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Storm surge effects on a back-barrier tidal flat of the Danish Wadden Sea

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Abstract The most severe storm passing Denmark in the 20th century occurred on 3–4 December 1999. During this event, waves, currents and suspended sediment concentrations were measured on a muddy tidal flat vis-à-vis the salt marsh of Skallingen in the inner part of the Hobo Dyb tidal area on the Danish west coast. Locally generated wave heights increased during the storm to $H_s = 0.38$ m, with peak spectral wave periods of up to 3.6 s and wave-induced bed shear stresses reaching a maximum of about 2 N m^{-2}. The highest current velocity measured over the tidal flat during the surge was 0.64 m s^{-1} ($\sim$1 N m^{-2}). During the surge peak, the wind-induced water circulation close to the bed was about 0.2 m s^{-1} which is equivalent to a bed shear stress of about 0.2 N m^{-2}. The concentration of suspended matter on the tidal flat increased from about 10 mg l^{-1} to about 200 mg l^{-1}, and the mobile layer on the mud flat was removed. The estimated deposition on the salt marsh amounted to 133 g m^{-2} (0.15 mm) which is only about 50% higher than that of a previously monitored gale event, and corresponds to less than 10% of the yearly deposition at the site.

It is suggested that deposition on the salt-marsh surface caused by storms or gales depends on the prevailing meteorological and hydrographical conditions as well as on the sequence of previous import and high-energy events. The results indicate that a single extreme storm can be of less importance for annual variations in salt-marsh deposition than more regularly occurring gales.

Introduction

Storm surges are a relatively common phenomenon along the Wadden Sea coast of northwest Europe. They form primarily during the autumn/winter seasons, resulting from deep eastward-migrating low-pressure systems (cyclones) developed at the Polar Front in the North Atlantic. Due to wind setup and low barometric pressure, the most powerful cyclones are capable of increasing water levels by several metres above the astronomical tide along the Danish west coast.

The storm surges can cause conspicuous geomorphological changes such as extreme beach and foredune erosion, and the formation of washover channels when they breach the foredune system. They also pose a hazard to human lives, livestock and infrastructure, published records of storm surges dating far back in time for the region (cf. Gram-Jensen 1991; Oost 1995).

In order to inundate the salt-marsh areas with turbid water, and thus contribute to salt-marsh growth, windy periods and high-water levels are required to achieve salt-marsh deposition. However, based on measurements in the main channels of the Grådyb tidal basin, storms and gales appear to generate hydrographical conditions causing an export of fine-grained material to the North Sea (Bartholdy and Anthony 1998; Bartholdy 2000). Thus, import situations with fine-grained material coming in from the North Sea must be separated in time from the salt-marsh depositional events. These are capable of maintaining, and even increasing the level of the salt-marsh surface relative to the rising sea level in the Grådyb tidal basin (e.g., Nielsen 1935; Jakobsen 1953; Bartholdy and Madsen 1985) as well as in numerous other estuarine environments (e.g., Pethick 1981; Allen and Rae 1988; Allen and Pye 1992; French et al. 1994; Kirchner and Ehlers 1998; Esselink et al. 1998). In order to improve our understanding of this complicated system, it is of primary interest to examine the dynamics on the tidal flat and the salt marsh during storms.

It is generally thought that storm surges have a significant sedimentological effect on barrier islands in general. However, direct measurements of storm dynamics in the transition zone between tidal flats and salt marshes are scarce. Moeller et al. (1996) examined the wave climate in such a region, and found that waves quickly lose a major part of their energy over a salt-

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marsh surface. French and Spencer (1993) showed for a back-barrier salt marsh in Norfolk (UK) that storm events accounted for a significant fraction of the long-term sedimentation. For the same site, Steers et al. (1979) reported that the marsh surface appeared remarkably unaffected by a severe storm surge in 1978. Letzsch and Frey (1980) found that autumn and winter storms resulted in erosion of the marsh surface in a back-barrier setting in Georgia (USA). Flemming and Bartholomä (1997) observed in an East Frisian back-barrier tidal basin that the system accommodated less fine-grained sediment in winter than in summer. In several studies dealing with salt-marsh deposition, attempts have been made to correlate layers in the salt marsh with seasonal change (e.g., Allen 1990) and known storm events (e.g., Bartholdy 1997; Kirchner and Ehlers 1998). However, few, if any, direct measurements exist to demonstrate the effects of waves and currents during extreme storm surges in back-barrier salt-marsh areas.

This paper reports on wave activity, hydrography and sediment transport in the back-barrier area of Skallingen in the Danish Wadden Sea during the extreme storm surge of 3–4 December 1999. The primary objective is to evaluate various impacts of the storm, including the sediment budget of this back-barrier area under such extreme conditions.

Study area

Skallingen is a barrier spit connected to the mainland south of Blåvands Huk (Fig. 1). At present, the coastline is retreating about 3 m year^{-1} (Aagaard et al. 1995). This has lead to the formation of washover channels clearly visible on the aerial photograph in Fig. 1. In the back-barrier area, typical salt-marsh depositional rates are between 1.5 and 4 mm year^{-1} (Bartholdy 1997). The back-barrier tidal flats are connected to the main inlet, Grådyb, through the tidal channel Hobo Dyb. The intertidal area is bounded to the east by the island Langli. In the northeast the tidal flats continue across a tidal divide toward the tidal channel Hjerting Løb which is situated east and north of Langli. The mean tidal range in the area is about 1.5 m but, as wind tide plays an important role, the active intertidal zone is perhaps closer to 2.5 m in height. The tidal flats behind Skallingen are primarily sandy, with muddy mussel banks along the margins of the channels. In the innermost part of the channel area, a relatively small turbidity maximum associated with mud flats occurs. In this area, direct observations of waves, currents, salinity, temperature and suspended material concentration were made coincidentally during the December storm of 1999.

