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Modelling waves and surges during the 1953 storm

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Waves and sea levels have been modelled for the storm of 31 January–1 February 1953. Problems in modelling this event are associated with the difficulty of reconstructing wind fields and validating the model results with the limited data available from 50 years ago. The reconstruction of appropriate wind fields for surge and wave models is examined. The surges and waves are reproduced reasonably well on the basis of tide-gauge observations and the sparse observational information on wave heights. The maximum surge coincided closely in time with tidal high water, producing very high water levels along the coasts of the southern North Sea. The statistics of the 1953 event and the likelihood of recurrence are also discussed. Both surge and wave components were estimated to be approximately 1 in 50 year events. The maximum water level also occurred when the offshore waves were close to their maximum. The estimation of return period for the total water level is more problematic and is dependent on location. A scenario with the 1953 storm occurring in 2075, accounting for the effects of sea level rise and land movements, is also constructed, suggesting that sea level relative to the land could be 0.4–0.5 m higher than in 1953 in the southern North Sea, assuming a rise in mean sea level of 0.4 m.

Keywords: waves; surges; 1953 storm; modelling; statistics

1. Introduction

The storm surge of 31 January–1 February 1953 was one of the most devastating natural disasters in western Europe last century, with the loss of over 1800 lives in the Netherlands (Gerritsen 2005) and 300 deaths in southeast England (Baxter 2005). Storm surges (water levels raised by meteorological forcing) occur each winter when low pressure systems and gale-force winds cross northwest Europe. Generally, storm surges are accompanied by high waves which can damage or breach coastal defences. If the waves are combined with higher than usual sea level, the wave energy impacts near to the top of sea defences and can overtop them, attacking their rear face and increasing the likelihood of failure. In the North Sea, the most dangerous winds for generating high waves and sea levels are from the north to northwest. North to northwest gales often occur as depressions move across the north of the UK from west to east. If the peak of the storm surge

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One contribution of 14 to a Theme 'The Big Flood: North Sea storm surge'.

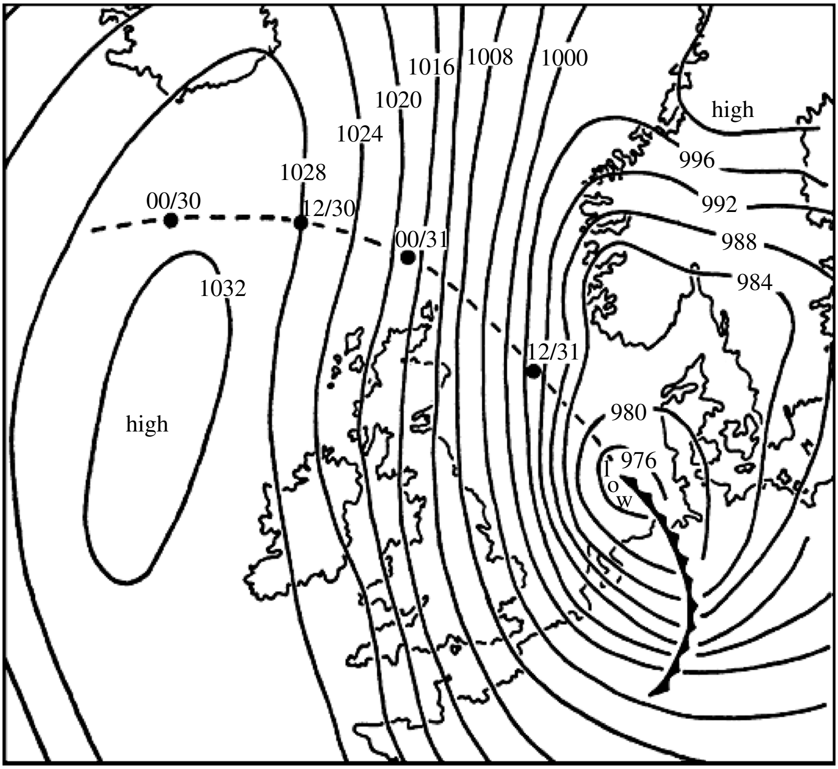

Figure 1. Meteorological chart for 00.00 1 February 1953 with the track of storm centre throughout the storm from 00.00 30 January to 00.00 1 February (replotted from Heaps 1967).

occurs near high tide, as in the 1953 event, this can produce particularly devastating conditions.

Since 1953, sea walls have been repaired and raised and other types of sea defences introduced. The Delta Plan in the Netherlands was established and led to construction of barriers across several of its estuaries over the next decades. A storm-surge barrage on the River Thames to protect London was agreed and officially opened in May 1984. Smaller offshore shore-parallel breakwaters at Sea Palling on the Norfolk coast and 'soft' defences such as beach nourishment and the reintroduction of saltmarshes are some alternatives (Huntington *et al.* 2003; Thomalla & Vincent 2003). In order to design flood defences for the future, it is important to know whether, in the light of current knowledge about climate change, future storm events could generate more severe conditions.

Here, we examine the severity of waves and storm surge in the 1953 event by means of extreme value statistics and model studies. The characteristics of the 1953 event are examined using statistics derived from observations and long-term climatic and hindcast model studies. Two EU projects, WASA (WASA group 1998) and STOWASUS-2100 (http://web.dmi.dk/pub/STOWASUS-2100) have produced multi-year time-series of wave model and tide-surge model data over the northeast Atlantic. WASA included present day wave and surge hindcasts for 1955–1994 and predictions for a twice-CO_2 scenario. STOWASUS used meteorological data from the ECHAM4 global climate

