (1):

$$\left| \sum_{i=1}^m V_{erosion,i}(const1) \right| = \left| \sum_{j=1}^n V_{accretion,j}(const2) \right| \quad (1)$$

In this equation, the Left Hand Side (LHS) represents the total mass volume lost from the eroded coastal zones and the Right Hand Side (RHS) represents the total mass volume accumulation at the accreted coastal zones. *Const1* is the uniform ratio between paleo- and present-curvature coefficients of all cross-shore profiles at the retreating coastal zones, and *const2* is the ratio at the advancing coastal zones. The value of *const1* can be determined by the iterative inverse procedure when *const2* is set to 1, or *const2* can be determined by the iterative inverse procedure when *const1* is set to 1. For the full treatment of the DESM model, we have to refer to the paper by Deng et al. (2014), as the past reconstruction is not the main objective of this paper.

Method for deriving the forward mode of the DESM model

The forward mode of the DESM is formulated by 1) deriving the component of the sea level rise effect on coastal erosion through the morphodynamic equations, and 2) creating a link between the morphodynamic equations and the DESM model. The DESM model uses coastal profiles as basic computation units (Fig. 2). To link the coastal profile changes with sediment transport mathematically, a sediment mass balancing equation (e.g. Roelvink and Reniers 2012) is adopted here:

$$\frac{\partial Z_b}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0 \quad (2)$$

where Z_b is the bed elevation (m). q_x and q_y are respectively the longshore and cross-shore volumetric sediment flux per unit (m^3). Equation (2) can be converted into the form incorporating

relative mean sea-level changes. By using water depth h (m) instead of the bed elevation Z_b (Fig. 1), the water depth h is related to the relative mean sea level: $Z_b = RSL - h$:

$$\frac{\partial(RSL - h)}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} = 0 \quad (3)$$

which can be integrated over the closure distance, L (Fig. 1),

$$V'_{external} + s * L + Q_{y,cliff/dune} - Q_{y,ba \sin} + Q_x = 0 \quad (4)$$

where s is the relative mean sea-level change (m) and L is the closure distance (m), $V'_{external}$ is the sediment mass volume per coastal profile (m^3) when s is zero. $V_{external} = V'_{external} + s * L$ is the overall submarine sediment mass volume per coastal profile (m^3). $Q_{y,cliff/dune}$ (m^3) stands for the sediment mass volume eroded from the subaerial cliff or dune being a function of cliff height H_{cliff} or dune height H_{dune} and coastline change c (m), as described in Deng et al. (2014).

$$Q_{y,cliff/dune} = c (h_{cliff_foot} + h_{cliff} + s / 2) \quad (5)$$

$Q_{y,ba \sin}$ is the basin-ward sediment sink (m^3) that is usually minor in the sandy coastal environment. Q_x is the lateral sediment flux per coastal profile (m^3). Q_x is equal to the sum of submarine and subaerial volume changes when $Q_{y,ba \sin}$ is neglected. $V'_{external}$ is assumed to be a function of coastline change c (m). The longshore sediment flux Q_x determines the large-scale shoreline change (e.g. Murray et al. 2013), when the relative sea level change is disregarded. Therefore, Equation (4) can be rewritten as:

$$V'_{external}(c) + sL + Q_{y,cliff/dune}(H_{cliff}, c) + Q_x = 0 \quad (6)$$

$V'_{external}(c)$ is the first-order polynomial function of coastline change c by the least-square regression method on the past modelling data:

$$V'_{external,i} = E * c_i + F \quad (7)$$

where $V_{external,i}$ is the numerical value of the function $V_{external}(c)$ at the i th coastal profile, E and F are the coefficients of the slope and intercept of the linear function, and c_i is the coastline recession distance at the i th cross-shore profile. The numerical value of F expresses the eroded sediment mass volume when the relative sea level and coastline changes are both zero. The numerical value of E is the approximated slope of the linear function between the sediment volume and coastline changes. When the static equilibrium profile is adopted, E is equal to the closure depth, as the Bruun Rule model.