Materials and methods

The data presented are part of a time series recorded over 35 days from 29 October to 5 December in 1999. The location of the instruments on the Skallingen tidal flat (site T) is indicated in Fig. 1. Waves and water levels were recorded by means of a wave recorder of the type Pacer Model 688WTG. Waves were recorded at a frequency of 2 Hz over 8.5 min every 1.5 h, while the mean water level was measured continuously every minute. Current velocity and direction, salinity, water temperature, and concentration of suspended sediment (C) were recorded every 10 min by means of an Aanderaa RCM 9 current meter placed adjacent to the wave recorder. This instrument was mounted in a steel tube on a concrete slab with its sensors located 0.5 m above the tidal flat. Current speed and direction were averaged over 600 acoustic doppler scans during each recording interval. Suspended sediment concentrations were estimated on the basis of back-scattered, infra-red light recordings (OBS) once every measuring cycle. The OBS records (NTU, calibrated by the manufacturer) were transformed into concentrations of suspended matter by an in-situ calibration [$C_{mg/l} = 2.289 \times NTU_{OBS}$; $r^2 = 0.95$] carried out during earlier measurements in the area.

Water levels were recorded at four locations in the area (Fig. 1). In a salt-marsh creek on the Skallingen peninsula (site S), the water level was recorded every 10 min by means of a ventilated pressure transducer maintained by the Institute of Geography, University of Copenhagen (Hasholt et al. 1998). This was also the case in the Varde Å estuary (site V) where the instrument is maintained by the County of Ribe. On the Skallingen tidal flat (site T), the measured (not ventilated) pressure was adjusted for the atmospheric pressure recorded at the salt-marsh station. The reference level on the tidal flat was adjusted to equal the level at the salt-marsh station during a relatively calm and long spring high-tide period. In the harbour of Esbjerg (site E), the water level was measured by means of a traditional floating device maintained by the Danish Coastal Authority. In all pressure transducer records, the measured (salt marsh and tidal flat) and estimated (Varde Å estuary) salinities and temperatures were used to further adjust the recorded water levels. Wind velocity (averaged over 10 min) was measured 10 m above the land surface at the Skallingen salt-marsh station (site S).

Results and discussion

The storm

With maximum gusts of 51.4 m s^{-1} (185 km h^{-1}), the storm of 3–4 December 1999 was the most severe cyclone passing Denmark in the 20th century (www.dmi.dk). It represented the culmination of an otherwise typical fall/winter situation in the North Sea, associated with relatively large temperature differences between the southern Atlantic air masses and the cold polar air. Following a period during which a

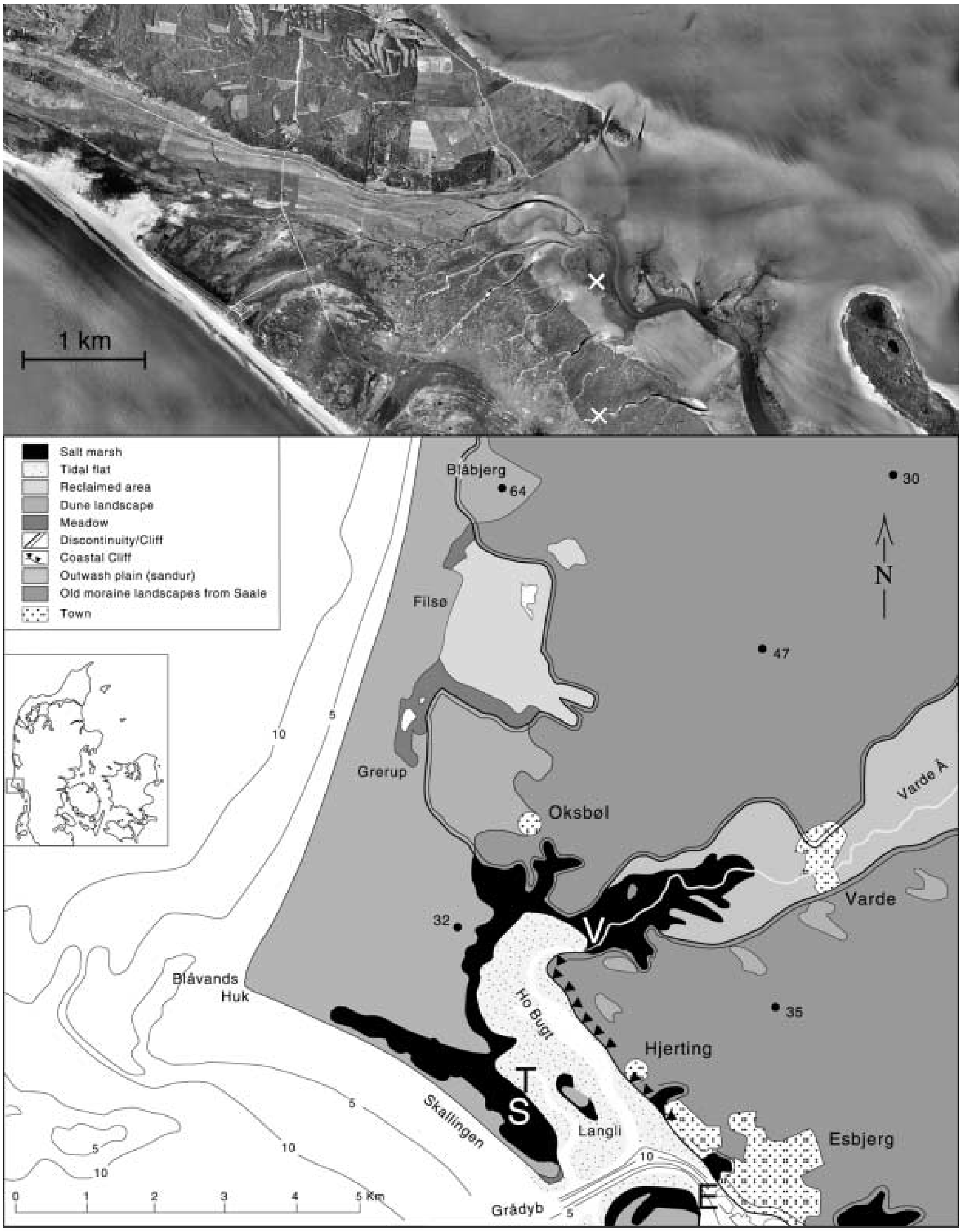


Fig. 1 Map and aerial photograph of the study site (geomorphological map modified after Smed 1978). *S*, *V*, *E* (map) Locations of tide gauges on the salt marshes of Skallingen, the Varde Å Estuary, and the harbour of Esbjerg, respectively. *T* (map) Study site on the Skallingen tidal flat. *White crosses* (aerial photograph) Locations of study site and tide gauge in the Skallingen area

