Influence of wind climate changes on the mean sea level and current regime in the coastal waters of west Estonia, Baltic Sea*

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Ülo Suursaar
Tiit Kullas

Estonian Marine Institute,
University of Tartu,
Mäealuse 10a, EE–12618 Tallinn, Estonia;

e-mail: ulo.suursaar@ut.ee

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Abstract

The response of semi-realistic wind speed increase scenarios to the mean sea level and current regime of semi-enclosed sub-basins in the Baltic Sea is studied with a 2D hydrodynamic model. According to the model output of spatial mean sea levels, an increase in the westerly wind component by 2 m s^{-1} leads, for example, to a mean sea level rise of up to 3 cm in windward locations in the study area. The sea level change patterns depend on the wind scenario and coastline configuration. The increases in wind speed considered here also lead to enhanced water exchange through the straits, strengthening of the basin-scale circulation, enhancement of up- and downwelling, and increased bottom stresses near coasts.

1. Introduction

Climate change is a complex and spatially inhomogeneous process which, apart from global or regional warming, manifests itself through various other phenomena, including changes in the wind climate. Recently, climatologists have indeed revealed some significant changes in the regional wind regime

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above the North and Baltic Seas. The increase in the winter westerly wind component and intensification of cyclonic activity has been reported e.g. by Alexandersson et al. (2000), Gulev et al. (2002) and Jaagus (2006). These developments are associated with changes in the winter NAO-index during the last half-century; in the summer, however, no significant trends in circulation patterns have been recorded.

Compared to the global mean trend estimates (1–2 mm yr^{-1}, e.g. Gornitz et al. 1982, Woodworth 1990, Ekman 1999), the sea level trends are steeper (1.5–2.5 mm yr^{-1}, according to Suursaar et al. 2006a) in the windward coastal waters of Estonia. Also, vitalisation of coastal processes and alteration of depositional coasts have been observed in recent decades in Estonia (Orviku et al. 2003). These changes can occur mainly as a result of changes in the local wind climate. However, local historical wind data do not provide sufficiently reliable quantifications of actual changes in wind speed and directional distribution: such time series are very sensitive to building activity and changes in vegetation around the meteorological stations. Additional problems are posed by the inconsistencies introduced by instrument changes, e.g. from mechanical weathercocks to automatic anemorhumbometers, as well as modified definitions of mean and gusty wind speeds. The changes in wind climate are further confirmed by changes in the frequencies of local storm-days (Orviku et al. 2003, Jaagus 2006), by the increased mean zonal (west–east) component of the geostrophic airflow above northern Europe (e.g. Ekman 1998, Siegismund & Schrum 2001), as well as by Estonian upper-air wind data. According to aerological soundings, the zonal component of December–February upper-air winds increased by about 3 m s^{-1} above Estonia in 1954–98 (Keevallik & Rajasalu 2001). Besides that wintertime change, the only statistically significant changes occurred in March, when the meridional component changed its sign as a result of enhanced southward flow and reduced northward flow. This may be due to a northward shift in cyclone trajectories in the course of the last 50 years (Sepp et al. 2005).

In general, the change estimates based on trends in annual or seasonal mean wind speeds have remained around 1–3 m s^{-1} during the last fifty years. Also, the simulation results of climatologic models (e.g. ECHAM4, CGCM2) predict further warming by the 2080s, as well as an increase of up to 2–3 m s^{-1} in the mean westerly wind component in winter and spring, and an increase of up to 1–2 m s^{-1} in summer and autumn wind speeds above northern Europe (e.g. Ulbrich & Christoph 1999, Jylhä et al. 2004, Räisänen et al. 2004). The gridded model results of climate and related sea level variables are downloadable from the IPCC Data Distribution Centre.

Some projections for the Baltic Sea level are presented by e.g. Johansson et al. (2001) and Wroblewski (2001).

1–3 m s^{-1} is also the magnitude of changes in wind speed we applied to our hydrodynamic model by modifying its forcing conditions. It would be important to acquire knowledge of the consequences of such changes (observed in the past and anticipated in the future) in the wind climate on the hydrodynamic regime of the sea. Does such a seemingly quite small change affect the mean sea level and current patterns at all? The regional ('climatological') Baltic Sea level rise component has been discussed mainly in relation to changes in NAO (e.g. Ekman 1998, Andersson 2002, Stigebrand & Gustafsson 2003); only a few studies have used direct hydrodynamic models (e.g. Schrum 2001, Meier et al. 2004). We carried out a short study of past sea level trends and the possible reactions of sea level to changes in wind forcing conditions (Suursaar et al. 2006a). In that investigation we produced time series of modelled sea levels at a few selected points in the west Estonian coastal sea and studied mainly temporal changes in sea level in relation to different 'scenario runs'. Indeed, the annual mean wind-driven sea level appeared to rise by up to 5 cm in some windward bays in our study area (Fig. 1). In this study we present results concerning

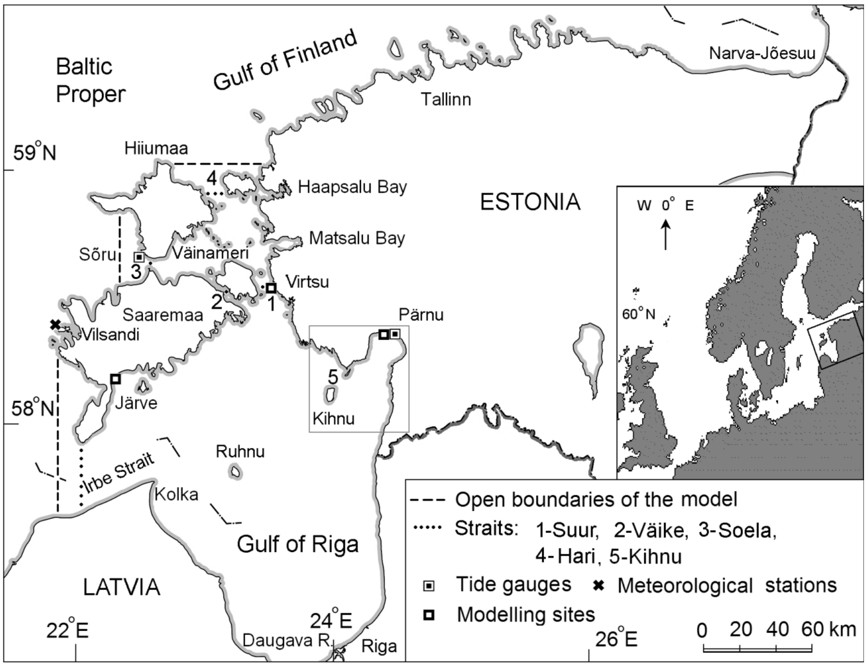

Fig. 1. Map of the study area with model boundaries and stations for forcing data. The Pärnu Bay – Kihnu Strait sub-region is marked separately

spatial changes in the mean sea level based on similar scenario runs, as well as on some additional scenarios. Possible changes in the mean flow regime are also discussed.