Substituting Equation (5) and Equation (7) into

$$V_{external,i} = E * c_i + F + s_i l_i + c_i (h_{cliff_foot,i} + h_{cliff,i} + s_i / 2) = -Q_{x,i} \quad (8)$$

Equation (6) yields:

where $V_{external,i}$ and $Q_{x,i}$ is the total eroded sediment mass volume of coastal erosion at the i th cross-shore profile. Equation (8) can be converted to a forward formula to predict the coastline recession distance c_i :

$$c_i = \frac{V_{external,i} - F - s_i l_i}{(E + h_{cliff_foot,i} + h_{cliff,i} + s_i / 2)} \quad (9)$$

For the future projection, coefficients E and F remain unchanged, and the parameter $H_{cliff,i}$ can be obtained from the modern DEM. In Equation (9), $V_{external,i}$ is approximately equal to the lateral sediment flux $Q_{x,i}$, which can be obtained from the application of the DESM model on modelling the paleo-coastal DEM.

Under the condition of a stable wind-wave induced lateral sediment flux, the future coastline recession distance is solely determined by the future relative sea level rise to have a balancing sediment budget, which is similar to the approach by Bray and Hooke (1997). When assuming a zero net lateral sediment flux, the static cross-shore equilibrium profile and zero sediment exchange with the subaerial part, E is equal to the closure depth h_s and F is zero. Equation (9) can be simplified to the original Bruun Rule model: $|c_i * h_s| = |s_i l_i|$.

Method for analyzing the wind climate change

The regional future projections for the Baltic Sea by Meier et al. (2011), which have adjusted model wind using a gustiness parameterization are driven by boundary data from the global General Circulation Models (GCMs) forced with the IPCC greenhouse gas emission scenarios A1B and A2 (Nakicenovic and Swart 2000). The applied regional climate model (RCM) called the Rossby Centre regional Atmosphere and Ocean Model (RCOA) (Döscher et al. 2002, 2010) is a fully coupled atmosphere-ice-ocean-land surface model. Table 1 lists the investigated scenario simulations, driving GCMs, emission scenarios, and the time spans. The analyzed data are 3-hourly wind speeds covering the time span from AD 1960 to AD 2099.

Table 1

Data of wind climate projection from AD 1960 to 2100 (Meier et al. 2011)

Scenarios	Forcing data	Emission scenarios	Time span
1	ECHAM5 (MPI-Met, Germany)	A1B	1960-2099
2	HadCM3 (Hadley Centre, UK)	A1B	1960-2099
3	ECHAM5 (MPI-Met, Germany)	A2	1960-2099

The analysis of climate changes is designed as follows. We analyze the wind speed components in two orthogonal directions: east-west and north-south. The period of AD 1960-2009 is selected to represent the control period (the past) as the modern DEM data set is measured between the years 2000 and 2009. The period of AD 2050-2099 is selected to represent the future. Ratios of wind speed probability of the control and future periods are plotted as histograms to show the changes. The year to year variability of histograms is depicted by adding the standard deviation to each bar as whiskers. We divide the 50-year period into 50 one-year periods. Then, we have 50 values for each bar and could compute the standard deviation among these 50 values. By using these 50 histogram values for each bar (each range of wind speed), the two-sample *T*-test was conducted to test changes in histograms between the control and future periods at 5% and 1% significance levels. In order to test whether there is a statistically significant change of

wind speed other than the probability or frequency, we also carried out the Mann-Whitney *U*-test for the time series of wind speed at the 5% significance level. Changes in wind direction are given by comparing the wind roses.

Application

Application area

The coastal system of the southern Pomeranian Bay is part of the Oder River outflow system described in general by Borowka et al. (2011). This system consists of the Oder River mouth and the Szczecin Lagoon which is separated from the Pomeranian Bay in the north by two barrier Islands: Usedom Island in the west and Wolin Island in the east. The Szczecin Lagoon discharges into the Baltic Sea (Pomeranian Bay) through three gates: the Swina River to the north, the Peenestrom to the west, and the Dziwna River to the northeast. The whole system is framed by the Pleistocene Highlands in the west and the east. The shape of the present coastal landscape is a result of a continuous process having started already during the Late Pleistocene when the Paleo-Oder River incised its valley into the headlands of glacial sediments. During the Holocene Littorina transgression onset at around 8000 cal. BP, the coastal erosion started to deliver its product to the formation of the Swina Gate barrier from the moraine cores of Usedom Island and Wolin Island. The Swina Gate barrier was built up stepwise by progradational sand accumulation from west-south to west-north directed from Wolin and from east-south to east-north directed from Usedom Island. The Baltic Sea is tideless, and wind wave-induced longshore sediment transport is the main sediment transport mechanism along the southern Baltic coast. As in Fig. 3, the longshore sediment transport capacity simulated by using the CERC formula (USACE 1984) has clearly shown that the area consisting of Wolin Island in the west and Usedom Island in the east can be regarded as a semi-enclosed area in terms of sediment budget. The driving forces of waves are winds, mainly from the west, while the shelter effect of the Usedom Island (limited wind fetch) has prevented the full development of waves to the east in this area. There is a nodal point of the semi-enclosed coastal system at the Dziwna River