number of less severe cyclones passed Denmark in late November, this particular storm moved from Scotland over the North Sea and then Denmark with its centre about 100 km north of Skallingen. Its intensity peaked over Skallingen in the evening of 3 December and, at 18:15, it was associated with mean wind speeds of 32.7 m s^{-1} (10-min average) and a low pressure of 959 hPa (Fig. 2). As a consequence of the storm centre being located north of Skallingen and the counter-clockwise rotating cyclones in the northern hemisphere, the wind direction changed from due south when the wind started to pick up, veering southwest and west to northwest during the passage of the cyclone.

The astronomical high tide occurred at noon and midnight, and the actual water level reached a maximum of 4.00 m DNN (Danish Ordnance Datum) at Esbjerg at 18:30. About 30 min earlier (18:00), the maximum water level on the back-barrier tidal flat of Skallingen (site T, Fig. 1) reached 3.37 m DNN. The water level on the salt marsh of Skallingen (site S, Fig. 1) reached a maximum of 3.24 m DNN at approximately 18:15. At the time of maximum water level in Esbjerg, water levels on the salt marsh and on the tidal flat were 3.24 and 3.30 m DNN, respectively. This illustrates the effect of wind setup, and stresses the fact that maximum storm water levels in the Wadden Sea are significantly lower along the barriers than along the mainland coast.

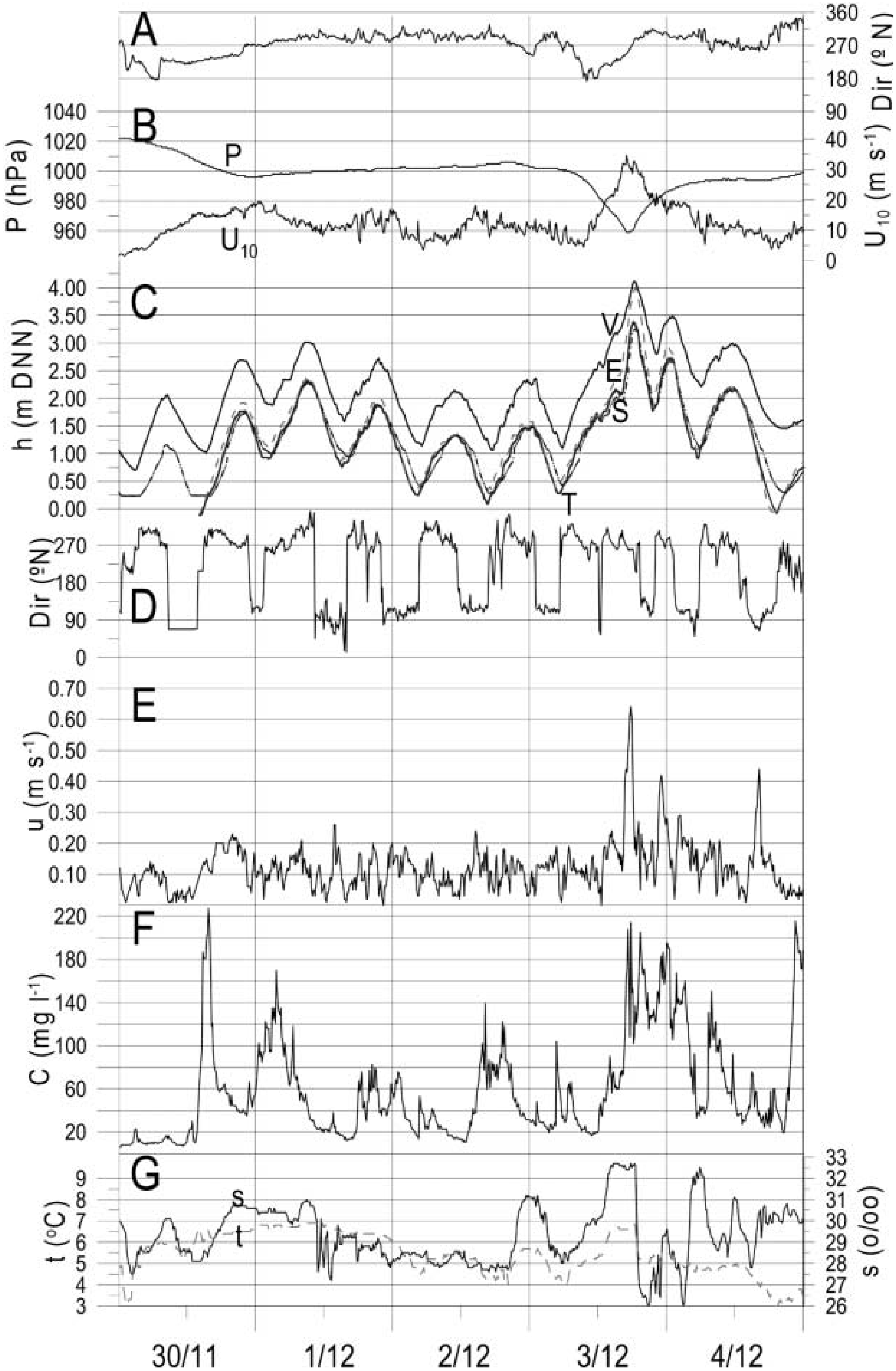


Fig. 2A–G Meterological and hydrographical parameters measured during the storm of 3 December 1999. **A** Wind direction, and **B** mean wind speed 10 m above the surface (U_{10}) and air pressure (P) at the salt marsh of Skallingen. **C** Water levels at Skallingen (*dash-dotted line S* Salt marsh; *full line T* Tidal flat), the Varde Å estuary (*full line V*), and Esbjerg (*dashed line E*; for locations see Fig. 1). **D** Current direction. **E** Current velocity. **F** Concentration of suspended sediment. **G** Salinity (s) and water temperature (t)

