Processes Influencing Storm-Induced Currents in the Irish Sea

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ABSTRACT

Although the problem of predicting storm surge elevations has received significant attention, the simulation of currents has suffered because of lack of current observations during surges. Current measurements made during surge conditions are presented here and are used in combination with three-dimensional models to understand processes producing storm currents in the Irish Sea. A coarse-grid (resolution of order 7 km) model of the west coast of Britain together with a fine-grid (of order 1 km) model of the eastern Irish Sea is used to examine the processes, namely, open boundary forcing of the west coast model and wind fields, that produced flows within the eastern Irish Sea during the storm surge of November 1977. Simulations of the surge show that the fine-grid model nested within the west coast model can reproduce observed coastal changes in surge elevation. However, an observed major inflow that was recorded by current meters in the region, prior to a storm surge elevation peak, is not represented, although subsequent inflows and outflows are reproduced. The flow fields in the west coast model giving rise to these currents are analyzed in detail. Also, computations are performed with idealized open boundary forcing and wind fields to understand their role in determining the circulation within the region. An analysis of computed flows shows that outflows from the eastern Irish Sea following major storm events are determined by sea surface elevation gradients in the region and topographic effects. Observed flows under these conditions are reproduced by the model. Inflows, however, are more difficult to compute and depend upon a delicate balance of northern and southern boundary forcing of the west coast model and wind fields over the region. The first observed inflow event, which was not reproduced in the model, was associated with a current from the south. A second inflow event that was reproduced arose from a combination of an inflow from north and south, and a third event was again reproduced in the model due to a current from the north. Without a more comprehensive observational dataset, it was not possible to determine the exact reason why the first inflow was not reproduced.

1. Introduction

In the past the main emphasis in storm surge prediction has been the computation of surge elevations. Very significant progress has been made in this field since the early work of Heaps (1983) and Heaps and Jones (1979), and two-dimensional vertically integrated hydrodynamic models are now in operational use providing an early warning of flooding. In recent years there has been an increasing emphasis in predicting currents during extreme wind conditions. The episodic nature of these events, and the strength of the currents associated with them, is such that little observational data exists for model validation. However, such a dataset does exist in the eastern Irish Sea for the storm event of November 1977, and is the focus of this paper. The significant spatial change, both in the vertical and horizontal, found in wind-induced currents (Davies et al. 1998, 2000) means that three-dimensional hydrodynamic models with a fine horizontal grid are required in these calculations.

Although three-dimensional shelfwide models have been developed and used to examine major storms (Davies et al. 1998), the finite-difference grid in such models is rather coarse (of order 12 km) and, consequently, cannot resolve regions such as the eastern Irish Sea. To improve resolution, Heaps and Jones (1979) developed a two-dimensional model of the west coast with a grid resolution of order 7 km (Fig. 1a) and used it to successfully simulate storm surge elevations. Subsequently a three-dimensional version of the model was developed (Davies and Jones 1992) and applied to the computation of surge currents. This model was also used (Jones and Davies 1998) to examine the role of the external surge (viz. that propagating into the region, as compared with the internal surge, the component produced by wind-forceing over the region) in determining the total surge. The external surge was only examined in terms of its contribution to surge elevation since the west coast model used an elevation-specified open boundary condition. To improve storm surge elevation predictions in the east-
Fig. 1. (a) Finite-difference grid of the west coast model showing location of grid points used in the comparison with coastal gauges. Also shown are positions D (solid square) and B (open circle) where current observations are made. (b) Finite-difference grid of the high-resolution eastern Irish Sea model, showing location of positions D and B where current observations were made.

In this paper the west coast and eastern Irish Sea models are used initially to examine the accuracy of the time series of computed currents at locations B and D (Fig. 1b). Differences and similarities between computed and observed currents at these locations are related to the flow fields along the boundary between the west coast and eastern Irish Sea models. As shown previously (Jones and Davies 1998), although never analyzed in detail, the flow along this boundary is important in determining currents within the eastern Irish Sea. To understand how currents along the open boundary of the eastern Irish Sea are affected by west coast winds, and “far-field” flows, the west coast model is forced by idealized winds and open boundary current input. To accomplish this, the west coast model is modified so that a radiation condition can be applied. The response of the Celtic Sea and Irish Sea to external surge events, introduced as currents through the open boundary of the model, is examined in detail to determine the role of these flows upon currents in the eastern Irish Sea.