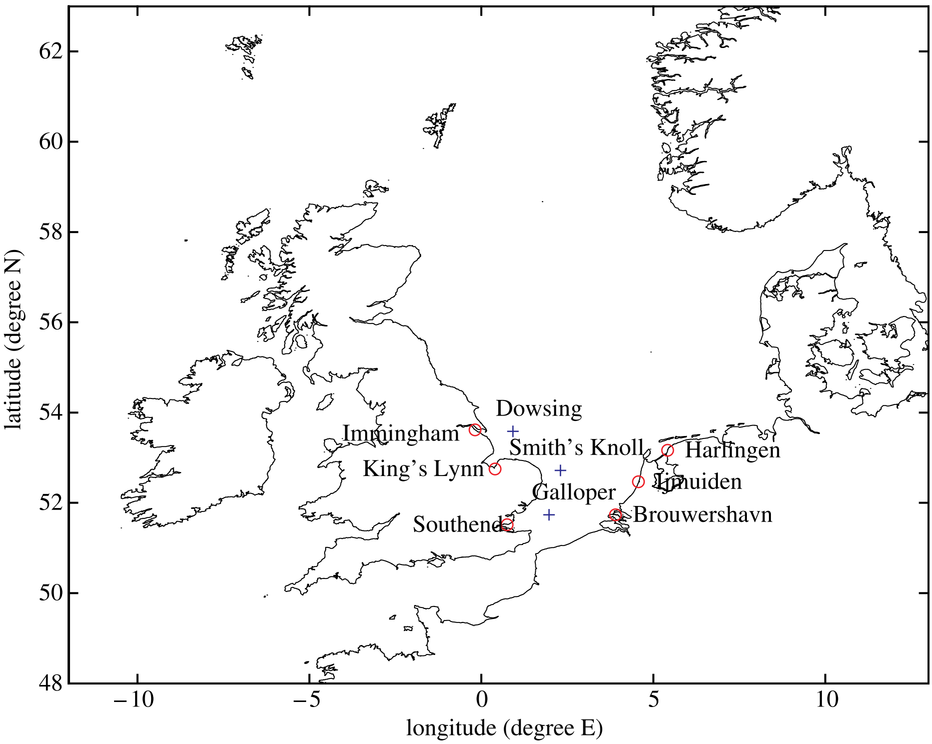

Figure 2. Locations of stations: blue crosses are light vessels, red circles are coastal tide gauges.

model for 'control' and '$2\times CO_2$' scenarios to investigate possible changes in the storm surge and wave climate. The JERICHO project examined trends in offshore wave climate from satellite and buoy data and used the SWAN wave model to transform offshore wave climate to the coast (Hargreaves *et al.* 2002).

A careful study was made of the 1953 storm and surge event by Flather (1984), with a model simulation on a 36 km grid. The tide and surge for the 1953 storm event were modelled for the present work using a 12 km-resolution shelf-scale model. The waves were modelled by using the same 12 km model nested in a 1 degree northeast Atlantic model. An important component of the modelling is reconstruction of realistic wind fields for the event. A novel component of this paper is the comparison of two sets of wind fields from different sources and the inclusion of a wave forecast which has not been reported previously.

Global and local increases in sea level owing to isostatic readjustment and global warming may produce an increased frequency of damaging conditions. We therefore examine the likelihood of such an event recurring, or being exceeded, in the light of estimates of climate change.

2. The 1953 storm surge and waves

The 1953 storm generated a large storm surge, which combined with a high spring tide to produce particularly high sea levels. The track of the storm (figure 1) brought exceptionally strong northerly winds over shallow areas of the western and southern North Sea where the main surge effect was generated

(Flather 1984). The elongation of the storm to the north also produced a long fetch and generated large wind waves. Water levels during the event were recorded at tide gauges in the North Sea and were estimated visually at other locations. Several tide gauges on the east coast of England failed during the storm. Rossiter (1954) obtained all available tide-gauge data and listed them as hourly time-series, providing a valuable record of the observed sea levels during the event. Rossiter (1954) also reproduced meteorological charts using information obtained from the UK Meteorological Office (the Met Office), and using the surge data, attempted to reconstruct the surge distribution over the North Sea and its evolution with time, aiming to understand the mechanisms of the surge.

In comparison with the tide-gauge records, evidence for the wave heights during the 1953 storm is sparse and often subjective and contradictory. The 1953 storm surge and its impact on the north Norfolk coast is discussed by Jensen (1953), Steers (1953), Steers *et al.* (1978) and Pye (1992). Only very approximate estimates for wave height exist, e.g. an estimated 4–6 ft above still water level, which may be equivalent to 2.5–3.5 m significant wave height, at the coast. Other estimates of wave height are given by Smith (1954), who states that '12 foot waves were reported on the first high tide (night of 31 January–1 February) and 9 foot waves on the second high tide (evening of 1 February) in the Low Countries'. These waves are presumably quite nearshore so that offshore waves could be much higher than this. Also, it is not clear whether the height referred to is trough-to-crest or from the undisturbed water level. Winds were reported as up to 114 m.p.h. at times. At the Smith's Knoll light vessel, the seas were reported to be at least 20 ft above mean sea level (MSL; Lawford 1954), which could imply a significant wave height of over 12 m. An approximate value given on the Met Office website (http://www.metoffice.com/education/historic/flood.html) is of waves exceeding 8 m in the North Sea.

Several light vessels (Dowsing: 53°35′ N, 0°55′ E; Smith's Knoll: 52°43′ N, 2°18′ E; Galloper: 51°44′ N, 1°58′ E, see figure 2) recorded winds and waves through the period. Maximum wave heights reported at these stations were 3, 2.5 and 4.5 m, respectively. It is difficult to reconcile these data with the known severity of the storm. An ambiguity in the reporting scale means that, potentially, these waves could have been much higher (see §5). Draper (1991) reports waves heights reaching over 6 m from a shipborne wave recorder (SWR) at Dowsing LV between 1970 and 1985 and over 5 m for SWR at Galloper, 1970–1971. Wind speeds of over 30 m s^{-1} were reported in Aberdeenshire on 31 January 1953 and close to the Dutch coast during the night of 31 January–1 February (Flather 1984). A fully developed sea in 30 m water depth for 30 m s^{-1} winds would have $H_S = 7.8$ m and peak period, $T_P = 14$ s (Hurdle & Stive 1989). An offshore significant wave height of 7.8 m may thus be regarded as a conservative estimate of the 1953 conditions just off the Norfolk coast.