The maximum water level of 4.00 m DNN in Esbjerg has been exceeded only twice in the 20th century (4.40 and 4.13 m DNN in 1981 and 1990, respectively), and equalled once in 1928. Had the storm culmination coincided with the astronomical high tide, the water level would undoubtedly have been the highest ever recorded in the harbour of Esbjerg.

Dynamics on the back-barrier tidal flat

Waves

Waves are in general very small in this back-barrier environment. Recorded significant wave heights are usually less than 0.1 m with wave periods between 1.5

and 2 s. During the storm, however, wave heights increased to $H_s = 0.38$ m with peak spectral wave periods reaching 3.6 s (Fig. 3). The wave field was locally generated on the salt marsh over a fetch close to 3.5 km long with a water depth of approximately 2 m. Waves from the North Sea did not penetrate into the back-barrier area. If such waves had made their way to the back-barrier area, they would have had far longer periods than the ones recorded (Fig. 3). As the wind decreased and shifted to the northwest (Fig. 2), water levels and fetch were reduced, and the wave activity dropped considerably. During the high tide at midnight on December 3–4, waves were again small (~0.1 m) even though the marsh was still inundated and winds were blowing at 20 m s^{-1}.

The bed shear stress under waves can be determined by the relationship $\tau_{0,w} = 1/2\rho \cdot f_W \cdot u_m$ (e.g., Nielsen 1992) where f_w is a wave friction factor, and u_m is the maximum orbital velocity. The wave friction factor depends on the ratio between the hydraulic roughness and orbital diameter. Wave orbital velocities at the bed were in this case computed from measured wave heights using linear theory. During the storm, wave-induced bed shear stresses increased by about an order of magnitude above background levels. Using an estimated bed roughness for the mud flat of $k_s = 0.006$ m (Soulsby 1983), wave-induced bed shear stresses exceeded 2 N m^{-2} (Fig. 3). However, such stresses are by no means extraordinary,

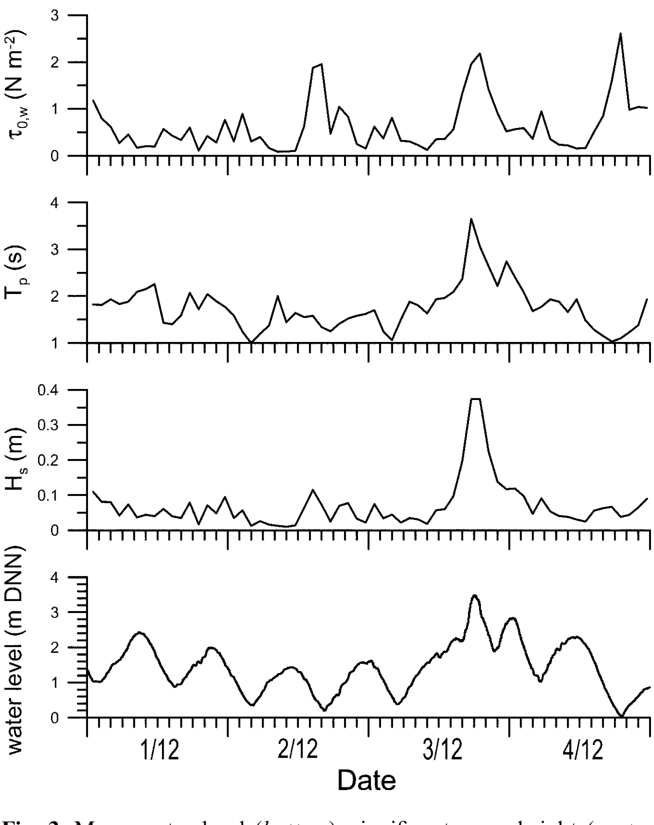


Fig. 3 Mean water level (*bottom*), significant wave height (*centre bottom*), peak spectral wave period (*centre top*), and computed wave-induced bed shear stress (*top*) on the Skallingen tidal flat

Table 1 Significant wave heights (H_s), spectral wave periods (T_p) and computed wave-induced bed shear stresses ($\tau_{0,w}$) during the storm

Time	H_s (m)	T_p (s)	$\tau_{0,w}$ (N m^{-2})
99-12-03			
13:00	0.060	1.96	0.36
14:30	0.097	2.09	0.56
16:00	0.197	2.37	1.33
17:30	0.373	3.65	1.96
19:00	0.374	3.06	2.18
20:30	0.224	2.64	1.42
22:00	0.138	2.22	0.91
23:30	0.117	2.74	0.52
99-12-04			
01:00	0.119	2.40	0.57
02:30	0.097	2.08	0.59
04:00	0.047	1.68	0.31

as they are often reached or exceeded during normal low-tide periods (Fig. 3). Table 1 lists the recorded significant wave heights and peak spectral wave periods as well as the computed wave-induced bed shear stresses ($\tau_{0,w}$) during the storm.