The main objectives are: (1) to investigate the reaction of the sea level and flow regimes to changes in different semi-realistic and constructed wind regime changes; (2) to study some spatial change patterns in the sea level and wind-driven circulation; (3) to study possible changes in the flow regime and to discuss briefly their influence on coastal environmental processes, such as mixing, coastal erosion and deposition.

2. Study area

The prerequisite of this hydrodynamic modelling study is the distinctive semi-enclosed shape of our study area, which allows relatively short open boundaries to be used for the model domain (Fig. 1). It consists of two nearly tideless brackish-water (average salinity 5.6 PSU) sub-basins: the Gulf of Riga and the Väinameri Sea (also called the Moonsund or Archipelago Sea). The Gulf of Riga measures roughly 140×150 km^2 and has a surface area of 17 913 km^2. The Väinameri is approximately 50×50 km^2, with a surface area of 2243 km^2. The maximum depth of the Gulf of Riga is 52 m (av. 23 m). The Väinameri is even shallower with an average depth of 4.7 m. The sub-basins are connected to each other through the nearly meridional Suur Strait, and with the Baltic Proper by three straits lying in the SW (Irbe Strait), W (Soela Strait) or NW (Hari Strait) directions. Both sea level variations and hydrodynamic processes are therefore subject to the prevailing westerly winds and the passage of cyclones. Water exchange through the whole system involves the contribution of all the four major straits (Otsmann et al. 2001). According to our extensive in situ current measurements in the Suur Strait in 1993–95, the gross-mean velocity modulus was 25 cm s^{-1} and the maximum measured value was 100 cm s^{-1} for both in- and outflowing currents (Otsmann et al. 2001). In general, both the flow direction and magnitude varied considerably depending on wind conditions, but the annual sum of northward flows (165 km^3) exceeded southward flows, yielding an annual accumulated outflow from the Gulf of Riga of about 30 km^3. In order to maintain this balance, roughly similar quantities should also exit via the Hari Strait and enter via the Irbe Strait. With a cross-section area of 0.37 km^2, the latter is the widest strait in the system, where under certain conditions in- and outflows can take place simultaneously (Lilover et al. 1998, Talpsepp 2005).

Because of the semi-enclosed configuration of the study area and the presence of some shallow bays exposed to the direction of the strongest possible winds, sea level variations here include a considerable local component. Besides the Baltic Sea areas of St. Petersburg in the NE and Schleswig in the SW, some of the highest storm surges are registered there. The two highest sea level events off the Estonian coast (since 1924) were both registered at Pärnu: 253 cm above the Kronstad zero on 19 October 1967 and 275 cm on 9 January 2005 (Suursaar et al. 2006b). The lowest sea levels were also measured at Pärnu – 125 cm below the Kronstad zero. Bearing in mind that Pärnu is also an important port and summer resort, we have placed special emphasis on this region in the present study (Fig. 1).

3. Material and methods

3.1. Hydrodynamic model

We use a two-dimensional (2D) hydrodynamic model, applied and briefly described in some of our previous studies (Suursaar et al. 2002). It is a type of shallow sea depth-averaged free-surface model, consisting of momentum balance and volume conservation equations:

$$\frac{DU}{Dt} - fV = -g(H+\xi)\frac{\partial \xi}{\partial x} + \frac{\tau_x}{\rho_w} - \frac{kU}{H^2}(U^2+V^2)^{1/2}, \tag{1}$$

$$\frac{DV}{Dt} + fU = -g(H+\xi)\frac{\partial \xi}{\partial y} + \frac{\tau_y}{\rho_w} - \frac{kV}{H^2}(U^2+V^2)^{1/2}, \tag{2}$$

$$\frac{\partial \xi}{\partial t} + \frac{\partial U}{\partial x} + \frac{\partial V}{\partial y} = 0, \tag{3}$$

$$\frac{D}{Dt} = \frac{\partial}{\partial t} + \frac{1}{H}\left(U\frac{\partial V}{\partial x} + V\frac{\partial}{\partial y}\right), \tag{4}$$

where U and V are the vertically integrated volume flows in the x and y directions respectively, ξ is the sea surface elevation (above the equilibrium depth, H), f is the Coriolis parameter, ρ_w is the water density, k is the bottom frictional parameter ($k = 0.0025$, e.g. Jones & Davies 2001), and τ_x and τ_y are the wind stress $\vec{\tau}$ components along the x and y axes. The wind stress ($\vec{\tau}$) was computed using the parameterisation by Smith & Banke (1975), which includes a non-dimensional empirical function of the wind velocity:

$$C_D = (0.63 + 0.066 \mid \vec{W_{10}} \mid)10^{-3}, \tag{5}$$

where $\mid \vec{W_{10}} \mid$ is the wind velocity vector modulus [m s^{-1}] at 10 m above sea level.

The model domain encompasses the Gulf of Riga and the Väinameri sub-basins with a model grid horizontal resolution of 1 km, yielding a total of 18 964 marine grid-points (including 2510 in the Väinameri). The Latvian bathymetric database (Berzinsh et al. 1994) and Estonian nautical maps were used to establish the bottom topography and the line of the coast. Wetting and drying in response to variations in the sea level were not included. A staggered Arakawa C grid was used with the positions of the sea levels in the centre of the grid box and velocities at the interfaces. At the coastal boundaries the normal component of the depth mean current is considered to be zero. The effect of neglecting the historical viscosity is negligible in our case. The model equations were numerically solved using the finite difference method with an integration time step of 30 seconds. Because of the chosen time step and horizontal grid resolution, the numerical diffusion generated by the numerical scheme was relatively low (the Courant number is approximately 0.6).

The 2D model performance was previously studied in comparison with the Helmholtz model (Otsmann et al. 2001) and flow measurements in the straits from 1993–95 (Kullas et al. 2000). Since the depth of the study area is less than 50 m (and effectively of the order of 5–10 m at the study sites), the use of a 2D model is acceptable. Moreover, hindcast simulations for 1999 and 2005 demonstrated the model's success in simulating sea levels (Suursaar et al. 2002, 2006b). Outside the straits, in situ flow measurements for comparison with the model are not available. This study has therefore not taken small-scale flow features into consideration, which may appear as modelling artefacts imposed e.g. by the digitalisation of the coastline and the bathymetry. The inferences drawn from the numerical simulations are quite general for the circulation.

3.2. Forcing conditions and scenario runs

The model was forced by the local wind and open-boundary sea-level data. As we are dealing with changes in wind-driven hydrodynamics, the primary factor forcing currents is the wind. Another important factor forcing sea level in the semi-enclosed study area is the Baltic (border-) sea level (Otsmann et al. 2001), especially on a time-scale longer than one or two days. The local wind is important in shorter periods and particularly in the case of storm surges (Suursaar et al. 2002). It is also responsible for the specific difference from the Baltic Proper sea level. The remaining factors (tides, seiches, precipitation-evaporation, river inflow, inverse-barometer response, thermohaline effects, etc.) are less important for this study. Their influence is considerably smaller than that of the two primary factors stated above. Moreover, since the aim of this study was to investigate specific aspects of the wind-driven circulation and sea level regime, the focus was on differences between the control and scenario runs (see Table 1), with less attention being paid to other possible factors (see also Suursaar et al. 2006a).