3. Reconstructing wind fields for modelling

For use in numerical model simulations of the 1953 surge, Flather (1984) reconstructed the essential meteorological data from a combination of surface pressure observations from the Met Office's daily weather report and values read from charts. Winds were then estimated from the pressure distributions using empirical relations used at that time, to give meteorological fields at 3 h intervals. The lack of observations over the North Sea to define the evolution and movement of the depression, especially during the critical period on 31 January, was a problem, and limited the accuracy of the surge simulations. This contrasts with the large amount of data now available from offshore platforms and buoys in the North Sea and the meteorological forcing from atmospheric models with resolution typically O (10 km). The same data and methods were used to reconstruct wind and pressure fields on a 12 km grid covering the North Sea and continental shelf for the present study. This model grid is used for the POL tide-surge model and is referred to as CS3.

National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis winds for 1953 are available from the dataset from 1948 to the present (http://www.cdc.noaa.gov/cdc/reanalysis/reanalysis.shtml). Winds were produced with data assimilation from a global spectral atmospheric model (Sela 1982) on a latitude/longitude grid at approximately 2.5° resolution over the northeast Atlantic. These winds were used to force the northeast Atlantic 1 degree wave model to provide wave-boundary conditions for the continental shelf model. Two sets of wind data are provided: (i) regular gridded data at the model level $\sigma=0.995$ (where σ is the vertical coordinate used in the atmospheric model, representing the fraction of the total surface pressure, i.e. $\sigma=1$ at the Earth's surface and $\sigma=0$ at the outer limit of the model atmosphere) and (ii) winds at 10 m above the surface on a Gaussian grid (in which the grid size varies with latitude to preserve equal areas). The height above the surface corresponding to $\sigma=0.995$ will vary, but typically may be over 30 m. The ratio of winds at 10 m to winds at $\sigma=0.995$ is close to 0.8 over most of the sea areas, lower over land and higher near centres of depression (where there may be a difference in location owing to interpolation). A further discussion of the scaling of the wind in the atmospheric boundary layer is given in Appendix A.

Observed winds from three light vessels are compared with the model winds in figure 3. The nearest grid point to the location of the light vessel is taken in each case. The winds labelled CS3 are those from Flather's reconstruction. They may be seen to be consistently greater than the NCEP winds, but in quite good agreement with the observed winds. Both models overestimate the winds on 30 January. Flather's model picks up the secondary peak in wind speed on 1–2 February. The main peak in wind speed may be overestimated in the central-southern North Sea at Smith's Knoll.

The CS3 winds are in reasonable agreement with the NCEP reanalysis winds at $\sigma=0.995$, with allowance for differences in spatial and temporal resolution. Best agreement is in the northern and central North Sea. Off northwest Scotland there is a discrepancy in the N-component and in the southern North Sea in the E-component. The magnitude of the wind speed appears very similar. The coarser spatial grid in the NCEP data cannot capture the details of the winds in the southern North Sea. Note that the NCEP winds have been modelled dynamically rather than derived directly from the pressure maps so are much smoother but due to poorer resolution cannot capture smaller features. In the NCEP data, the time of maximum winds progresses smoothly as the storm moves down the North Sea, whereas in the CS3 winds, the time of maximum wind is