Currents

The data for key meteorological and hydrographical parameters for the storm are presented in Fig. 2. Water level in the Varde Å estuary (site V, Fig. 1) is generally high because of river flow. These data demonstrate that, during the peak of the surge, the water level in the estuary (max. 4.12 m DNN) was approximately equal to the water level at Esbjerg. This suggests that, along the mainland coast vis-à-vis the recording station on the Skallingen tidal flat, the water level can be regarded as largely equivalent to the water level in Esbjerg during the surge peak. Based on this assumption, the water level was more than 0.5 m higher along the mainland coast than on the tidal flat over a period of 4 h. The maximum difference over the 4-km-long distance was 0.78 m about 1 h after high tide. During wind setup, the surface water is blown downwind at a rate depending on the wind speed. Using the results presented by Allen (1984), the proportionality factor between the wind speed 10 m above the surface and the surface current is close to 0.04 at high wind speeds. This means that the easterly directed surface current recorded during the storm could have reached a maximum of about 1.3 m s^{-1}.

Table 2 includes an attempt to distinguish between the shear stress derived from the wind-induced circulation and that resulting from the tidal current flowing toward the salt marsh during the storm peak. During wind setup, the equation of motion in the wind direction (neglecting wave contributions) reads

$$\tau_a = -\tau_o - \rho \cdot g \cdot h \cdot [\delta h/\delta x] \quad (1)$$

where τ_o is the bottom shear stress, h is the water depth,

Table 2 Meteorological and hydrographical parameters for the Skallingen study site at the peak of the storm on 3 December 1999. Mean wind and current directions were 267° (coming) and 255° (going), respectively. *U_{10}* Mean wind speed 10 m above the salt-marsh surface. *h* Water level at the salt-marsh tide gauge. *D_t* Water depth at the current meter on the tidal flat. *$u_{0.46}$* Measured current velocity 0.46 m above the bed. *τ_a* Wind shear stress on the water surface. *τ_o* Current-induced bed shear stress. *τ_{oa}* and *τ_{ot}* Bed shear stresses induced by the wind-driven current and by the tidal current, respectively

Time	U_{10} (m s^{-1})	h (m DNN)	D_t (m)	$u_{0.46}$ (m s^{-1})	τ_a (N m^{-2})	τ_o (N m^{-2})	τ_{oa} (N m^{-2})	τ_{ot} (N m^{-2})
16:40	31.59	2.13	2.85	0.41	3.74	0.46	0.15	0.31
16:50	31.51	2.23	2.93	0.44	3.72	0.53	0.15	0.38
17:00	33.61	2.32	3.01	0.43	4.24	0.50	0.17	0.33
17:10	32.15	2.45	3.31	0.50	3.88	0.68	0.16	0.52
17:20	31.38	2.64	3.44	0.53	3.69	0.77	0.15	0.62
17:30	30.49	2.83	3.57	0.61	3.49	1.01	0.14	0.87
17:40	30.19	2.99	3.71	0.64	3.42	1.12	0.14	0.98
17:50	29.64	3.12	3.73	0.60	3.29	0.98	0.14	0.84
18:00	30.62	3.20	3.78	0.43	3.52	0.50	0.14	0.36
18:10	32.64	3.23	3.77	0.31	4.00	0.26	0.16	0.00
18:20	32.36	3.24	3.77	0.18	3.93	0.09	0.16	0.00
18:30	31.63	3.24	3.71	0.22	3.75	0.13	0.16	0.00

ρ is the water density, g is the acceleration due to gravity, and τ_a is the wind shear stress on the water surface calculated from

$$\tau_a = \rho_a \cdot C_a \cdot U_{10}^2 \qquad (2)$$

where ρ_a is the air density, C_a is an atmospheric drag coefficient which is approximately 0.003 for the measured wind speeds (Wu 1982), and U_{10} is the wind velocity at 10 m above the surface.

Near the bed, the local current velocity at level z above the bed (u_z) is expected to follow a logarithmic velocity distribution which is related to the bed shear stress τ_o by the relationship

$$u_z = [8.5 + 2.5\ \ln(z/k_s)]\sqrt{(\tau_o \cdot \rho^{-1})} \qquad (3)$$

where k_s is the equivalent bed roughness, in this case estimated at 0.006 m (Soulsby 1983).

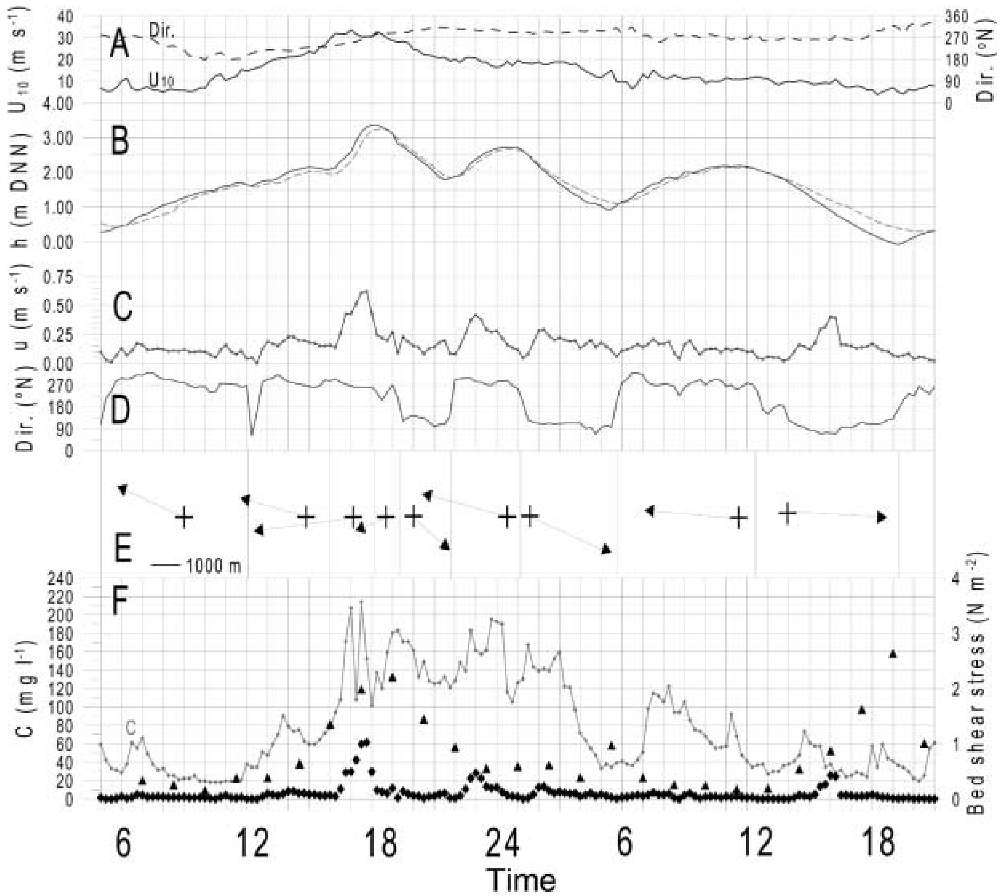
The bed shear stress consists of two components, $\tau_{0,a}$ and $\tau_{0,t}$. The former is the component derived from the wind-induced circulation, whereas the latter is derived from the tidal flow, in this case induced by water running toward the salt marsh.