Table 1. Explanation and statistical data of the wind forcing time series used in the sea level modelling: average wind speeds (AV), maxima (Max) and standard deviations (SD) in m s^{-1}, increase in average wind speed compared to CR (AV Inc., %)

Scenario	Explanation for forcing scenario	Wind forcing statistics in 1999			
		AV	AV Inc.	SD	Max
CR	Measured Vilsandi wind 1999	5.85	0%	3.42	24.0
SRM	2 m s^{-1} added to moduli (except 0)	7.83	33%	3.42	26.0
SRS	2 m s^{-1} added to S–N component	6.43	10%	3.63	25.9
SRN	−2 m s^{-1} added to S–N component	5.82	−1%	3.36	22.1
SRW	2 m s^{-1} added to W–E component	6.36	9%	3.79	24.8
SRE	−2 m s^{-1} added to W–E component	5.88	1%	3.20	23.4
SRSW	2 m s^{-1} added to S and W components	6.91	18%	3.98	26.6

The model has relatively short open boundaries, which were shifted by 5–20 km outside the narrowest parts of the straits. Fluxes across the boundaries were allowed. Although the model performs well within the sub-basins, the results obtained between the open boundary and the straits are probably less reliable. We used hourly-measured sea level time series obtained from the Sõru tide gauge, which is located just outside the Soela Strait. The Sõru data (Fig. 2a), applied at the three breaks in the open boundaries near the Irbe, Soela and Hari Straits, were used in exactly the same way as the sea level boundary conditions in all the control and scenario runs for the year 1999.

The wind stress was calculated from the wind data measured at the Vilsandi meteorological station (58°22′59″N, 21°48′55″E). This station is located on an island west of the Estonian mainland (Fig. 1), so the data is not affected by the presence of land to the east. The data had a 1 m s^{-1} value interval, a 10° angular resolution and a 6-hour time step subsequently interpolated into the hourly interval. A spatially homogeneous wind was applied to all the grid-points of the modelled area. This is justified, as the west Estonian meteorological stations in our relatively compact study area generally display rather coherent wind data (Soomere 2001, Suursaar et al. 2006b).

The wind scenarios used in the study (Table 1) were based on 1999 data and included a realistic simulation, regarded as the ‘control run’ (CR), and

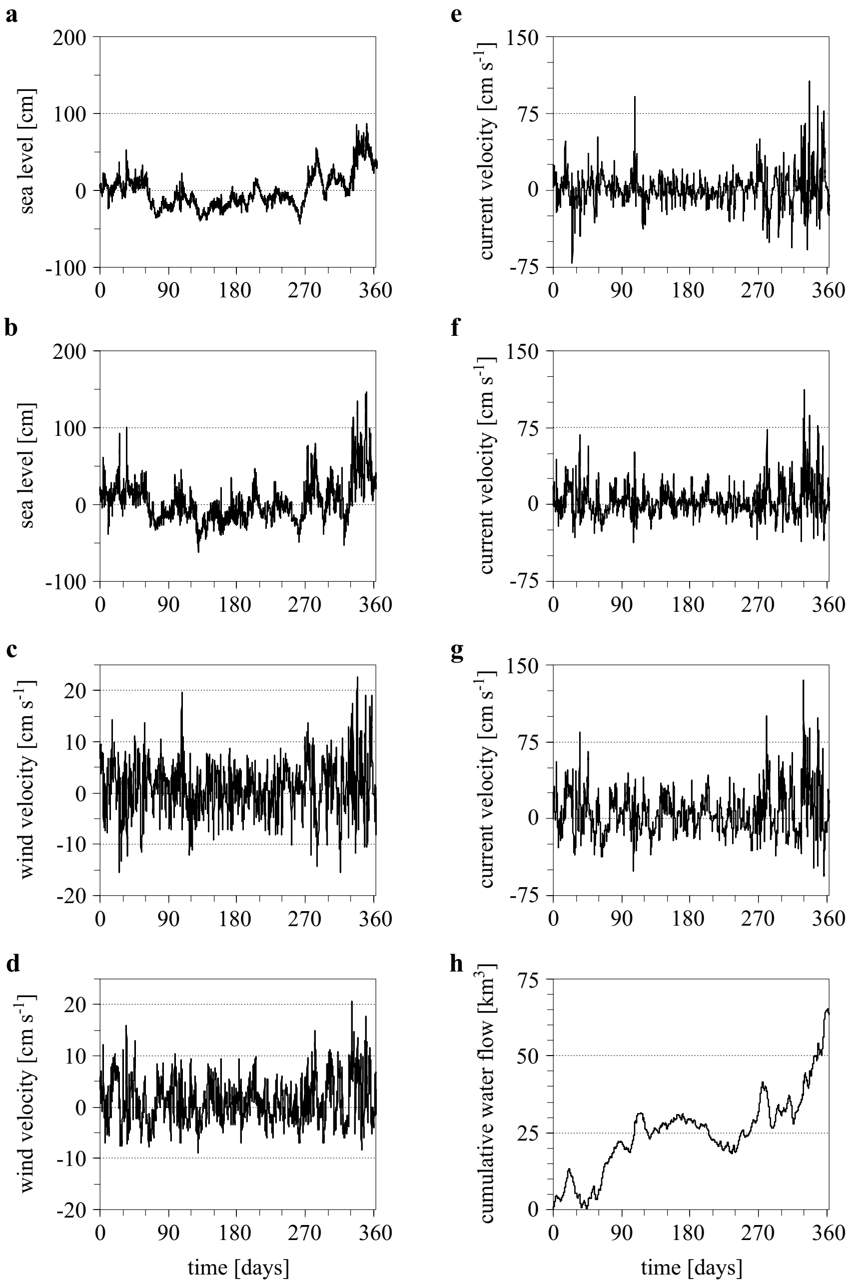

Fig. 2. Time series of measured sea level variations at Sõru (a) and Pärnu (b); Vilsandi wind speed u (c) and v (d) components; modelled currents at Pärnu (u-component) (e), Virtsu (v-component) (f), and Järve (v-component) (g); modelled cumulative flows through the cross-section of the Suur Strait (h) in 1999. The positive direction is eastwards for the u-component and northwards for the v-component

several semi-realistic ‘scenario runs’ (SRM, SRN, SRS, SRW, SRE, SRSW) with slightly modified wind forcing. These scenarios are basically the same as in the previous study (Suursaar et al. 2006a), except that the SRN-scenario was added. (The main differences lay in the spatial model outputs of this study and in the current outputs.) The directional distributions of the winds are identical for the CR and the SRM scenarios, as 2 m s^{-1} was added to the wind speed regardless of its direction. The five other modified 1999 scenarios emphasised the N, S, W, SW and E directions and reduced the occurrence of S, N, E, NE and W winds, respectively. The actual wind speed increment was nearly 2 m s^{-1} in the case of the SRM scenario, but less in the other cases. The resulting increase in annual average wind speed (Table 1) was somewhat smaller than in the sea level modelling scenarios used by Schrum (2001) and Meier et al. (2004). We stress here that the magnitude of the wind speed increase we have considered is quite realistic, but we do not unconditionally expect such changes in the future. Although many studies forecast a further increase in westerly winds, it is also possible that the increase in westerlies and winter NAO in 1950–90 could be followed by a recession in westerlies (i.e. relative increase in easterlies) (e.g. Hagen & Feistel 2005).