mouth where a longshore component of wave power is close to zero and in the east of this point. In this “cliff-dune” coastal system of the Pomeranian Bay, major sediment sources for the dune development come from the adjacent cliff coast. This “cliff-dune” coastal system is thus an ideal site to apply the concept of three-dimensional sediment mass conservation (Deng et al. 2014). A validated 2DH wave model for the period of 1997-2006 by Zhang et al. (2013) indicates that the wave breaking height ranges from 1.5 m to 2 m in this area, which is mainly caused by storms. Zhang et al. (2011a) did a statistical analysis of the high-resolution hindcasted winds at the southern Baltic Sea covering the period from 1948 to 2007 (with a temporal resolution of 1 hour and spatial resolution of 25 km provided by Weisse et al. (2009)). The analysis indicates that the change either in the direction or the strength on a decadal scale is so small that its impact on the morphodynamics can be neglected.

Data used for the application of the DESM model for the past and future projections

The hindcast of the coastal morphology serves as input data for the future projection. To apply this mass balancing model, the high resolution modern DEM, relative sea level records and historical coastline changes are needed. The model DESM has been applied to generate the DEM in the 1980s and the DEM in AD ~1900 by Deng et al. (2014). In this application, the validation of the inverse model of DESM has been conducted by comparing with the nautical chart data in the 1980s and the results of the multi-scale long-term morphodynamic model BS-LTMM (Deng et al. 2014). Here, we describe the input data needed for the past and future projection by applying the DESM model and its forward mode.

A high-resolution modern DEM of the study area (as shown in Fig. 3) is constructed using five different datasets: (1) a bathymetric dataset for the Swina Gate nearshore area (including the artificial channel) with a spatial resolution of 15 m × 15 m; (2) a bathymetric dataset covering the whole Pomeranian Bay with a spatial resolution of 50 m × 50 m; (3) a terrestrial DEM for the Swina Gate barrier with a spatial resolution of 15 m × 15 m; (4) a terrestrial DEM for the coast of the Island of Usedom and the Island of Wolin with a spatial resolution of 25 m × 25 m; and

Figure 3

The Pomeranian Bay: (a) the wind rose plotted based on the climate modelling data; (b) Baltic Sea bathymetry; and modern DEM and (c) longshore sediment transport capacity for the period 1957–2006: direction and magnitude

(5) high-resolution (~1 m) cross-shore bathymetric profile measurement data with a spacing of 500 m alongshore and length of ~2000 m. Tide gauges are densely distributed along the Baltic Sea and have been providing the long-term continuous and precise measurements of a relative sea level. The

relative sea level rise in this area is around 1.1 mm yr^{-1} for the period of AD 1811–2000 on the basis of a tide gauge measurement at the Swina River mouth by Rosentau et al. (2007). By using the same tidal gauge data, Richter et al. (2012) determine a recent relative sea level rise of 1.3 mm yr^{-1} for the period of

AD 1908-2007. In the Baltic Sea region, melting of the Scandinavian Ice Sheet (SIS) caused the isostatic uplift of the Fennoscandian Shield and subsequently a subsidence in the southern Baltic (Harff et al. 2007). The subsidence rate in the study area is less than 0.2 mm yr^{-1} (Harff et al. 2011). The historical coastlines are reconstructed using the German topographic map “Messtischblatt” at a scale of 1:25 000. “Messtischblatt” maps used in this study were measured during the period from the nineteenth century to AD 1945, when the coastal area of the research key areas belonged to the Prussian state. These “Messtischblatt” maps from the 1880s to the 1920s are available for the public on the websites of (http://greif.uni-greifswald.de/geogreif/?coll_id=72&map=overview) and Adam Mickiewicz University in Poznań (http://mapy.amzp.pl/tk25_list.cgi?show=all;sort=g). By a comparison with a modern aerial photograph and topographic map, the geo-referenced maps have an accuracy of 5 m to 10 m (Deng et al. 2013).