Water levels on the salt marsh and on the tidal flat were nearly constant ($\delta h/\delta t \sim 0$) from 18:00 to 18:30 (Table 2). Therefore, at 18:10, 18:20 and 18:30, the local tidal current is expected to have been negligible, and Eq. (3) can be solved for $\tau_{0,a} = \tau_0$ with u_z as the measured current velocity at $z = 0.46$ m. According to Van Rijn (1994), the wind-induced bed shear stress can be expected to vary as a proportion of the wind shear stress at the water surface. This value (mean for the three data sets) was only 4% in the present case, corresponding to a mean wind speed of about 32 m s^{-1}, a mean water depth of about 3.8 m, and a current velocity of about 0.2 m s^{-1}.

The highest current velocity measured during the surge was 0.64 m s^{-1}. This represents the culmination of the flood current peak which lasted about half an hour, with current-induced bed shear stresses close to 1 N m^{-2}. After the culmination of the storm, the flood current induced by the astronomical tide gave rise to a velocity of about 0.42 m s^{-1} ($\tau_0 = 0.5\,\mathrm{N\,m^{-2}}$; Fig. 4), which was similar to the maximum value recorded during the third ebb period of the storm surge. During the smaller surge on 1 December (high-water level at Esbjerg: 2.39 m DNN), the highest measured current velocity on the tidal flat occurred during the ebb and was only 0.26 m s^{-1} ($\tau_o = 0.18\,\mathrm{N\,m^{-2}}$). Judging from the recorded time series, the maximum bed shear stress of 1 N m^{-2} therefore most likely represents the highest possible current-induced bed shear stress acting on the mud flat. Based on water-level records from Esbjerg Harbour (1889–1989), a high-water level varying between 2.25 and 3.0 m DNN can be expected about twice a year (twofold standard deviation). Consequently, the maximum current-induced bed shear stress during the major storm was about six times larger than that of a more frequently occurring gale like the one on December 1 and, in contrast to such an event, clearly flood-dominated.

Constituting only 4% of the wind shear stress on the water surface, the bed shear stress originating from the wind-induced return flow near the bed was insignificant compared to that of the tidal flow. The water displacement produced by the wind, however, was large enough to have significant hydrographic effects. This is demonstrated by a rapid shift in salinity only minutes after the current direction changed from west to southeast (Fig. 2). The salinity dropped from 32.6 to 28.4‰ over a 15-min period, and it continued to drop during the following 2 h to reach 26.0‰, this being the lowest value measured during the two weeks preceding the storm. Because of the high water level, and hence the large quantities of North Sea water in the region, freshwater from the river Varde Å (the only significant freshwater source in the area) is expected to have been pushed back up into the river valley. The existence of brackish water during this early stage

Fig. 4A–F Tidal, wind and wave dynamics at the study site in the back-barrier area of Skallingen during the storm of 3–4 December 1999 (see Fig. 1 for locations of tidal flat and salt marsh). **A** Wind velocity (*full line* U_{10}, 10 min mean) and direction (*dashed line Dir.*) 10 m above the salt-marsh surface. **B** Water levels on the tidal flat (*full line*) and in the salt marsh (*dashed line*). **C** Current velocity on the tidal flat 0.5 m above the bed. **D** Current direction on the tidal flat 0.5 m above the bed. **E** Integrated tidal displacement based on the measured current velocity on the tidal flat 0.5 m above the bed. Each vector indicates the current direction and integrated "length" in the time span bordered by the vertical lines. **F** Concentration of suspended matter on the tidal flat 0.5 m above the bed (*full line*), and bed shear stress induced by tidal currents (*diamonds*) and by waves (*triangles*)



of the ebb surge can therefore only be explained if freshwater derived from the river was mixed with seawater by the setup circulation against the direction of the wind.

During the storm, the direction of the tidal current on the Skallingen tidal flat roughly followed the slope of the water surface between the tidal flat and the salt marsh (Fig. 4). While the tide was rising on 3 December from 5:00 to about 1 h after high water at 19:20, the currents on the tidal flat were directed toward the salt marsh. Because of the complicated velocity distribution caused by the wind-induced circulation, it is meaningless to integrate velocity over depth. Instead, the actual measured water displacement (the time-integrated current velocity at 0.5 m above the bottom) is used as an indication of the near-bottom water movement. During the period in question, this displacement amounted to 8,426 m toward the west (Fig. 4), whereas in the following short ebb period the displacement of the return flow amounted to 1,323 m. The flood current prior to the astronomical high water at midnight moved 2,763 m, and the following ebb 2,754 m. During the final flood and ebb currents of the storm, the displacements amounted to 2,978 and 3,136 m, respectively. Adding these flood and ebb current displacements together results in 14,167 m in the flood direction, and 7,213 m in the ebb direction. Thus, in the near-bed region of the tidal flat, only about half of the water flowing toward the salt marsh returned. This imposes a near-bed sediment transport toward the salt marsh.