The annual means were calculated on the basis of hourly time series of sea level and current moduli for each forcing scenario at all the grid points. In order to study the possible influence on coastal geomorphology, time series of current speeds (8760 hourly readings) at some grid points near the coast were also produced (Fig. 1). A set of sensitivity runs with constructed wind forcing schemes was performed to investigate the steady state flow and sea level patterns depending on different wind directions and speeds. Possible current speeds in the straits were studied depending on wind speed and direction, a snapshot of quasi-stationary flow patterns was produced for the Pärnu Bay – Kihnu Strait area, and flow conditions in selected points of this area (Fig. 1) were investigated.

4. Results and discussion

4.1. Control run for 1999

The statistics of Vilsandi wind data used in the modelling study are shown in Table 1 (scenario CR) and in Fig. 2c, d. Typically, W–SW winds prevailed in 1999 with a maximum sustained wind speed of 24 m s^{-1} (gusts of up to 32 m s^{-1}) blowing from the SW in December. At Sõru the measured sea level varied between −43 and +87 cm, while near Pärnu, it ranged from −62 cm to +146 cm (Fig. 2a, b).

In 1999 the average measured sea level was −0.9 cm at Sõru and 3.2 cm at Pärnu. According to CR (Fig. 3a), this difference is due mainly to the annual wind statistics. CR also serves as the further basis for the 'scenario runs' (Figs 3–5). It shows the annual mean sea level above the borderline (i.e. Sõru) sea level. The difference is the largest (3 cm) in the bays of

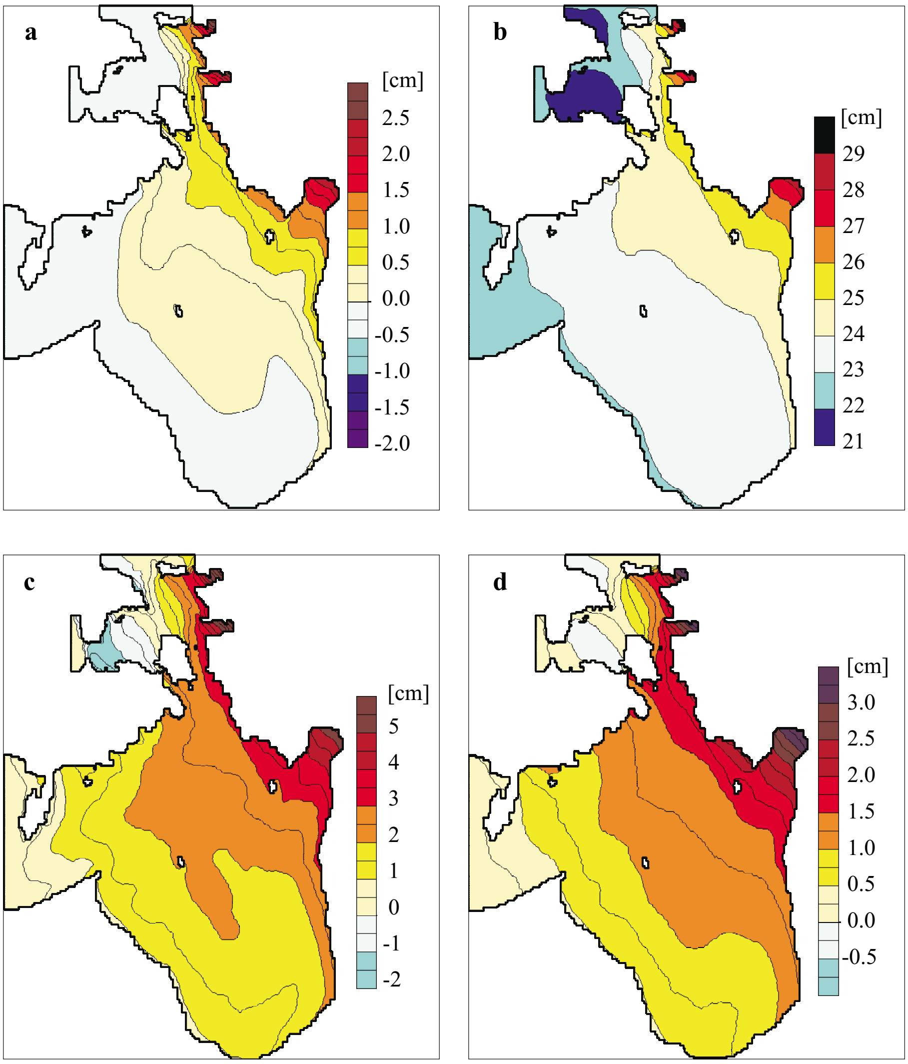

Fig. 3. Horizontal distribution of mean sea level, modelled as the control run (CR), above the open boundary sea level (a); modelled (as CR) standard deviations (b); study period mean sea level differences calculated between the SRSW and CR simulations (c); differences in standard deviations between the SRSW and CR simulations (d); see also Table 1