A brief overview of the three-dimensional models is given in the next section with subsequent sections dealing with the storm surge of November 1977, in particular, currents in the eastern Irish Sea, and subsequently the circulation induced by current forcing through the open boundary (external surge), and the response to uniform and spatially varying wind forcing (internal surge). A final section summarizes the main findings.
2. The three-dimensional hydrodynamic models

As details of the models have been presented elsewhere (Jones and Davies 2001; Davies et al. 1997a,b) only the main features will be outlined here. The models are identical except in their geographical extent and grid resolution (Figs. 1a,b). In the simulation of the storm surge discussed subsequently, an elevation-specified open boundary was used in the west coast model, since only elevations based on observations were available for model forcing. This boundary forcing is identical to that used by Heaps and Jones (1979) in a two-dimensional simulation of the surge. The eastern Irish Sea model used a radiation condition with elevations and currents interpolated from the west coast model. Since this model could resolve near coastal shallow regions (Fig. 2), wave-current interaction effects (Grant and Madsen 1979) were included (Davies and Lawrence 1994, 1995; Jones and Davies 1998).

In subsequent calculations with idealized current forcing and winds the west coast model had a radiation open boundary condition. Since the region (Fig. 1a) covers a substantial range of latitude, the models are based on the three-dimensional, nonlinear hydrodynamic equations expressed in spherical coordinates, namely,

$$\frac{\partial \zeta}{\partial t} + \frac{1}{R \cos \phi} \frac{\partial}{\partial \phi} \int_{-h}^{h} \nu \cos \phi \, dz$$

$$+ \frac{1}{R \cos \phi} \frac{\partial}{\partial \chi} \int_{-h}^{h} u \, dz = 0 \quad (1)$$

$$\frac{\partial u}{\partial t} + S_v + f v = -\frac{g}{R \cos \phi} \frac{\partial \zeta}{\partial \chi} + \frac{\partial}{\partial \chi} \left( A_v \frac{\partial u}{\partial \chi} \right) \quad (2)$$

$$\frac{\partial v}{\partial t} + S_v + f u = \frac{g}{R \cos \phi} \frac{\partial \zeta}{\partial \phi} + \frac{\partial}{\partial \phi} \left( A_v \frac{\partial v}{\partial \phi} \right), \quad (3)$$

where $S_v$ and $S_v$ are the nonlinear momentum advective terms, with $\chi$ and $\phi$ being longitude and latitude, respectively; $z$ vertical coordinate; $h$ undisturbed depth; $\zeta$ free surface elevation; $t$ time; $f$ Coriolis term; $R$ radius of earth; $g$ acceleration due to gravity; $u$ and $v$ eastward and northward currents, respectively; and $A_v$ vertical eddy viscosity. The assumption of a homogeneous sea region is valid during storm conditions due to extensive mixing. The numerical solution involves initially transforming the equations to a normalized sigma ($\sigma$) coordinate in the vertical. Discretization is then accomplished using an expansion of functions through the vertical. These functions are chosen as eigenfunctions of the eddy viscosity profile. This expansion method yields a continuous current profile from sea surface to sea bed. Discretization in the horizontal is accomplished using the Arakawa C finite difference grid, and a time-split method is used to integrate the hydrodynamic equations in time. In the region of open boundaries or narrow coastal channels where there were insufficient grid points to enable the advective terms to be accurately computed, they were omitted. This yielded an accurate well-posed solution. On the C grid in a shallow high friction dissipation region a stable solution was obtained with the function model without the horizontal viscous terms. These are required in low friction dissipation regions (e.g., Davies et al. 1998, 2000).