Application of the method to convert the inverse modelling adopted by the DESM model to the forward mode

Analysis of climate modelling data from AD 1960 to AD 2099

In this study, future projections for AD 2100 are based on the assumption about a stationary wind-wave climate. There are several sophisticated wind wave models such as SWAN (Booij et al. 1999) to simulate the near-shore wave dynamics. In the near shore area, coastal morphology is dynamically changing from a seasonal to decadal time scale. Therefore, when incorporating the morphological change, real time simulations of wave dynamics in the Baltic Sea for more than 100 years are computationally expensive and can only be conducted by implementing the multi-scale morphodynamic model like the Baltic Sea – Long Term Morphodynamic Model by Zhang et al. (2011a). Furthermore, small changes in wind climate have almost no impact on the coastal morphodynamic development on decadal to century time scales as indicated by Zhang et al. (2011a). Therefore, by adopting the previous research results in this area, this study presents a simple method to analyze

changes in wind fields available from the regional climate models to examine a possible coastal morphodynamic response to these changes. Several climate modelling studies focusing on the future wind climate change over Europe have indicated that there is no consensus on future mean wind climate change by AD 2100 (e.g. Von Storch et al. 2008). In this study, results of recent climate simulations with high spatial and temporal resolution for the Baltic Sea region (e.g. Meier et al. 2011, 2012; Schimanke et al. 2014) have been analyzed.

The overall impression of these histogram comparisons with whiskers representing standard deviations in Fig. 4 is that there is always an overlap of whiskers between the control and future periods. In the central parts of the histograms, the ratios are almost close to 1; and in all scenarios listed in Table 1, the probability of winds blowing toward the east is increasing. However, such probability changes are not significant in all ranges of wind speed. In scenario 1, for example, the probability changes are statistically significant in the east-west wind speed ranges of $(-4, 0)$ and $(6, 12)$ at the 5% significance level. However, by setting the significance level at 1%, these probability changes in the three scenarios are almost statistically insignificant by using the two-samples *T*-test. The histogram of north-south wind speeds appears to be more stable than the east-west wind speeds. The Mann-Whitney *U*-test results show that changes in north-south wind speed is insignificant in scenario 1 and 2 and the north-south wind speed significantly increases in scenario 3, whereas the east-west wind speed in scenario 1 and 3 significantly increases and the change is insignificant in scenario 2. The analyses of the strength and frequency of wind speed are consistent. In conclusion, our analysis shows that the signal of wind climate changes is weak in terms of the probability and strength of wind speed.

The ratios in Fig. 5 also show apparent probability changes in northerly extreme wind speeds, but these changes are not significant according to the result of the two-sample *T* tests. However, the limited and unequal samples of extreme wind speed frequency between the two periods may make the result of the two-sample *T* tests incorrect. By plotting the wind roses for both the control and future periods, we found that in all these scenarios there may be an increasing probability of extreme winds from the

Figure 4

Comparisons of the probability distributions of east-west wind speed (positive value means wind directed to the east) and north-south wind speed (positive value means wind directed to the north) between the period of AD 1960-2009 and AD 2050-2099: a) Scenario 1: ECHAM5, A1B; b) Scenario 2: HadCM3, A1B; c) Scenario 3: ECHAM5, A2. The GCMs forcings of the three scenarios are listed in Table 1. Note that the right y-axis is the ratio of wind speed probabilities of the future and control periods and filled triangles means that changes are significant at the 5% significance level.

west. This increase is due to changes in the east-west component of winds according to the above analysis shown in Figure 4. But these changes might not have an important impact on the coastline changes. The extreme northerly winds in the study area are able to generate southerly waves advancing the southern Baltic Sea coast. As evidenced by our morphodynamic modelling results, the extreme wind events with an incoming direction ranging from NNW to NE play an important role in the

coastal changes (particularly coastal erosion) of the study area (Deng et al. 2014). However, Fig. 5 shows that changes are mainly in the wind direction range from N-W to S-W. The changes shown in Fig. 5 should not have a significant impact on coastal morphodynamics. To sum up, our analysis shows no remarkable changes in the average wind climate and we regard it as a support of our assumption about stationary wind climate.