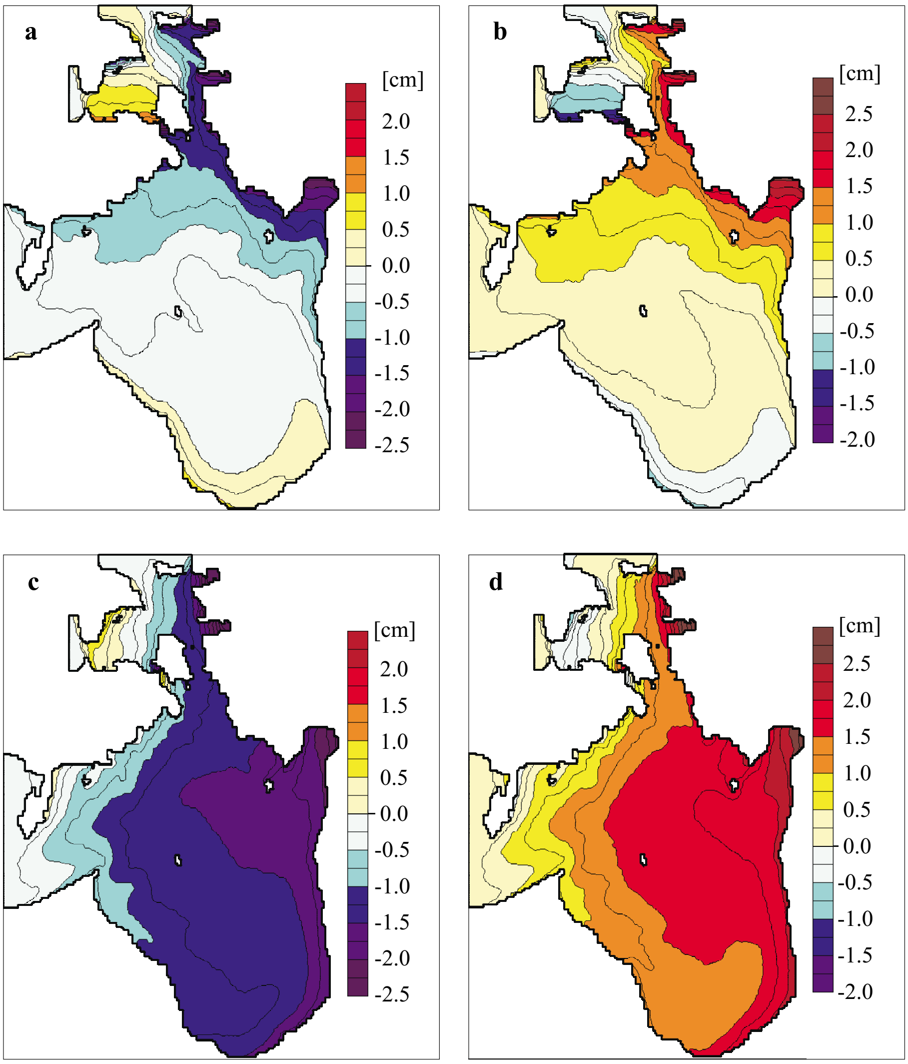

Fig. 4. Horizontal distribution of study period mean sea level differences between SRN and CR (a); SRS and CR (b); SRE and CR (c); SRW and CR (d)

Pärnu, Matsalu and Haapsalu, which are exposed to the climatologically most expected SW and W winds.

Compared with the modelled differences, the somewhat larger difference between the measured sea levels (4.1 vs 3 cm between Sõru and Pärnu) can be explained by the influence of the inflow from the River Pärnu, which averages 60 m^3 s^{-1}; according to our simulations, however, the influence of this inflow is very small (< 1 mm), even at the point of inflow entry. Nevertheless, since the Pärnu tide gauge is located 1–2 km upstream on

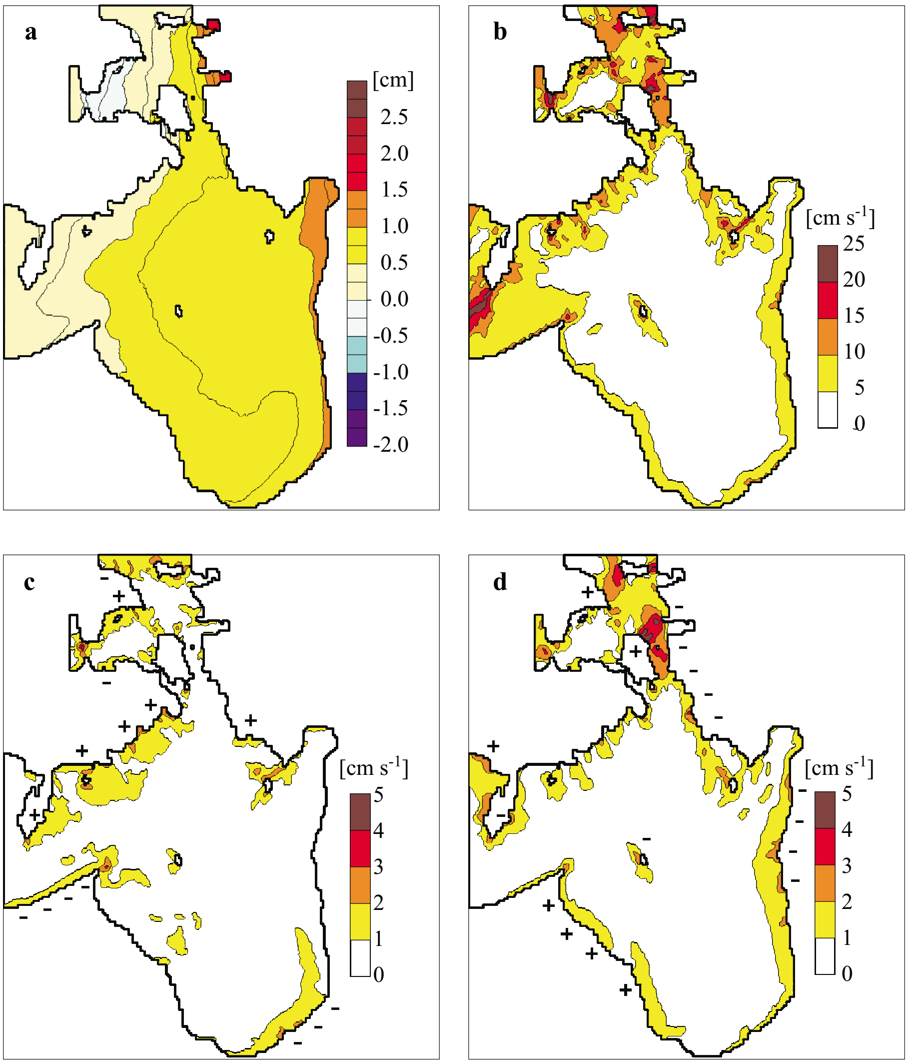

Fig. 5. Horizontal distribution of modelled mean sea level differences between SRM and CR (a); horizontal distribution of the mean 2D flow moduli, modelled as the control run (CR) (b); horizontal distribution of modelled 2D current moduli differences between SRW and CR (c), and SRS and CR (d). Zones of expected downwelling (−) and upwelling (+) enhancement are marked on (c) and (d)

the Pärnu, a bias of 1–2 cm is possible. The mean flow velocity at the 5 grid points surrounding the entry point of the Pärnu is up to 2–3 cm s^{-1} as a result of the inflow, and decreases further with distance. The influence of the River Daugava is probably more pronounced, but still applies only to the vicinity of the river mouth. Hence, though our study deals mainly with wind-driven climatological effects, we can assume that river inflow has a very small and local impact on the sea level and current regime in the study area. The impact could possibly be smaller than that reported earlier (e.g. Myrberg & Andrejev 2003) for the Gulf of Finland.

The standard deviations (based on hourly data) are large (28–30 cm) in the bays and smallest (21–22 cm) near the open boundaries (Fig. 3b). For comparison, the standard deviations of the Baltic Proper sea level vary roughly between 17–18 cm near the Swedish coast and 25–30 cm at the far (windward) heads of the Gulf of Finland and Bay of Bothnia (Lazarenko 1961, Johansson et al. 2001).