In the calculations described subsequently the vertical eddy viscosity $A_v$ is expressed as $A_v = \alpha(\chi, \phi, t)\psi(\sigma)$, with $\psi(\sigma) = 1.0$ (viz. eddy viscosity constant in the vertical), although the influence of varying $\psi(\sigma)$ was examined by Jones and Davies (1998). The coefficient $\alpha = K_1(\pi^2 + \tau^2)^{1/2}h$ with $\pi$ and $\tau$ depth mean currents, and $K_1 = 0.0025$. This parameterization of viscosity is consistent with measurements and results from turbulence closure models (Davies 1991). At land boundaries the normal component of the current was zero. The radiation condition used on open boundaries is given by

$$q = \frac{C}{h} (\zeta - \zeta_r - \zeta_s) + q_r + q_s,$$

where $\zeta_r$ and $\zeta_s$ are tidal elevation and current normal to the open boundary, respectively, taken from Jones.
and Davies (1996) as described later (section 3); and $\zeta$, and $q$, are the surge elevation and normal current from the west coast model, respectively (see section 3), with $C = (gh)^{1/2}$. For storm surge computations, the surface stress was set equal to the external wind stress components $F_s$ and $G_s$. At the sea bed a quadratic friction law was applied in terms of bed stresses $F_B$ and $G_B$ in which the coefficient of bottom friction can increase because of wave–current interaction effects (Davies and Lawrence 1994, 1995).
To understand the importance of the various terms in determining the storm-induced currents within the eastern Irish Sea, Eqs. (2) and (3) were integrated in the vertical to yield their depth mean forms, namely,
\[
\frac{\partial \eta}{\partial t} - f \eta + \bar{S}_x = \frac{-g}{R \cos \phi} \frac{\partial \bar{\zeta}}{\partial \phi} + \frac{F_s}{\rho h} - \frac{F_a}{\rho h} \quad \text{and} \quad (4)
\]
\[
\frac{\partial \bar{v}}{\partial t} + f \bar{v} + \bar{S}_y = \frac{-g}{R \cos \phi} \frac{\partial \bar{\zeta}}{\partial \phi} + \frac{G_x}{\rho h} - \frac{G_y}{\rho h} \quad \text{and} \quad (5)
\]
Time series of the various terms in (4) and (5) at location D (similar results were found at B) where current measurements are available will be presented in the next section.

3. Storm-surge computation

a. Overview of storm surge

The storm-surge period 7–17 November 1977 is particularly important in that current measurements are available for model comparison in addition to shore-based elevation records. Also, the two main surge events (in which there was a major increase in elevation at Liverpool, United Kingdom, although there were a number of inflows and outflows during the period) that took place during the period were produced by different wind field distributions. The time series of wind stresses in the central eastern Irish Sea (Fig. 3a) illustrate the temporal variability of the wind. In the model the wind stress varied with both space and time and was derived from accurate meteorological wind fields produced after the event. The first surge elevation peak at Liverpool (11–12 Nov) was produced by a depression that moved from west to east across the area to the north of Scotland and then onto Norway. In the second (13–15 Nov) a depression followed a more southerly track moving from off the west coast of Scotland over the North Sea toward Denmark. The first surge event occurred close to the time of tidal high water. In fact total sea surface elevation was one of the largest ever recorded at Liverpool. The second surge event occurred close to the time of low water.

The atmospheric forcing situation at the time of the first surge is characterized by strong winds (stresses exceeding 0.5 Pa) from the southwest at 1200 UTC 11 November 1977. The magnitude of this wind stress increased with time (Fig. 3a) and direction changed to a wind from the west, of magnitude 1.5 Pa off Anglesey, United Kingdom, decreasing to 0.5 Pa in the north of the eastern Irish Sea at 0000 UTC 12 November 1977. Over the next 6 h the wind stress decreased. The second surge was produced by strong winds from the northwest, and by 0000 UTC 14 November 1977 a region of wind stress of over 1.0 Pa occurred off the west coast of Scotland, although over the Irish Sea the stress was less than 0.5 Pa. The wind stress off the west coast of Scotland increased to 2.0 Pa over the next 12 h and drove water through the North Channel and into the eastern Irish Sea, producing the storm event in the region.