Figure 5

Wind roses comparison between the control and future periods in the three scenarios (unit: m)

Forward function for assessing the coastline responses to sea level rise

In the simple Bruun rule model assuming a two dimensional equilibrium profile, the sediment mass volume change has a perfect linear correlation with a coastline recession distance and the sea level rise: $|c_i \cdot h_i| = |s_i l_i|$. While considering the dynamic changes in coastal profiles, the perfect linear correlation between the sediment budget and coastline change is impossible to obtain. Therefore, the method is only a linear approximation of these dynamic coastal profile changes. In the eroded part of the Wolin Island from AD ~1900 to AD 2000, the sediment

budget was already estimated by the DESM model (Deng et al. 2014).

Fig. 6 presents the coefficients of E and F obtained by the least square regression using the data of historical coastline and relative sea level changes and the associated sediment budget. The relatively high R^2 value of 0.82 indicates that the linear function is sufficient to reflect the primary relationship between the coastline and sediment mass volume changes. Substitution of E and F values in Equation (7) yields the total submarine sediment volume change $(V_{external})_i$ at the i th profile:

$$(V_{external})_i = V'_{external,i} + s_i l_i = 15.4 \cdot c_i - 2268.6 + s_i l_i \quad (10)$$

Figure 6

Linear relationship between $V'_{external,i}$ and coastline change c_i on basis of the modelled sediment mass volume change from AD ~1900 to AD 2000 by the DESM model

The empirical Equation (10) means that the submarine sediment mass volume can be determined with the input of coastline recession distance c_i between present and historical coastlines, closure distance l_i and relative sea level rise s_i . However, this forward equation is derived from the data of a given period. To test whether this function is valid when implementing it in a different period, its predicted sediment mass volume change from AD 1829 to 2000 is then compared with the one

produced by the DESM model. In this comparison, c_i and s_i are time-dependent and the closure depth of 12.45 m is still used for the reconstruction of the morphological past in AD 1829. The first version of the Prussian topographic map “Ur-Messtischblatt” published in AD 1829 is used to reconstruct the historical coastline. The accuracy of this historical map is around 40 m (Deng et al. 2013). The relative sea level change s_i is set to $\sim 1.2 \text{ mm y}^{-1}$ in this area (Rosentau et al. 2007). The iterative inverse modelling procedure of the DESM yields a result that *const1* is 0.63 for the eroded coastal segment and *const2* is 1 for the accreted coastal segment. This value of *const1* is smaller than the one obtained for the period AD ~1900 to AD 2000. This difference indicates a tendency for dynamic changes in coastal profiles in this area since AD 1829.

The comparison of volume changes between Equation (10) and the inverse modelling result by the DESM model is presented in Fig. 7. The Root Mean Square Error (RMSE) between the results of these two methods is 793 m^3 that accounts for $\sim 25\%$ of the average volume approximated by the DESM model. This satisfied coincidence between the two methods and the high R^2 value in Fig. 7 indicates that the derived forward model (Equation (10)) represents

Figure 7

Comparison of volume changes from AD 1829 to AD 2000 estimated by the DESM model and the Equation (10) for the Swina Gate area. For the location, see Figure 3.

a centennial dynamic relationship between the sediment volume, coastline and relative sea level changes, and the closure distance. By including the subaerial part of coastal change, Equation (10) is thus rewritten to estimate the coastline recession distance:

$$c_i = \frac{V_{external,i} + 2268.6 - s_i l_i}{(15.4 + h_{cliff_foot,i} + h_{cliff,i} + s/2)} \quad (11)$$

$V_{external,i}$ representing a lateral sediment flux can be either derived from the measured data or the DESM model result. Parameters $h_{cliff,i}$ and $h_{cliff_foot,i}$ can be obtained from the modern DEM. For example, for the projection of coastline changes from AD 2000 to AD 2100, $V_{external,i}$ can be obtained by the paleo approximation of the DEM at AD ~1900.