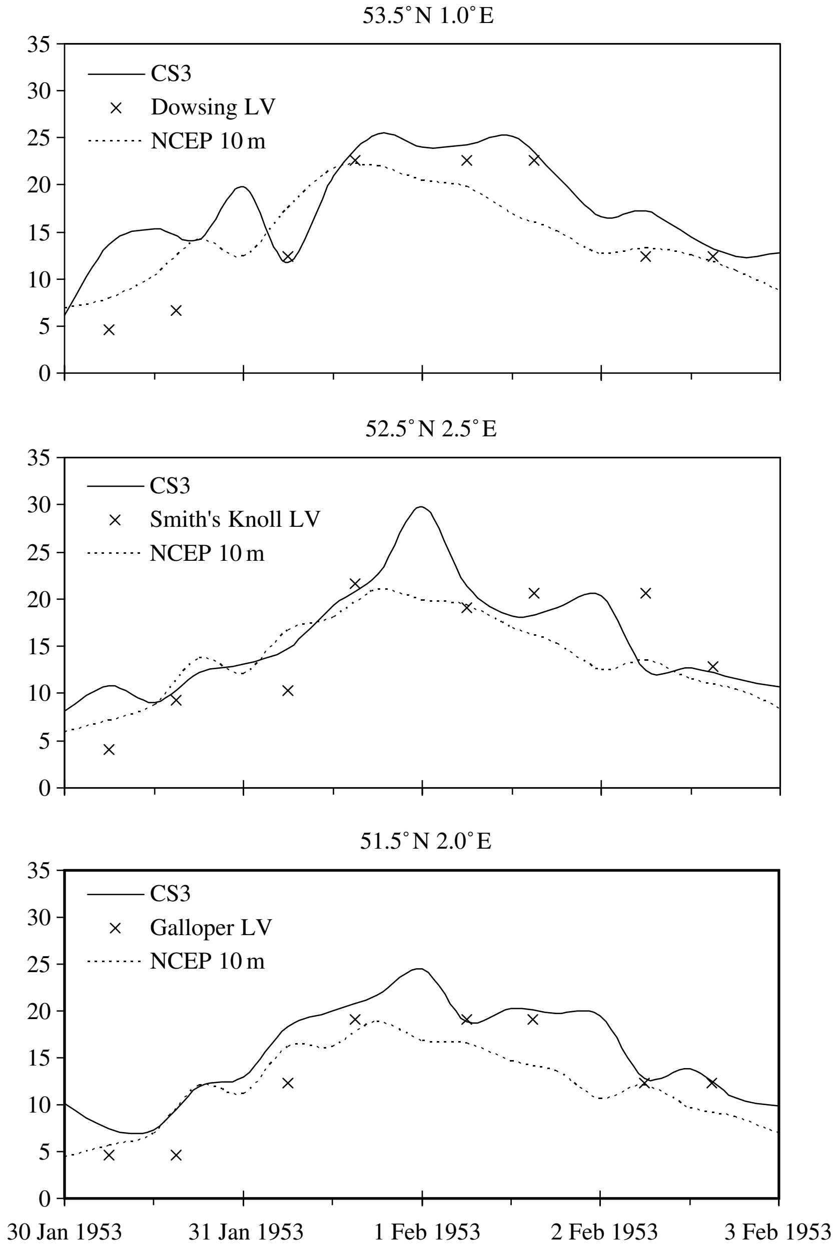

Figure 3. Comparison of model and observed wind speed in m s^{-1}. CS3 and NCEP 10 m are model winds from Flather's reconstruction and the NCEP reanalysis, respectively. Dowsing LV, Smith's Knoll LV and Galloper LV are observations at the named light vessels (see figure 2 for locations).

earlier in the southern North Sea. It is difficult to differentiate between these reconstructions due to the lack of data in the North Sea, for both wind and waves. The light vessels at Dowsing, Smith's Knoll and Galloper all showed a maximum of wind speed at 15.00 on 31 January. Nothing was reported between

06.00 and 15.00 so the maximum could have occurred even earlier. This supports Flather's reconstruction based on the barometric charts, therefore, there may well have been a local intensification of the winds.

4. Modelling the 1953 storm surge

Flather (1984) used numerical tide-surge models with his reconstructed meteorological data to simulate the surge on a 36 km grid. Despite the uncertainties of the forcing, the models reproduced the observed surges reasonably well, but some errors were noted. In particular, the first surge peak on the east coast of England late on 31 January was not reproduced, most probably because of deficiencies in the wind data. The analysis suggested that the surge was exceptional as a result of the exceptionally strong northerly gales west of the storm centre, the elongation of the storm in the north–south direction, the track of the depression along the axis of the North Sea bringing the gales to bear on the shallow waters in the west and south, and the slow speed with which the storm moved away, increasing the duration of the northerly gales.

New simulations of the 1953 event have been carried out using the CS3 tide-surge model developed at POL and run operationally since 1991 at the Met Office to predict storm surges. The predictions are used for flood warning on the coasts of England and Wales. Similar systems are run in all European countries bordering the North Sea (Flather 2000). Time-series comparisons have been carried out (figure 4) and show similar results to those presented in Flather (1984). Note that the surge peak at about 00.00 GMT, 1 February at King's Lynn, Southend and Brouwershaven is not reproduced by the model, but results at Ijmuiden are better. The largest observed surge (3.90 m) occurred at Harlingen, with corresponding model maximum 3.65 m.

The maximum computed surge elevation was extracted and its distribution plotted (figure 5). This shows maximum surges exceeding 225 cm over most of the Southern Bight of the North Sea, with values of 300 cm and more on parts of the Dutch coast. The maximum surge height is underestimated in the Thames estuary and in the Wash, where the model cannot resolve the approach channel to King's Lynn.

Figure 6 shows the computed difference in maximum sea level caused by the surge, defined as the difference between maximum observed water level and the maximum predicted tide. This is sometimes referred to as the 'skew surge'. Comparing this with figure 5, the maximum surge and skew surge are almost the same along the Dutch coast. This indicates that the maximum surge occurred very close to the time of high water in this area, resulting in the very high sea levels that produced the flooding.

5. Modelling waves during the 1953 storm

The wind data were used to drive a wave model of the continental shelf on the same 12 km grid as the surge model, i.e. 1/6° longitude by 1/9° latitude. Open sea boundary conditions for the shelf model were calculated from a northeast Atlantic model on a 1 degree grid (Wolf *et al.* 2002) extended to 70° N, driven by 10 m surface winds from the NCEP reanalysis data. The wave model used is