The flow simulations express stronger currents and water exchange during the meteorologically highly variable last months of the year (Fig. 2e–h). The modelled cross-section average velocity reached 1 m s^{-1} and the maximum flow rate was –2.1 km^3 day^{-1} in the Suur Strait. Entering currents prevailed in the Irbe Strait and exiting currents in the Suur and the Hari Straits. The mean current moduli remained below 30 cm s^{-1} in all the grid-cells, with maximum velocities in the straits (Fig. 5b). The currents were also somewhat stronger (usually 5–15 cm s^{-1}) near the coast, but in the extensive central parts of the sub-basins the mean velocities were < 5 cm s^{-1}. The maximum current moduli never exceeded 30 cm s^{-1} in the same areas. However, maximum flow readings reached up to 1.6 m s^{-1} in the Suur and Hari Straits and even up to 2 m s^{-1} at a few grid-points of the Soela Strait.

4.2. Changes in sea level

While the difference between the open boundary sea level and realistic-year (1999) hindcast simulation levels (CR) expresses the specific influence of wind forcing on the sea level (Fig. 3a), the differences between CR and scenario runs (Table 1) express the mean sea level change components as a result of hypothetical changes in the wind climate (Figs 3c, d, 4, 5a).

The results from different scenario runs demonstrate that every change in the long-term average wind speed, wind variability (even if the average remains unchanged), or directional distribution of the wind, has a specific effect on the established sea level regime of a location (Suursaar et al. 2006a). The effect is different along the coastline, depending on the morphometric features of the straits (direction in relation to prevailing winds, water exchange capacities), the dimensions of the sub-basins, and aspects of the configuration of the bays, such as the distance of the head of the bay from the nodal point of the sub-basin, breadth convergence and depth reduction. It is easy to see that if northerly winds strengthened by 2 m s^{-1} (Table 1), the annual mean sea level would increase somewhat in the southern part of the Gulf of Riga and in the Väinameri ahead of the Väike Strait road dam. The sea level would decrease by up to 3 cm in the leeward bays (Fig. 4a). Roughly opposite mean sea level changes would occur if the wind regime underwent changes with enhancement of S winds (Fig. 4b). A somewhat larger effect appears in the SRE and SRW scenarios (Fig. 4c, d), as the major strait, the Irbe Strait, lies in a roughly zonal direction.

The SRM scenario, which increased every wind event regardless of its direction, produced a sea level increase as well (Fig. 5a). The SRSW scenario, which included both increased S and W wind components, produced the most pronounced effect (Fig. 3c). Interestingly, even if the sea level was raised by up to 5–6 cm in some locations, there were areas where the sea level decreased. This happened on the leeward coasts of Hiiumaa Island and in the western part of the Väinameri Sea owing to the limited flow capacity of the Soela Strait. SRWS also produced the largest increment in standard deviations (Fig. 3d), and the largest increase in annual maximum sea levels. Although the average wind speed increased by 18% (Table 1), the standard deviations in the western part of the Väinameri surprisingly decreased by up to 0.3 cm (Fig. 3d). Since the relevant sea level mechanisms there lie simultaneously in a regionally windward and a locally leeward location, they partially cancel each other out.

It is also worth mentioning that the sea level differences exhibit short-term and seasonal variations, which are discussed in more detail by Suursaar et al. (2006a). While an up to 5–6 cm sea level increase can occur in the annual mean sea level (Figs 3, 4), the rise is up to 9–11 cm in winter, which sees more storms. In summer months, the sea level change is small, implying a further strengthening of the seasonal signal in sea level records, which has been already reported on the basis of measurements e.g. by Ekman & Stigebrandt (1990), Suursaar et al. (2006a).

Our analysis predicted a specific sea level change within our well-defined study area. We should expect an additional sea level rise component within the Baltic Proper as well, because the water inflow in the Danish Straits is sensitive to the same wind change scenarios. Intensified westerlies and storminess will cause an additional mean volume surplus in the Baltic Sea, and a change in sea level inclination, which is due to the fjord-like shape of the Baltic Sea. According to simulations by Meier et al. (2004), an up to 3–4 cm sea level rise component would occur in the central Baltic (outside our study area) in the case of a 30% increase in wind speed. Considering both the regional (Baltic) and the local sea level change components discussed in this study, a total average sea level rise of 8–10 cm could occur e.g. at Pärnu as a result of a change in the wind regime component, the magnitude of which is quite realistic and has been forecast by many authors.

4.3. Changes in the current regime

Since every change in wind speed and direction affects the sea level regime, it also affects wind-driven currents and water exchange through the straits (Figs 6, 7). We have investigated whether these changes are substantial and whether they have any significant effect.

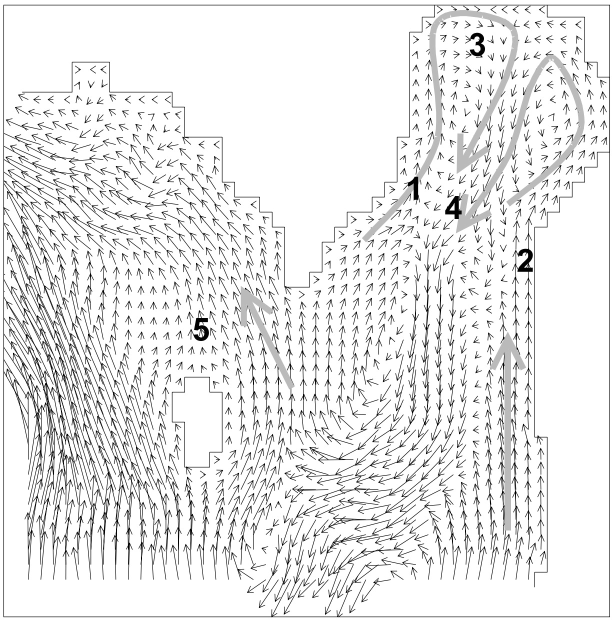

Fig. 6. Flow patterns produced by a 20 m s^{-1} stationary wind from the south; the arrow lengths are proportional to the flows, which are integrated from the surface to the bottom; modelling locations (appearing in Fig. 7a, b) are marked as '1', '2', '3', and '4'; Virtsu Strait is '5'

Because of the shallowness and small density differences in the study area, the currents there are determined mainly by the wind and modified by coastline and bottom topography. In the central Gulf of Riga, the influence of the Coriolis force and the basin scale flow pattern in the form of a topographic wave is more visible (Raudsepp et al. 2003). We concentrate on the straits and coastal areas, where velocities are larger and a possible influence on coastal processes should be expected. In the straits and the small, oval-shaped sub-basins (bays), the flow patterns follow certain simple rules. In the narrow straits, the flow velocity and direction are determined mainly by the projection of the wind vector on to the direction of the straits' axis (e.g. Otsmann et al. 2001). Except in the wide Irbe Strait, the currents rarely flow counter to the wind in the straits of Väinameri and in the Kihnu Strait (Fig. 7).