Results

Scenarios of future coastline changes

On the southern Baltic coast, the scenario of relative sea level rise by AD 2100 is assumed to be that the eustasy might be accelerated by a factor of two

compared with the 20th century (Meyer et al. 2008), whereas the rate of isostasy remains unchanged. This scenario of a relative sea level has also been used by Zhang et al. (2011) for the morphodynamic projections. The coastal section along Wolin Island was selected for the application of Equation 11, as coastal processes are not intervened significantly by the coastal protection activities during the last century, which is disregarded in this model. By projecting the same eustatic sea level rise of 0.12 m in the past 100 years to AD 2100, an almost linear rate of coastline retreat is induced. The projection of future coastline changes is presented in Fig. 8. Apparently, coastline responses to the relative sea level rise are spatially variable. From sites 61 to 62.5, an accelerated sea level rise of 2.4 mm yr⁻¹ for the 21st century has just induced an increment of less than 10 m in the coastline retreat distance, and an increased increment of up to 40 m from sites 63 to 71. From sites 61 to 62.5, the value of $s_i l_i$ only accounts for not more than 20% of the lateral sediment flux $V_{external,i}$ due to a high cliff height and short closure distance compared with the eastern part of the coast. The coastal section between sites 67 and 73 has low cliff heights and a long closure distance of more than 6000 m. The longer closure distance means the larger accommodation space induced by the relative

Figure 8

Projection of coastline changes by AD 2100 on the basis of the scenarios of eustatic sea level rise of 1.2 mm yr⁻¹ and 2.4 mm yr⁻¹ at the Swina Gate area. The comparisons with Bruun rule, modified Bruun model, and SCAPE approximation (m=0.5) are also given. Note that the isostatic subsidence varies around ~ -0.1 mm yr⁻¹ here. For the location, see Figure 3.

sea level rise. Here, the value of $s_i l_i$ accounts for more than 40% of the lateral sediment flux $V_{external,i}$, which results in strong impact of relative sea level rise. These results indicate that the future coastline changes in this area are determined by the competition between the wind-wave induced lateral sediment flux and the relative sea level rise.

Comparison with other simple models adopting the concept of equilibrium coastal profile

The forward mode of the DESM model is also compared with the original Bruun rule model, the modified Bruun model by Bray and Hooke (1997) and SCAPE (Soft Cliff And Platform Erosion) approximation ($m=0.5$) by Walkden and Dickson (2008). The formulations of the latter two models are presented in the Appendix. In the comparisons (Fig. 8), the Bruun rule model results in the largest estimates of coastline retreat compared to other models. The Bruun rule model overestimates the forward mode of the DESM model by a factor of about 2.0 in this study area. However, the forward mode of the DESM model, the modified Bruun model and the SCAPE approximation produce quite similar results. In particular, the DESM model result is only slightly smaller than the ones obtained by the modified Bruun model with a mean difference of -1.7 m. The mean difference between the forward mode of the DESM model and the SCAPE approximation results is 6.5 m. This comparison thus shows that the forward mode of the DESM model has an estimation between the modified Bruun model and the SCAPE approximation ($m=0.5$). In case of the increasing rates of the relative sea level rise by AD 2100, Fig. 9 shows that the differences between the linear forward mode of the DESM model and the SCAPE approximation ($m=0.5$) are accelerated. However, higher rates of relative sea level rise may need more time to attain an equilibrium state as it is shown in the model experiments by Walkden and Dickson (2008). This constraint may also apply to the forward mode of the DESM model, as the assumption about a constant longshore sediment flux may not be valid due to the high accelerated rate of the sea level rise of more than 3 times compared to the past. But, such high acceleration of the sea level rise may not happen in this study area.

Figure 9

Comparison of the model DESM, original Bruun rule model and modified Bruun model and SCAPE approximation ($m=0.5$) at the point 69 of Figure 8. For the location, see Figure 3.