b. Storm-surge computation

Since tidal currents are significant, the five dominant tidal constituents, namely $M_2, S_2, N_2, K_1,$ and $O_1,$ which have previously been accurately reproduced (Davies and Jones 1992), were included as input along the open boundary of the west coast model. Although the storm surge currents in the eastern Irish Sea were significant (of order 0.5 m s$^{-1}$) during the period, the flow is dominated by the tidal currents, which are of order 1.0 m s$^{-1}$. In the absence of elevations and currents from a large-area model, observed surge elevations from Castletownsend and Newlyn, United Kingdom (Fig. 1a), were interpolated along the southern boundary of the model, with observations from Malin, United Kingdom (Fig. 1a), imposed along the northern boundary. By this means some account of surge events outside the region covered by the west coast model was included. These lines and the interpolation method were identical to those used by Heaps and Jones (1979), who showed that they gave appropriate open boundary conditions for an accurate simulation of surge elevations (although currents were never examined) in a two-dimensional surge model of the area. The surge elevations and currents computed with the model were determined by subtracting a tide-only solution from the tide and surge calculation. Surge elevation and current inputs along the western open boundary of the eastern Irish Sea model were derived by interpolation from the west coast model. Since both elevations and currents were available, a radiation condition was used along this boundary. Also, an accurate description of the $M_2, S_2, N_2, K_1,$ and $O_1$ tides (Jones and Davies 1996) was available for tidal boundary forcing.

Time series of the computed storm-surge elevations at a number of eastern Irish Sea ports (Fig. 3b) show that at Douglas, United Kingdom, a relatively deep water port, the major features of the surge peaks at 0000 UTC 12 November and 0000 UTC 14 November are reproduced by the model. At Workington, United Kingdom, a shallow water port, the observations show a strong signal at the tidal period that was not present in the model, although the main features of the peaks are reproduced. At Heysham and Hilbre, United Kingdom,

![Fig. 4. Depth mean currents (i) over whole region, and (ii) “blow up” in region of eastern Irish Sea model open boundary at (a) 1800 UTC 11 Nov, (b) 0600 UTC 12 Nov, (c) 0000 UTC 14 Nov, (d) 0000 UTC 15 Nov, and (e) 0600 UTC 15 Nov, computed with the west coast model.](image-url)
the gauge failed during the period. Prior to this failure there were timing errors at these gauges, which may account for the difference in time between the computed and observed surge peak. The time of the surge peak was correctly reproduced at other gauges. At Liverpool the main features of the two surges were reproduced, although not the magnitude of the second surge peak. The root-mean-square error (rmse) at Liverpool for the period was 24.5 cm, reflecting the large surge at this port compared with rmse’s of 9.9 and 32.7 cm, respectively, at Workington and Douglas, United Kingdom. At the other locations, because of gauge failure, the rmse could not be computed.

1) Flow Field During First Surge Event

Based on a comparison of observed and computed \( u \) components of current (Fig. 3c) (the \( v \) component of currents was negligible; Howarth and Jones 1981) at the current meters located at current rigs D (three measurements in the vertical corresponding to \( \sigma = 0.48, \sigma = 0.62, \sigma = 0.80 \)) and B (one measurement at \( \sigma = 0.50 \)), the observed inflow to the eastern Irish Sea, which occurred at 1800 UTC 11 November, prior to the surge peak at Liverpool, was not present in the model. However, the outflow at 0600 UTC 12 November as the surge elevation decreased was present, with the model tending to overestimate the outflow. The small but persistent inflow at 0000 UTC 14 November, which occurred prior to the second surge peak, is found in the computed flow. The short inflow event at 1200 UTC 14 November, which occurred at rig D, was not present in the model. Also, it was not found in the model or observations at rig B. The significant and persistent outflow that took place between 1800 UTC 14 November and 0600 UTC 15 November at both locations as the surge elevation at Liverpool decreased is reproduced in the model. This set of results suggests that the model can reproduce outflow events, or persistent inflows, although the rapid inflow at 1800 UTC 11 November was absent. Despite this the model accurately reproduced the storm surge elevation at Liverpool that immediately followed this
inflow, which suggests that the model must have had some form of inflow into the eastern Irish Sea at this time. The computed rmse’s at the current meters derived from the time series shown in Fig. 3c were as follows: at rig D, $\sigma = 0.48 (3.7 \text{ cm s}^{-1})$, $\sigma = 0.62 (4.2 \text{ cm s}^{-1})$, $\sigma = 0.80 (4.0 \text{ cm s}^{-1})$; and for rig B, a value of 4.0 cm s$^{-1}$.