Sediment transport

During the 5-day period shown in Fig. 2 (30 November – 4 December 1999), the concentration of suspended sediment over the muddy tidal flat (at $z = 0.46$ m) varied from 10 to 200 mg l^{-1}. In general, the concentration was highest during low water when even small waves are capable of resuspending the mobile layer of the mud flat. This pattern was interrupted during the storm which generated persistently high concentrations of 100–200 mg l^{-1} (Fig. 4). When the wind started to pick up just before the astronomical high water at noon on 3 December, resuspension of the mobile layer began as the wave bed shear stress reached a level of about 0.3 N m^{-2} (Fig. 4). At this time, the current-induced bed shear stress was practically zero. The maximum concentration (214 mg l^{-1}) occurred just before high water when the combined maximum wave- and current-induced bed shear stress (2.0 and 1.0 N m^{-2}, respectively) reached 3.3 N m^{-2} (computed according to Soulsby 1997). During the following short ebb period, the values dropped to about 120 mg l^{-1}. In the flood period prior to the astronomical high tide at midnight, the concentration increased to about 200 mg l^{-1}, and it dropped to 35 mg l^{-1} during the following ebb. During the last two tidal periods of the storm surge, the turbidity was thus significantly lower during the ebb than during the flood.

On the basis of these observations, the storm surge can be divided into an early phase (from the time the wind speed began to increase up to the peak of the storm) during which the concentration of suspended matter increased from about 10 to 200 mg l^{-1}, and a late phase (from the peak of the storm until the back-barrier water level was back to normal). In this late phase, the concentration dropped as a result of the water exchange between the tidal flat and the salt marsh. Flood water flowing from the tidal flat to the salt marsh during this phase lost a substantial amount of suspended sediment to the salt marsh before returning during the following ebb. Furthermore, the mobile layer of the mud flat was removed as the increased bed shear stress during low water was evidently not able to markedly increase the concentration of suspended sediment during these last two low-water periods. Even in the last low-water period, when the wave-induced bed shear stress reached the highest level measured during the storm surge (2.6 N m^{-2}; Fig. 4), suspended sediment concentrations did not exceed about 60 mg l^{-1}.

The transport of fine-grained suspended sediment is directly related to sediment concentration and water displacement. However, since only about half of the water flowing toward the salt marsh in the near-bed region returned the same way, calculations of net sediment fluxes based on the measured velocity and suspended sediment concentration will be meaningless. Instead, the transport-weighted sediment concentrations and high-water depths on the salt marsh (Table 3) were used to evaluate the sediment accumulation on the salt marsh during the storm surge, the differences between the flood and ebb concentrations being best suited for this purpose. For comparison, Table 3 also contains the results for the gale on December 1 prior to the storm.

The combined results suggest that the deposition during the gale was 94 g m^{-2}. During the storm surge this value increased to 133 g m^{-2}. Deposition during the storm surge was thus only about 50% higher than during the gale. As this type of gale occurs relatively frequently in the region, this means that an episodic extreme storm, like the one reported here, is of lesser importance for annual variations in salt-marsh deposition than the more frequent gales.

The dry bulk density of the salt-marsh clay is approximately 0.9 g cm^{-3} (Bartholdy 1997), and the calculated storm-surge deposition therefore corresponds to a sediment thickness of about 0.15 mm. Since a mean deposition rate of about 2 mm year^{-1} can be expected in the northernmost salt marsh of the Skallingen peninsula (Bartholdy and Madsen 1985; Bartholdy 1997), these results suggest that the storm surge contributed less than 10% of the yearly deposition on the salt marsh at the time.

As in the case of the gale (cf. second low water of 1 December, Fig. 3), the mobile layer of the mud flat was removed by the storm down to a mud resistance of 3–4 N m^{-2}. Since the maximum bed shear stresses during the storm surge did not substantially exceed those occurring commonly during low water, it would appear that a major factor controlling the transfer of fine-grained sediment from the mud flat to the salt marsh is the availability of fine-grained sediment on the mud-flat surface during tides which reach the marshes. This antecedent condition thus appears to be essential for the depositional effect of a storm. As discussed by Bartholdy and Anthony (1998), and by Bartholdy (2000), the major import of fine-grained sediment from the North Sea to the tidal area seems to be episodic, and related primarily to summer conditions. In 1999, the December 3–4 storm occurred during the winter season after a series of gales, and it can therefore be regarded as having had a relatively small depositional potential. This was confirmed by a visit to the area two days after the storm. There were absolutely no signs of this high-energy event at the time, neither on the salt marsh nor on the mud flat. This contrasts sharply to a similar visit after the storm surge of November 1981, when a thin layer of newly deposited sandy mud on the salt-marsh surface was sampled and analysed (Bartholdy 1985, 1997). The only unusual feature in the back-barrier area after the storm was a mud layer draping ripples on an otherwise sandy tidal flat near the mouths of salt-marsh creeks.

Table 3 Measurements of suspended sediment concentrations on the muddy tidal flat adjacent to the Skallingen salt marsh (for location see Fig. 1). $C_{f,mean}$ and $C_{e,mean}$ Transport-weighted mean concentrations in the flood and ebb currents, respectively. D_{HW} Water depth on the salt marsh at high water. The deposition on the salt marsh was estimated as $(C_{f,mean} - C_{e,mean}) \times D_{HW}$. As the salt marsh surface is not expected to be eroded, negative deposition was considered as an artifact and was disregarded. This has only practical importance for the 99-12-03 figure. Here, during the peak of the storm, only 16% of the water returned in the ebb period

Date	$C_{f,mean}$ (mg l^{-1})	$C_{e,mean}$ (mg l^{-1})	D_{HW} (m)	Deposition (g m^{-2})
99-30-11	12	17	0.16	–
99-30-11	75	88	0.74	–
99-12-01	88	22	1.28	84.5
99-12-01	55	44	0.86	9.5
99-12-02	27	55	0.33	–
99-30-11 to 99-12-02				94
99-12-03	97	144	2.24	–
99-12-04	166	111	1.66	91.3
99-12-04	78	42	1.15	41.4
99-12-05	104	36	0.00	–
99-12-03 to 99-12-05				133

Conclusions

1. The Skallingen back-barrier environment was left surprisingly untouched by a severe storm passing Denmark on 3 December 1999. The maximum bed shear stresses on the back-barrier mud flat during the

storm did not substantially exceed those occurring commonly during low water.