Discussion

The paper presents a method to convert the DESM model to the forward mode. Whether a coastal profile shape is dynamic or not makes a difference between the forward mode of the DESM model and the modified Bruun Rule model for the cliff coast by Bray and Hooke (1997). The mean difference between the two models is only -1.7 m in our case study area. Below, we show the connection of two models in the equation forms. If the coastal profile shape is stable, the slope E of the linear function is equal to the closure depth h_* and the ordinate intercept F is zero. Equation (9) thus becomes:

$$c_i = \frac{V_{external,i} - s_i l_i}{(h_* + h_{cliff_foot,i} + h_{cliff,i} + s/2)} \quad (12)$$

By neglecting the relative sea level rise s and cliff foot height $h_{cliff_foot,i}$ that are minor compared with cliff height $h_{cliff,i}$ and closure depth h_* in the denominator, Equation (12) can be transformed to the model by Bray and Hooke (1997):

$$c_i = \frac{V_{external,i} - s_i l_i}{(h_* + h_{cliff,i})} \quad (13)$$

The method presented in the paper allowing for

the dynamic changes in coastal profiles has a linear slope E that is not equal to the closure depth h_c . The linear slope E depends on the curvature ratio of *const1* or *const2* in Equation (1) of coastal profiles. In the application area, E of 15.6 m is larger than the closure depth h_c of 12.45 m as the *const1* is less than 1. Therefore, a different numerical value of E also reflects the change of coastal profiles. As the coastal profile is not static, the ordinate intercept F means that there may be still erosion in the submarine profile even if there are no coastline changes. Such extreme cases may occur by the combined effects of storm erosion and re-accretion at the beach in the normal weather conditions, or because there are large amounts of sediment sources from subaerial cliff erosion. In our study area, the process-based model results by Deng et al (2014) show that strong submarine erosion takes place at the coastal segment between points 59 and 62 caused mainly by the storms, but the coastline recession distance is decreasing to almost zero from points 62 to 59.

Astthon et al. (2011) established a power-law relationship to understand the future potential response of a wide variety of cliff coastal environments (consolidated or unconsolidated, soft- or hard-rock) to the rate of sea-level rise by employing a different exponent value of m (Appendix 2). By adopting this general framework, Fig. 9 shows that the presented erosion formula of Equation 11 lies between the case of “negative response” (SCAPE approximation: $m=0.5$) and “instance response” (original Bruun rule model: $m=1$). Brooks and Spencer (2012) compared the equilibrium coastline response models for soft-rock cliff coasts, and the results show that the SCAPE approximation between the rates of sea level rise and coastline recession fits best the experimental data but are still similar to the modified Bruun model. The presented model gives a plausible estimation between the SCAPE approximation and the modified Bruun model, and provides an alternative tool to account for the variability of coastal profile morphology. Moreover, the presented linear function can be a rapid estimation method of sediment budget by using the information on coastline and relative sea level changes for the long-term coastal planning and coastal zone management.

In addition to those discussed in Deng et al. (2014) for the past projection, this method has several limitations. The main limitation of this method is

the assumption about a constant lateral sediment flux between historical and future periods. Here, a test of this assumption is conducted by comparing modeled wind for the periods of AD 1960-2009 and AD 2050-2099. We recognize some changes in increasing westerly wind speeds and changes in extremes. However, especially changes in extremes vary among the different climate realizations, which make it difficult to draw any general conclusion. This is in agreement with Nikulin et al. (2011). Moreover, our analysis takes the coastline geodetic orientation into consideration to explore possible impacts of changing wind climate on coastal morphological changes in a specific area. This combined analysis of coastline geodetic orientation and wind direction is then supported by our previous process-based modelling study in this area (Deng et al. 2014). Furthermore, Zhang et al. (2011a) indicate that small changes in the direction and strength of wind on a decadal scale can be represented by the stationary wind climate and storm frequency distribution, when coastal morphodynamic evolution is simulated on a decadal to centennial scale. Therefore, despite the fact that we do not analyze the wave climate change, the wind analysis supports the hypothesis about the “stationary” wind climate. Moreover, possible weak changes in the wind have almost no impact on the coastal morphodynamic evolution at this specific study area (i.e. the Pomeranian Bay, southern Baltic Sea). However, multi-decadal variations of wind climate may exist in this region (HELCOM 2013). Furthermore, effects of wind gustiness on wave climate in this coastal area remain unclear, despite that Abdalla and Cavaleri (2002) found that the gustiness effect is limited on enclosed seas like the Mediterranean Sea due to its fetch-limited conditions. Therefore, future studies may incorporate changes in the lateral sediment flux driven by wind-wave climate changes. Such lateral sediment flux changes can be parameterized by defining future scenarios of $V_{external,i}$ in Equation (9). An approach of Bonaldo et al. (2015) assuming a static equilibrium coastal profile, may be used for the first approximation to provide possible future changes in the lateral sediment flux as well as cross-shore sediment transport during storm events.