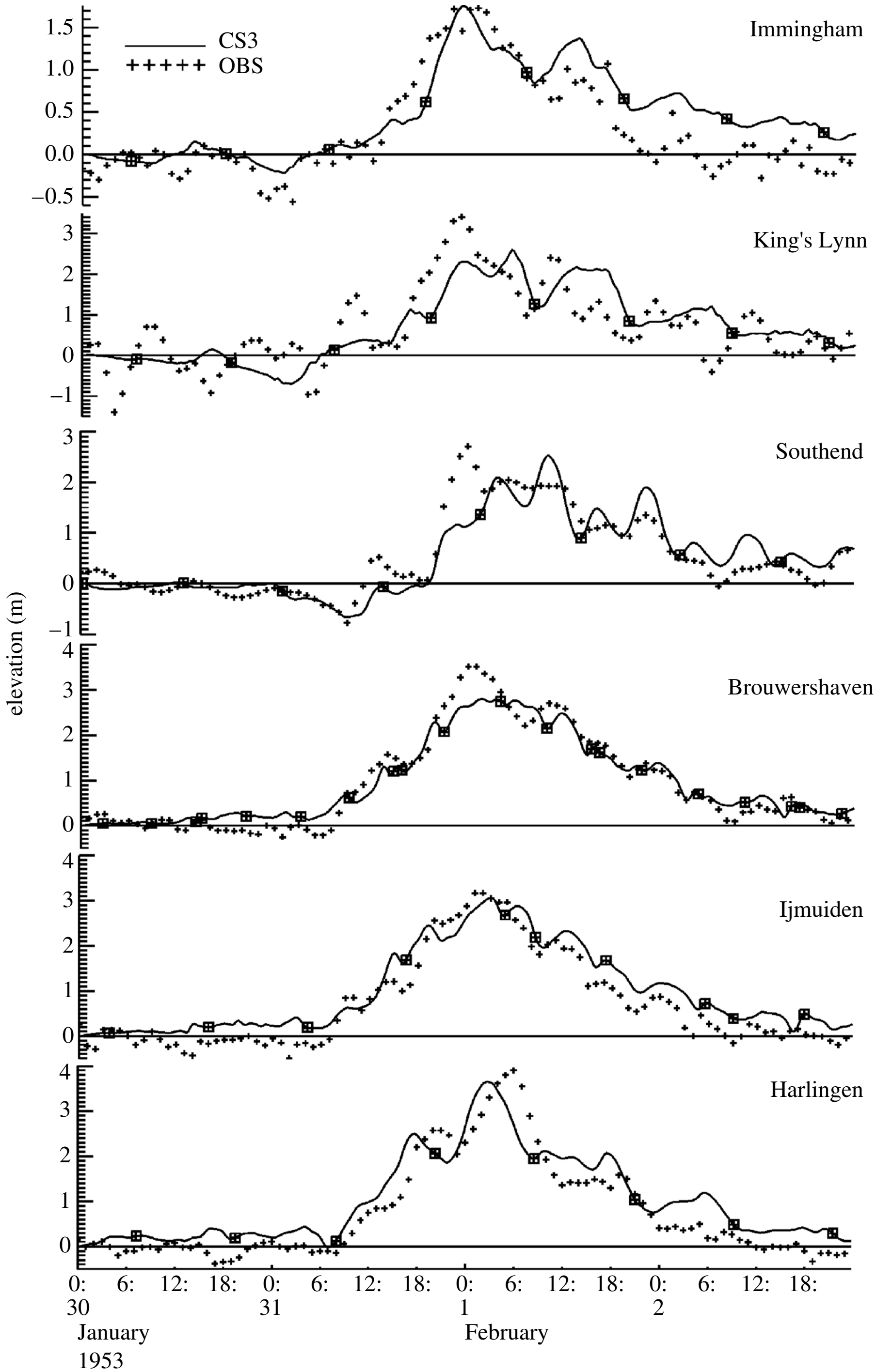

Figure 4. Comparison of computed (CS3) and observed (OBS) surge elevations at some UK and Dutch tide gauges. (Note scales vary.) The square symbol indicates approximate times of tidal high water in the model solution.

the version of the WAM model developed during the PROMISE project (Monbaliu *et al.* 2000).

Using the same (CS3) winds as for the surge model produced an overestimate of the wave height, and since these winds were probably not applicable to 10 m above the sea surface, a constant correction factor was applied to the wind speed to allow for the logarithmic relationship of wind speed with height (see Appendix A). Values of 0.72, 0.78, 0.80, 0.83 and 0.91 were tested. The optimum result was obtained for a scaling factor of 0.91, which brings the CS3 winds into closer agreement with the 10 m NCEP winds and the maximum wave height at the point nearest WASA/Cromer to 7.9 m. As already mentioned in §3, there are differences in the two wind fields. The NCEP reanalysis includes all available data but has a coarse spatial resolution. Flather's CS3 analysis is based on the barometric charts and may not be dynamically consistent, but can include more details of the local intensification of the wind; however, there are gaps in the information so that there are uncertainties in the reconstruction. A possible explanation for the different response of the wave and surge models is that the zone of higher winds needs to be nearer to the coast, i.e. in shallower water, since the wind-stress which drives the surge is more effective in shallow water unlike for waves where a larger wind-stress in deeper water generates larger waves. Reconstruction of the correct wind fields is an ongoing problem for hydrodynamic modelling.

The computed distribution of significant wave height at 00.00 GMT on 1 February 1953 is shown in figure 7. At this time, the wave height in the southern North Sea reached its computed maximum of about 10 m. Maximum recorded water levels along the Dutch coast occurred 3–7 h later, (Rossiter 1954) therefore coinciding with large offshore waves.

Time-series of wave height at the locations of the light vessels in the North Sea are shown in figure 8. It seems probable that the maximum observed wave heights at Dowsing and Smith's Knoll were underestimated, possibly owing to the ambiguity introduced by the method of recording visual observations, whereby waves with a significant wave height above 4.5 m have the same code as waves with height 5 m lower. The higher wave height should be flagged by adding a constant value to the wave direction. This may have been difficult to do under stress, and certainly the maximum recorded wave height of 2.5 m at Smith's Knoll does not agree with reports of wave crests over 20 ft high (Lawford 1954) or the statistics in Draper (1991). Both the lower heights, as recorded, and possible 'corrected' wave heights, assuming the alternative reading for the wave height code, are plotted. It seems plausible that at least the heights on 1 February were misreported. The model results for the maximum wave height at Dowsing seem in close agreement with the corrected values.

6. Surge, sea level and wave statistics

Various studies have been carried out to look at the frequency of occurrence of such extreme sea levels. To derive these directly from data requires a long time-series (at least 50–100 years) of annual maxima. Methods have been developed to use more of the data than just the block maxima, e.g. taking all exceedances over a threshold and the r-largest order statistics within a block (Coles 2001).

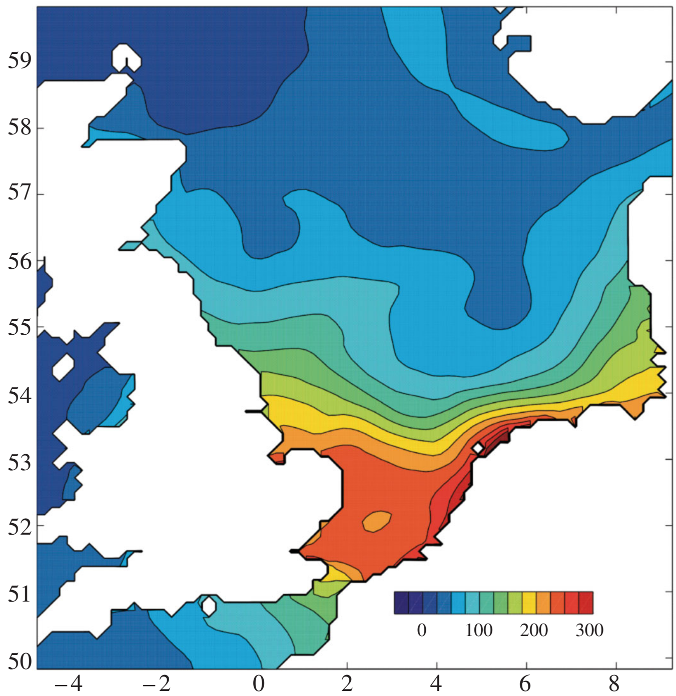

Figure 5. Maximum surge (cm) during the 1953 storm computed using the POL CS3 model.