2. During the storm surge the near-bed region of the Skallingen tidal flat was dominated by currents directed against the wind and toward the salt marsh.
3. The wind-induced water circulation during the storm had a substantial effect on the mixing of water masses in the tidal basin, but it was too small to have any marked effects on the near-bed dynamics over the back-barrier mud flat.
4. The depositional effects of storms depend on the season as well as the sequence of previous import and high-energy events. One extreme episodic storm can be of lesser importance for annual variations in salt-marsh deposition than more frequent gales.

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References

Aagaard T, Nielsen N, Nielsen J (1995) Skallingen – Origin and evolution of a barrier spit. Meddel Skalling Lab B, XXXV

Allen JRL (1984) Sedimentary structures, their character and physical basis, vol II. Developments in sedimentology 30. Elsevier, Amsterdam

Allen JRL (1990) Salt-marsh growth and stratification: A numerical model with special reference to the Severn Estuary, southwest Britain. Mar Geol 95: 77–96

Allen JRL, Pye K (1992) Coastal saltmarshes: their nature and importance. In: Allen JRL, Pye K (eds) Saltmarshes: morphodynamics, conservation and engineering significance. Cambridge University Press, Cambridge, pp 1–18

Allen JRL, Rae JE (1988) Vertical salt-marsh accretion since the Roman period in the Severn Estuary, southwest Britain. Mar Geol 83: 225–235

Bartholdy J (1985) Dynamic interpretation of grain-size relations in silt. Geo-Mar Lett 5: 67–70

Bartholdy J (1997) The backbarrier sediments of the Skallingen Peninsula, Denmark. Dan J Geogr 97: 11–32

Bartholdy J (2000) Processes controlling import of fine-grained sediment to tidal areas: a simulation model. In: Pye K, Allen JRL (eds) Coastal and estuarine environments: sedimentology, geomorphology and geoarchaeology. Geol Soc Lond Spec Publ 175: 13–29

Bartholdy J, Anthony D (1998) Tidal dynamics and seasonal dependant import and export of fine-grained sediment through a backbarrier tidal channel of the Danish Wadden Sea. In: Alexander C, Davis RA, Henry VJ (eds) Tidalites: processes and products. SEPM Spec Publ 61: 43–52

Bartholdy J, Madsen PP (1985) Accumulation of fine-grained material in a Danish tidal area. Mar Geol 67: 121–137

Esselink P, Dijkema KS, Reents S, Hageman G (1998) Vertical accretion and profile changes in abandoned man-made tidal marshes in the Dollard Estuary, the Netherlands. J Coast Res 14(2): 570–582

Flemming BW, Bartholomä A (1997) Response of the Wadden Sea to a rising sea level: a predictive empirical model. Ger J Hydrogr 49: 343–353

French JR, Spencer T (1993) Dynamics of sedimentation in a tide-dominated backbarrier salt marsh, Norfolk, UK. Mar Geol 110: 315–331

French PW, Allen JRL, Appleby PG (1994) 210-Lead dating of modern period saltmarsh deposit from the Severn Estuary (southwest Britain), and its implications. Mar Geol 118: 327–334

Gram-Jensen I (1991) Stormfloder. Dan Meteorol Inst Sci Rep 91(1)

Hasholt B, Thomas UP, Simonsen K (1998) A new climate and water-balance station on the peninsula Skallingen, South West Jutland. Dan J Geogr 98: 84–87

Jakobsen B (1953) Landskabsudviklingen i Skallingmarsken. Geogr Tidsskr 52: 147–158

Kirchner G, Ehlers H (1998) Sediment geochronology in changing coastal environments: potentials and limitations of the ^{137}Cs and ^{210}Pb methods. J Coast Res 14(2): 483–492

Letzsch SW, Frey RW (1980) Deposition and erosion in a Holocene salt marsh, Sapelo Island, Georgia. J Sediment Petrol 50(2): 529–542

Moeller I, Spencer T, French JR (1996) Wind wave attenuation over saltmarsh surfaces: Preliminary results from Norfolk, England. J Coast Res 12(4): 1009–1016

Nielsen N (1935) Eine methode zur exacten sedimentationsmessung. Meddel Skalling Lab I

Nielsen P (1992) Coastal bottom boundary layers and sediment transport. World Scientific, Singapore

Oost AP (1995) Dynamics and sedimentary development of the Dutch Wadden Sea with emphasis on the Frisian Inlet. Geol Ultra Utrecht 126

Pethick JS (1981) Long-term accretion rates on tidal salt marshes. J Sediment Petrol 51: 571–577

Smed P (1978) Landskabskort over Danmark. Geografforlaget, Brenderup

Soulsby RL (1983) The bottom boundary layer of shelf seas. In: Johns B (ed) Physical oceanography of coastal and shelf seas. Elsevier, Amsterdam, pp 189–266

Soulsby RL (1997) Dynamics of marine sands. Thomas Telford, London

Steers JA, Stoddart DR, Bayliss-Smith TP, Spencer T, Durbridge PM (1979) The storm surge of 11 January 1978 on the east coast of England. Geogr J 145: 192–205

Van Rijn LC (1994) Principles of fluid flow and surface waves in rivers, estuaries, seas and oceans, 2nd edn. Aqua, Amsterdam

Wu J (1982) Wind stress coefficients over sea surface from breeze to hurricane. J Geophys Res 87: 9704