Further consideration for applying this method is that there is an implied assumption about a stability of coastline shape features that have a stable

tendency for coastline changes. The instabilities can be incurred by 'high-angle' waves (Ashton et al. 2001; Ashton and Murray 2006a, 2006b). The wave climate and coastline orientations are spatially variable on the southern and eastern Baltic Sea coasts. The identification of coastal segments with coastline stability in this region can be studied by zero-crossings of the net longshore sediment flux (Soomere and Viška 2014).

Conclusions

A method for projecting coastline changes due to the relative sea level rise is presented by using historical data of paleo-coastline changes, modern DEM and relative sea level scenarios. This method is formulated by converting the inverse modelling of the DESM by Deng et al. (2014) to the forward mode by using a sediment mass balancing equation. A deterministic relationship between the relative sea level rise, coastline changes and submarine sediment volume changes by the linear least square regression method is created. The deterministic function was examined by approximate reconstruction of the DEM in AD 1829, which shows that it is able to estimate volume changes similar to the inverse modelling results. When applying the forward mode of the DESM model to the future projection, it successfully predicts an almost constant rate of coastline retreat in the scenario of a constant rate of sea level rise. By comparison with the SCAPE approximation and the modified Bruun model that are tested against the real data by Brooks and Spencer (2012), the presented model gives a plausible assessment between the SCAPE approximation and the modified Bruun model. This comparison also shows that the presented forward mode lies between the case of "negative response" (SCAPE approximation: $m=0.5$) and "instance response" (original Bruun rule model: $m=1$) in the general framework of potential coastline responses to the rate of sea level rise by Ashton et al. (2011). The forward mode of DESM provides an alternative method to assess the coastline recession responding to the relative sea level rise by accounting for morphological changes in coastal profiles, and an easy estimation of the sediment budget by using the information on the coastline and relative sea level changes. This is useful for long-term coastal planning and coastal zone management (for example, beach

nourishment). Further improvement of the method may incorporate the changes of wind wave climate in the model and consider variability of the coastline geometrical shape.

Reliable future coastline scenarios are extremely important to the coastal zone management and coastal planning (e.g. Dickson et al. 2005, Zhang et al. 2012). This requires modelling tools that are suitable for variable coastal environments. However, even if such models with a wide applicability are developed, they still require the input data of the driving forces such as a relative sea level and wind climate changes. The climate system on Earth is highly complex and influenced by plenty of internal and external factors (HELCOM 2013). Anthropogenic influences are also highlighted in the IPCC report (IPCC 2013). There may be still uncertainties in the numerical downscaling approach from GCM to RCM that may cause some results difficult to interpret (Bellafore et al. 2012). However, modelling of future coastline responses to climate change needs accurate ranges of future climate scenarios. This is essential and the climate change model is the only tool that can provide this kind of data.

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Appendix

1. *Modified Bruun rule model by Bray and Hooke (1997):*

$$R_2 = R_1 + \frac{(S_2 - S_1)L}{P(B + h_*)}$$

where P is the proportion (here, it is assumed to be 1) of cliff sediment that remains within the active profile (% sand and gravel) and B is cliff elevation (m); R_1 is the historical retreat rate (m yr^{-1}) and S_1 is the rate of historical relative sea level rise (m yr^{-1}); R_2 is the future retreat rate and S_2 is the future rate of relative sea level rise.

2. *Potential coastline responses to the rate of sea-level rise at the cliff coast by Ashton et al. (2011):*

$$R_2 = R_1 \left(\frac{S_2}{S_1} \right)^m$$

where R_1 and R_2 are historical and future retreat rates (m yr^{-1}), and S_1 and S_2 are the rates of historical and future relative sea level rises (m yr^{-1}). Different values of m describe various types of cliff coast whose potential coastline responses to the rate of sea level rise in the future. When $m = 0.5$, the equation represents the Soft Cliff And Platform Erosion (SCAPE) approximation by (Walkden and Dickson 2008). When $m=1$, the equation is the original Bruun Rule model.