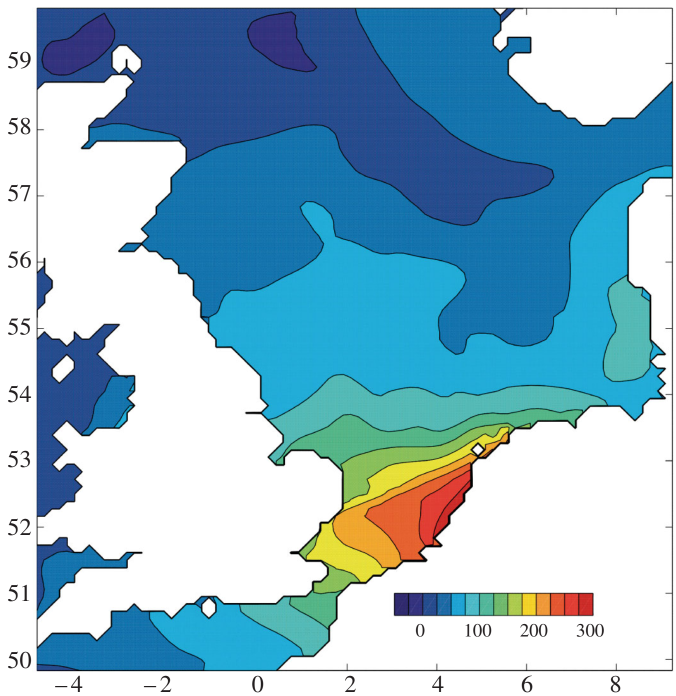

Figure 6. The difference between maximum water level and maximum predicted tide height during the 1953 storm computed using the POL CS3 model.

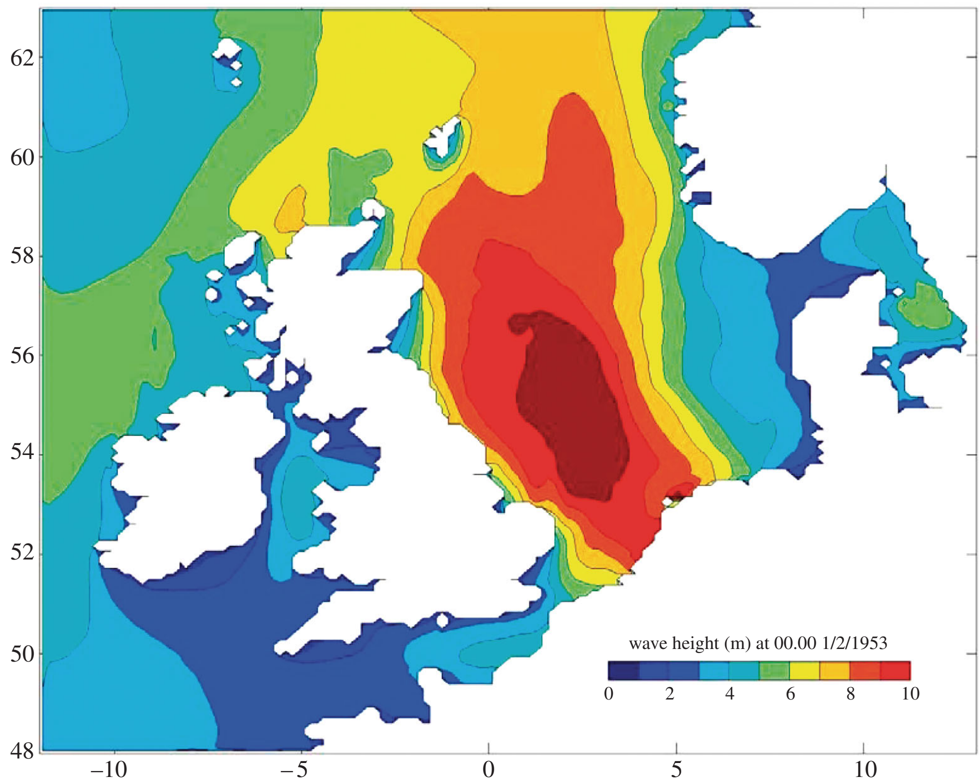

Figure 7. Map of significant wave height for time of maximum wave height in the southern North Sea.

The joint probability method (JPM) and revised JPM have been developed to allow prediction of the extreme value distribution of sea levels from relatively short time-series, e.g. 10 years or less (Pugh & Vassie 1979, 1980; Tawn 1988, 1992), by separating and recombining the effects of astronomical tide (which is predictable over long periods if the tidal constants remain unchanged) and meteorological surge. Trends in sea level and the effects of tide-surge interaction can also be included (Tawn 1992). Further refinement of the JPM is now being used for the interaction between waves and water levels (Hawkes *et al.* 1997).