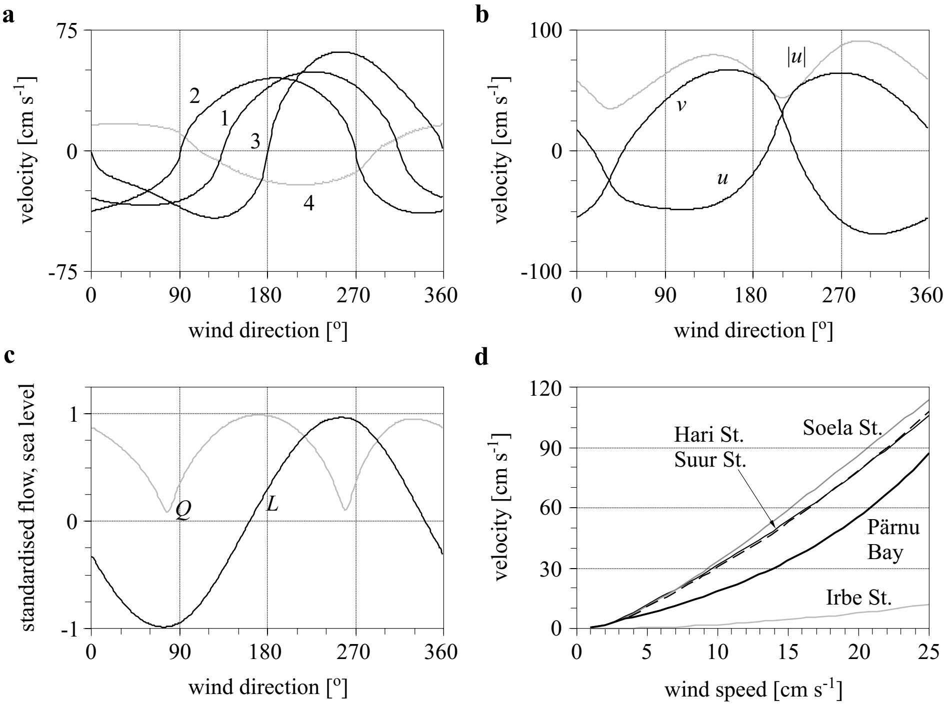

Fig. 7. Modelled dependence between the direction of the stationary wind (20 m s^{-1}) and current velocity components in Pärnu Bay (a, see also Fig. 6); between the wind direction and velocity components (u, v) and moduli $(|u|)$ in the Kihnu Strait (b); between wind direction and normalised flow through the Suur and Soela Straits, Q, and normalised sea level in the Väinameri, L, (c). Dependence of cross-section mean current velocities on the speed of the wind (d) blowing from 160° for the Hari and Suur Straits, from 250° for the Irbe Strait, and from 270° for the Soela Strait and for Valgerand Point (Pärnu Bay)

Simulations with constructed stationary and homogeneous wind conditions revealed two main, well-defined and persistent basin scale flow regimes and two mixed or transitional regimes for Pärnu Bay (Figs 6, 7a). Similar rules apply to the Gulf of Riga (Raudsepp et al. 2003). The modelled quasi-stationary flow-field (Fig. 6) shows cyclonic and anticyclonic circulation cells and a pattern with downwind flows near the shallower coastal areas and compensatory flows against the wind along the deeper middle section of the Bay. Such a flow pattern is typical of winds with a southerly component

(160–250°), while those with a northerly component give rise to reversed patterns in Pärnu Bay.

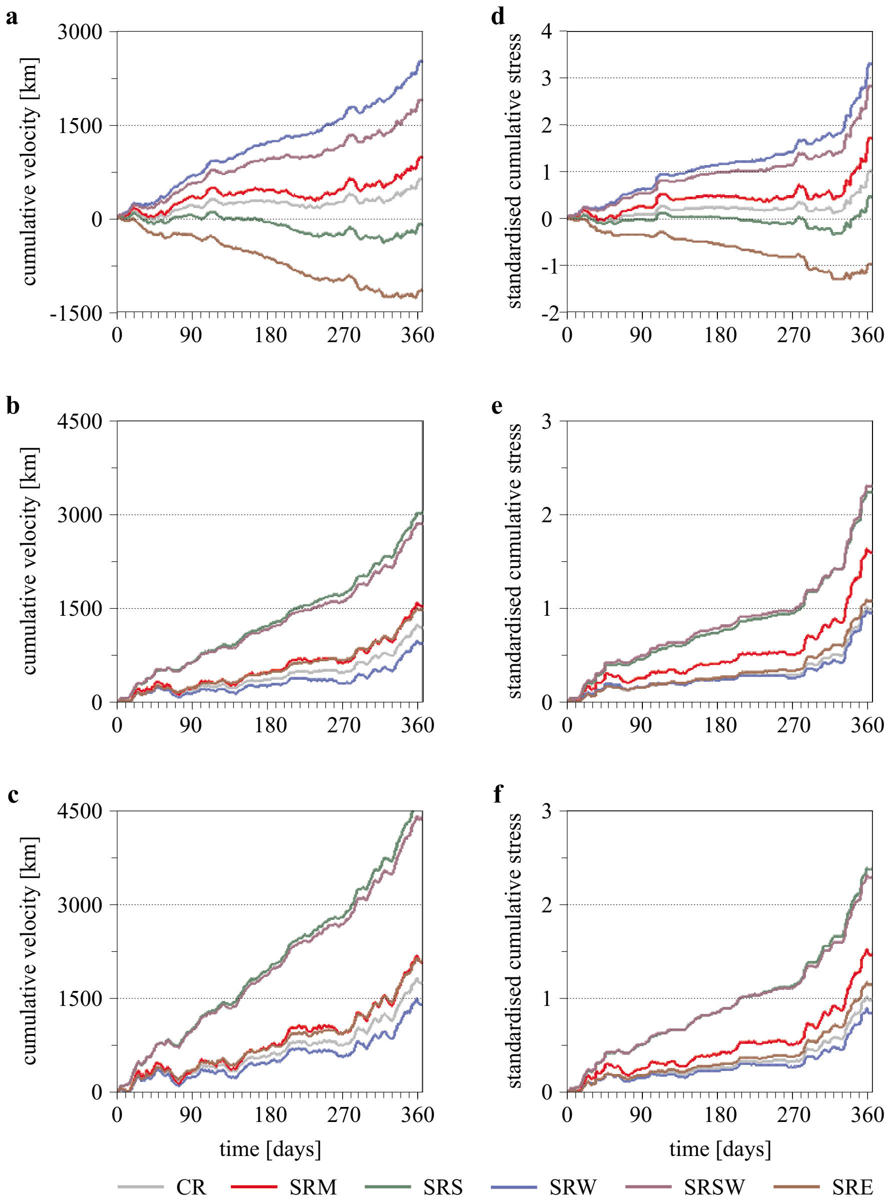

Fig. 8. Cumulative (integral) curves of current velocity vector components obtained with selected forcing scenarios (see Table 1) at Pärnu (a), Virtsu (b) and Järve (c) in 1999; corresponding variations in standardised (CR = 1) bottom stresses (d, e, f). See also Fig. 2, page 368

Hence, if we consider changes in the wind regime, the increase in wind speed intensifies the overall circulation within the sub-basin as well as in the straits. It is in the straits that the water exchange increases the most if the wind speed along their axes increases (Figs 7, 8). The current speed increases along the straight coastline parallel to the wind direction, and the counter-wind current speed increases somewhat in the middle of the sub-basin. For example, the SRS-scenario yields the highest current speed increments in the meridional Suur and Hari straits (Fig. 5d). Enhancement of the northward water flux occurs along meridional stretches between Riga and Pärnu, and between Riga and Kolka. W winds enhance flows in the Soela and Irbe Straits, the latter also producing curious meandering streams behind the Kolka peninsula (Fig. 5c). The absolute increase in current speeds is 1–3 cm s^{-1} near the straight coasts and 3–6 cm s^{-1} in the straits, which makes up roughly 15–20% of the corresponding mean current speeds.