To understand where this inflow came from it is necessary to examine the spatial distribution of depth mean current vectors at this time [viz. 1800 UTC 11 Nov; Figs. 4a(i),(ii)]. Also, time series of the $u$ component of velocity at selected points (positions 2, 8, 12, 16, and 17; Fig. 1a) in the west coast model adjacent to the open boundary of the eastern Irish Sea model are presented (Fig. 5a). In addition, the time series computed with wind forcing only, and the difference between the full surge computation and the wind-forced solution, which represents the effects from the open boundary, is shown. Current vectors and time series [Figs. 4a(i),(ii), 5a] reveal that at 1800 UTC 11 November, there was an outflow in the region to the north of the Isle of Man, with magnitude decreasing from north to south. The wind-forced solution at this time (Fig. 5a, position 2) does not show a significant peak in the current, but rather a general outflow with a small peak at 0600 UTC 12 November. The open boundary effect does, however, show an oscillatory flow with peak outflows at 1800 UTC 11 November and 0600 UTC 12 November, with the latter producing the maximum outflow in the surge solution. To the south of the Isle of Man there was a small inflow (location 8) at 1800 UTC 11 November, although this inflow only reached a maximum at 0000 UTC 12 November. At this position the open boundary influence is small at this time (Fig. 5a), and the inflow at 0000 UTC 12 November is entirely wind forced and is related to the maximum in the wind stress at this time (Fig. 3a). Current vectors [Fig. 4a(ii)] show that at location 8 the current is mainly aligned north–south with little west–east flow. Farther south (location 12) there was no evidence for the inflow, at 1800 UTC 11 November [Fig. 4a(ii); again the current is mainly aligned in a north–south direction], with only a slight inflow at 0000 UTC 12 November (Fig. 5a). As at location 8, the open boundary influence upon the current appears negligible (Fig. 5a). Also, here the water is much deeper, with a resulting small wind-induced inflow. The maximum inflow occurs at location 17, close to the coast of Anglesey [Figs. 4a(i),(ii), 5a] and it is this inflow that gives rise to the surge elevation increase at 0000 UTC 12 November in the eastern Irish Sea. The inflow of water close to the coast of Anglesey explains why the surge elevation is accurately predicted without the model reproducing the current inflow at positions B and D. In essence, provided there is sufficient inflow into the Liverpool Bay region between the Isle of Man and Anglesey, the surge elevation at Liverpool is accurately reproduced, although there may be errors in the spatial distribution of this inflow. From Fig. 5a, it is evident that at this time the southern open boundary effect upon the inflow to the south of the Isle of Man is small, and the inflow is due to the wind stress acting over shallow water near Anglesey producing a strong current. This suggests that the west coast model fails to reproduce the inflow at B and D yet produces the surge elevation at Liverpool, possibly due to an underestimation of the effect of the northern open boundary or due to an error in the wind field over the region. As we will show, this boundary influences the flow through the North Channel, which affects the spatial distribution of the flow into the eastern Irish Sea in the region both to the north and south of the Isle of Man. This point is reexamined later in this section after the surge event of 0000 UTC 14 November.

The outflow that occurred at 0600 UTC 12 November was reproduced by the model. It is evident from current vectors [Figs. 4b(i),(ii)] and time series at various points (Fig. 5a) that there was outflow to the south of the Isle of Man with the area of maximum outflow near location 12, to the east of D and B. The maximum outflow was confined to the deep water region where bottom frictional effects were a minimum. Prior to the outflow there was a balance between the sea surface elevation increase in the eastern Irish Sea, and the wind to the west in the region. As the wind field decreased, the surface ele-
vation pressure gradient gave rise to an essentially uniform outflow from the region that could be reproduced by the model. The outflow at location 2 was influenced by the boundary, although the major contribution was in the wind effects increasing to the south of this point. At location 8, boundary and wind effects were comparable, with wind effects increasing to the south of this point.