In order to assess the likelihood of recurrence of such a storm surge we can examine the return period of the surge component by fitting a generalized extreme value (GEV) distribution to the extreme surge levels. Alcock & Blackman (1985) showed that the total (tide plus surge) water level was a much more extreme event than the surge level alone. For the surge component of the 1953 extreme water level their estimates vary from a return period of 66 years at Lowestoft (visual observation of total water level only, 3.35 m; estimated surge, 2.41 m; predicted tide, 0.94 m), 26 years at Harwich (also visually observed, total level, 3.99 m; surge estimate, 2.32 m; tidal level, 1.67 m Ordnance Datum Newlyn (ODN)) to 8 years at Sheerness (recorded water level, 4.70 m; surge, 2.16 m). The lower surge-return periods at some ports are mainly indicative of

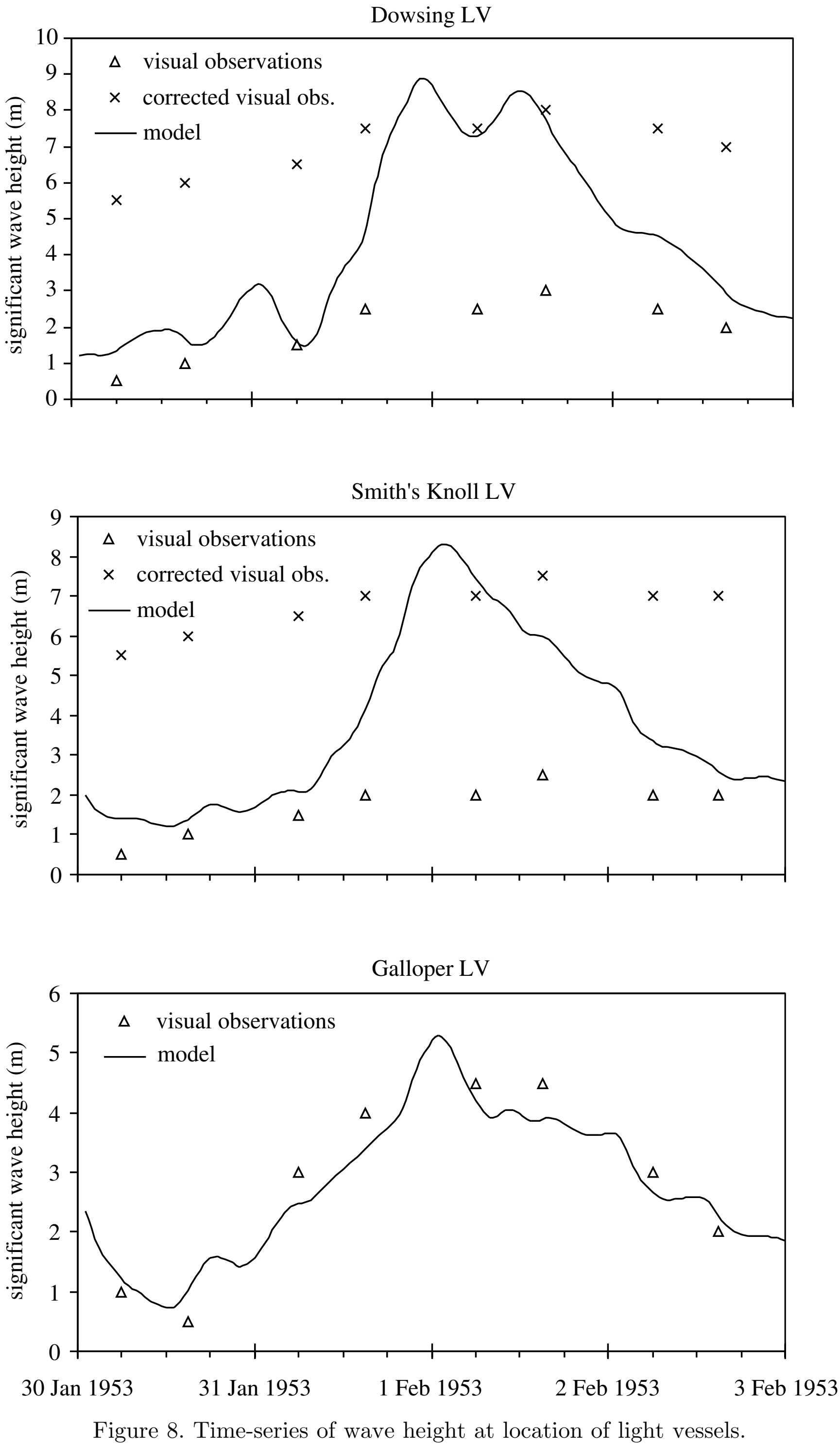

Figure 8. Time-series of wave height at location of light vessels.

the fact that the time of highest total water level is not necessarily the time of the maximum of the surge component.

However, there are problems with the JPM, as pointed out by Tawn (1988). Using a more sophisticated parameter-fitting procedure, Dixon & Tawn (1994) provide parameters which can be used to calculate the return period of the surge at Lowestoft. There is an allowance for a trend through time which produces the return period for the same event of 60 years in 1970, and 43 years in 1990. Lowestoft is a useful location because it has a minimum of surge-tide interaction (Wolf 1978), which complicates the analysis at other locations. The recorded surge level at King's Lynn was 2.4 m and at Southend was 2.29 m at the time of maximum total water level (Flather 1984), so it can be seen that the surge level was rather similar around the East Anglian coast. It would also be expected that the surge-return period would be similar. Comparison of the model-estimated maximum surge elevation (figure 3) with estimates of the surge elevation with return period 50 years, S_{50} (Flather *et al.* 1998), shows that the 1953 event exceeded the 50-year surge in the southern North Sea, but not in the German Bight. Therefore, we use the value of approximately 1 in 50 years as the approximate order of magnitude of the return period of the surge event.

From the STOWASUS and WASA model results and some offshore wave measurements near Cromer (53.07° N, 1.52° E in 31 m mean water depth, December 1985–June 1987), return periods for extreme wave heights off north Norfolk were calculated. The GEV and Gumbel distributions were fitted; in each case the Gumbel distribution was preferred as giving a better fit and the results are quoted below. Clayson & Ewing (1988) estimate a 50 year return period significant wave height, H_S, of 7.6 m for the Cromer wave data. The STOWASUS data gave a maximum simulated H_S of 7.8 m during the 30 year simulation at the nearest grid-point (53.18° N, 1.82° E) and 50 and 100 year return periods are 7.9 and 8.4 m, respectively. The WASA hindcast study generated 40 years of model wave heights on a 0.5°×1.5° grid. The output point nearest the Norfolk coast was used to examine extreme statistics (53° N, 1.5° E). This gives 50 and 100 year return periods of 7.1 and 7.5 m, respectively, for a location very close to the Cromer wave buoy. Based on the evidence discussed above, it was assumed that the 1953 event was at least a 1 in 50 year wave event.