4.4. Possible effect on coasts and the aquatic environment

Changes in the wind-driven hydrodynamic regime can affect the aquatic environment via numerous processes, including changes in water and matter exchange through the straits and intensification of horizontal and vertical mixing within the sub-basins (e.g. Blaas et al. 2001, Kowalewski & Ostrowski 2005). As shown earlier, both an increase in wind speed and a change in its direction alter the water exchange between the sub-basins (Fig. 7). Although every strait has its specific 'sensitive' direction (Fig. 7d), the water exchange in the whole Gulf of Riga – Väinameri system increases when the resultant wind direction turns either south- or northwards (Figs 5d, 7c, 8b). As the trophic status of these basins is higher than in the Baltic Proper, nutrient concentrations should also decrease somewhat and salinity should increase. On the other hand, when the winds turn more zonal, water exchange decreases even if the wind speed increases (Fig. 7c). Water exchange can still increase in some other locations, such as the Irbe Strait and Kihnu Strait (Fig. 7b). For bays like Pärnu Bay, the water circulation intensifies and the sea level rises when winds blow more intensively along the straits axis, i.e. S–SW (Fig. 7a).

Along straight coasts, changes in vertical velocities, i.e. up- and downwelling, ought to occur. In fresh wind conditions above sloping coasts, wind-driven Ekman transport is compensated by upwelling near the coast to the left of the flow and downwelling to the right of it. The zones affected reach 1–2 internal Rossby radii, which in the Baltic Sea is about 5 km (Fennel 1991). Changes in vertical fluxes in the Baltic Sea have been studied in greater detail using 3D modelling by Lehmann et al. (2002), Myrberg & Andrejev (2003), and Kowalewski & Ostrowski (2005). In this study, we do not quantify this effect for the Estonian coastal sea. However, following the ideas expressed in the above studies, an increase in W winds should increase downwelling events along the southern coasts of the Irbe Strait and Gulf of Riga, and strengthen upwelling events along the Saaremaa coasts (Fig. 5c). An increase in S winds should enhance upwelling along the coast between Kolka and Riga, and in the Suur Strait. Downwelling events will probably become more frequent mainly along the eastern sections of the Gulf of Riga and the Väinameri (Fig. 5d).

The possible influence on coastal processes is illustrated by a flow series chosen near the coasts of Salme, Virtsu and Pärnu (Figs 2e–h, 8a–c). The annual development of cumulative velocity components under different scenarios evidently depends on the direction of the coastline. Stronger longshore winds always increase current speeds. When considering corresponding time series of bottom stresses, it appears that a relatively small, up to 20% increase in mean current speed, may result in up to 2–3 times larger sums in bottom stresses (Fig. 8d–f). The changes are the most pronounced during the last months of the year (Fig. 8), whereas the relatively calm summer months are geomorphically inactive. (We would like to stress that we are not dealing here with coastal modelling; we are only outlining the rough proportions of bottom stresses here.)

In the coastal regions of Pärnu Bay, current-induced bottom stresses are predominantly directed to the bay's end (90% work against 10%, Fig. 8d) and the same applies to wave action. In addition to the influence of the asymmetry in the regional wind statistics (Soomere 2001), the local wind is considerably shielded by the land in the north. Furthermore, the attacking waves are always higher than the departing ones. The accumulation of sand in the region of the Pärnu beaches is a result of such asymmetry (Figs 6, 8) in wave and current action. The distribution of the modelled velocities in the Pärnu Bay region show a predominance of small velocities: only 0.9% of the velocity readings were > 45 cm s^{-1} in 1999. However, these 0.9% yield about 26% of the total annual bottom stress and 49% of the annual potential 'work', as bottom stress is proportional to the square of the velocity. Moreover, small velocities below a certain threshold value, e.g. 15 or 20 cm s^{-1}, yield 'wasted' stresses that are not able to erode, suspend or transport sediments at all.

As bottom stress is proportional to the square of the flow speed, and wave energy density is proportional to the square of the wave amplitude, the major share of annual work attacking the coast is concentrated in stormy periods. As SW and W storms are associated with sea level rise near Pärnu, the high flow velocities and strong wave action decisive in coastal erosion events and longshore sediment relocation act 1–3 m higher than the usual waterline (Fig. 7). As storms occur mainly during autumns and mild winters, the absence of protective ice cover in such winters usually coincides with strong westerly winds.

5. Conclusions

1. Semi-realistic simulations using a 2D hydrodynamic model have demonstrated that a relatively modest increase in wind speed could be responsible for a mean sea level increase of up to 2–5 cm within the study area. This was the outcome of adding 2 m s^{-1} wind speed components to realistic-year (1999) wind forcing data. Changes of a similar magnitude probably occurred already between 1950 and 1990, and several climate modelling studies anticipate further changes.
2. The local wind-driven sea level change component we have analysed applies within our semi-enclosed study area, with straits lying in the SW, W and NW directions. An additional analogous change in the Baltic mean sea level probably exists. According to the study by Meier et al. (2004), the latter component could be up to 3–4 cm in the central Baltic. Consequently, a total wind-induced average sea level rise of 7–10 cm could occur at locations like Pärnu and Matsalu. Increasing wind speed will lead to increasing sea level variability, especially in windward bays. However, both the mean sea level and standard deviations could marginally decrease in the eastern section of the Väinameri, where, because of the limited flow capacity of the Soela Strait, the regionally windward but locally leeward sea level features partially cancel each other out.
3. The increase in wind speed generally leads to enhanced water exchange through the straits, as well as to a strengthening of the basin-scale circulation. Although the flow in each strait has its own most effective driving wind direction, the water exchange in the whole Gulf of Riga-Väinameri system increases when the resultant wind direction turns either south- or northwards. Near coasts, the relatively small (up to 20%) wind speed increase we have considered presumably enhances up- and downwelling events, and produces an up to 2–3 fold increase in bottom stresses in cases where the direction of the particular coastline section coincides with that of the relevant wind speed change scenario.

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