7. What would the 1953 surge produce in 2075?

Flather *et al.* (2001) suggested that by 2075 the 50 year return period surge, S_{50}, could increase by approximately 10 cm on the east coast south of Flamborough Head, and on the Lancashire coast, but decrease by about 10 cm on the south coast. However, these results were subject to considerable uncertainty. The total water level experienced in the 1953 storm was more extreme than the surge event alone because it was nearly coincident with high tide. Purely as a result of the predicted increase in sea levels, which is the parameter that can be predicted with most confidence, the water levels achieved in 1953 are likely to be attained or exceeded more frequently in the future.

Fifty years after the 1953 event, given rising sea level and climate change, it is of interest to try to estimate what the 1953 surge would be like if it occurred towards the end of this century. In order to address this question, a scenario has

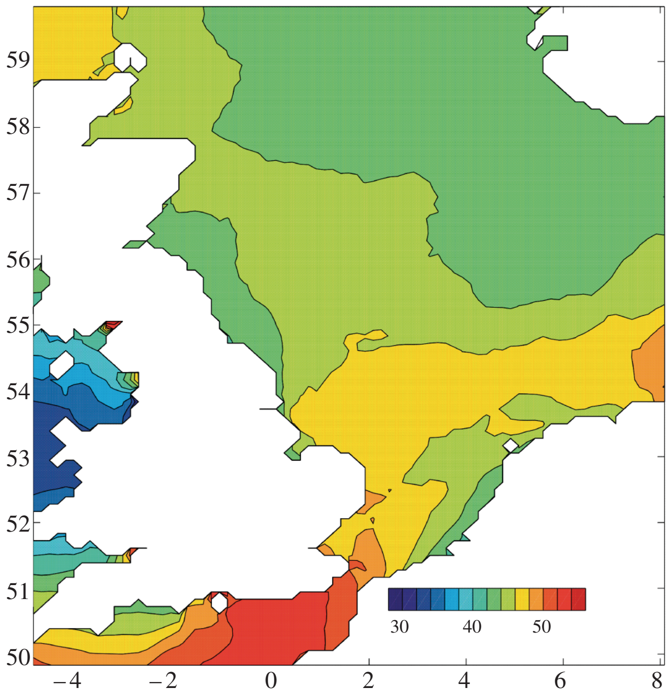

Figure 9. Change in maximum sea level relative to the land for the 1953 event occurring in 2075.

been produced by assuming an identical storm occurred with the same tidal conditions but modified by effects of sea-level rise and land movements. The regional change in MSL for northwest Europe was taken to be +40 cm (Church *et al.* 2001). This was combined with the change in land elevation estimated using the model of Lambeck & Johnston (1995), to give a distribution of change in 'relative' MSL; i.e. the MSL relative to the land. The POL CS3 model was then rerun with the same meteorological forcing and tidal input but the modified MSL, resulting in changed water depths. The difference in water depth modifies the tide and surge dynamics, producing water levels different from those in the original solution. Figure 9 shows the resulting difference between maximum water level in 1953 and projected to 2075. The main difference is due to MSL rise and the land movements but modified by effects of changes in tide and surge dynamics.

Estimating possible changes in wave height is more problematical. It has been demonstrated that wave heights in the northeast Atlantic have increased over the last few decades and this is correlated with a decadal cycle of atmospheric pressure difference called the North Atlantic oscillation (NAO; Woolf *et al.* 2002). It is uncertain whether the NAO will increase or decrease in the future (Tsimplis *et al.* 2005). Also, its impact in the North Sea is much less than on the west coast of the British Isles since it is related to the strength of westerly winds. The predictability of the type of storm events which cause severe storm surges and waves in the North Sea is not significantly related to this index. Prediction of changes of storminess, i.e. severity of storms, their frequency and storm tracks still requires the further refinement of climate models.

NCEP reanalysis data were provided by the NOAA-CIRES Climate Diagnostics Center, Boulder, Colorado, USA, from their website at http://www.cdc.noaa.gov/. Thanks to Emma Young and Mike Meredith for help in extracting NCEP data, and to Ian MacGregor at the Met Office who supplied light vessel records from the National Meteorological Archive.

Appendix A. Wind scaling in the atmospheric boundary layer

By the similarity hypothesis, the wind in the atmospheric boundary layer may be assumed to have a logarithmic profile with height above the surface, z.

$$U(z) = \frac{U_*}{\kappa}\log_e(z/z_0),$$

where κ is von Karman's constant, z_0 is a roughness length and U_* is the friction velocity, related to the wind speed at 10 m, U_{10} by $U_*^2 = C_D U_{10}^2$, where

$$C_D = \sqrt{\frac{\kappa}{\log_e(z) - \log_e(z_0)}},$$

is a drag coefficient.

Thus, for two heights, h_1 and h_2 above the surface, we have

$$\frac{U(h_1)}{U(h_2)} = \frac{\log_e(h_1) - \log_e(z_0)}{\log_e(h_2) - \log_e(z_0)}.$$

We need to know the appropriate roughness length, z_0.

The Charnock coefficient $\beta = gz_0/U_*^2$ is related to C_D and z_0 and hence to wave age but is often treated as a constant in models with values between 0.01 and 0.03. If a constant value for this is chosen, it can be used to derive the effective value of the roughness length from a given drag law model, e.g. Smith & Banke (1975) or Wu (1982). Assuming the wind scaling in the atmospheric boundary layer above the sea obeys this logarithmic law then winds at any level can be derived from winds at another level by a simple scaling factor.

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