2) Flow field during second surge event

For the second surge event, there was an inflow at all points (Fig. 5a) between 1800 UTC 13 November and 0600 UTC 14 November [see vector plots at 0000 UTC 14 Nov; Figs. 4c(i),(ii)]. At location 2, as for the previous surge, the main influence was the open boundary effect. However, for locations to the south of the Isle of Man, the two effects were comparable, with wind-forced flow slightly dominant. The contribution from the wind forcing, unlike previously, was constant along the whole of the eastern Irish Sea boundary. This result suggests that it was not local wind forcing, and this is confirmed by the time series of local winds (Fig. 3), which have no significant maximum at this time, unlike the wind-forced event at 0000 UTC 12 November. This inflow is the result of wind fields acting over the whole model domain. The spatially uniform nature and long duration of this inflow, and the fact that it was not produced by local winds, appears to be one reason that it is reproduced by the model at current locations B and D. The far-field (i.e., west coast model) flow field associated with this inflow [Fig. 4c(i)] has a flow to the north from the south, converging with a flow to the south from the north, in the region to the west of the Isle of Man. In effect the flow to the north is prevented from passing through the North Channel [the situation found previously; Fig. 4a(i)] and is diverted to the east, giving a uniform flow into the eastern Irish Sea. This suggests that previously at 1800 UTC 11 November [Figs. 4a(i),(ii)], where the inflow was confined to the region close to Anglesey, there was too large an outflow in the North Channel, possibly due to an error in the distribution of the inflow through the northern boundary of the west coast model, or incorrect wind fields. The sensitivity to these effects is examined later.

The following small outflow event, the subsequent inflow, and major outflow to the south of the Isle of Man at 0000 UTC 15 November [Figs. 4d(i),(ii)] are fairly uniform at locations 8, 12, and 16, although with an increasing magnitude of the outflow event in progressing from north to south. The sharp inflow peak at location D at 1200 UTC 14 November is not present at any of the inflow points, although the inflow at 0600 UTC 15 November [Figs. 4e(i),(ii)] is present at all points and reproduced by the model at locations D and B. This inflow is produced by a current to the south that passed through the North Channel, with part of the transport continuing south, and some flow deflected to the east into the eastern Irish Sea [Figs. 4e(i),(ii)]. The fact that the inflow is produced by a current from the north appears to be the reason why the current is not increased in the coastal region north of Anglesey as found with the inflow from the south [Figs. 4a(i),(ii)]. The three inflow events [Figs. 4a(i), 4c(i), 4e(i)] in the region close to rigs D and B are due to significantly different far-field flow regimes, and the processes producing these far-field currents are examined in the next section. Before considering the processes producing these far-field currents it is instructive to look at the various terms in Eqs. (4) and (5) in the region between the Isle of Man and the coast of Anglesey. For illustrative purposes these are examined at rig D, as comparable results apply elsewhere.

3) Balance of terms

Time series of the various terms in the $u$ equation of motion due to wind and open boundary forcing after removing the tide shows [Fig. 5b(i)] that at the time of the first surge elevation maximum at Liverpool (0000 UTC 12 November) the predominant balance is between the wind stress term and sea surface elevation gradient. The acceleration term is also significant at the time of maximum elevation gradient [Fig. 5b(i)]. A similar balance occurs at the time of the second surge elevation peak at Liverpool (1800 UTC 14 Nov). Although the wind stress term has a smaller maximum value at this time, it has persisted for longer. This appears to be the reason why the elevation gradient term is comparable to that during the first surge. Also, the wind stress shows some time variability close to its maximum value, which produces some variability in the acceleration term. The rapid changes in the acceleration term and the increase in the bed stress at times of maximum surge elevation are due to changes in the velocity field [Fig. 5d(i)], which shows a major outflow (negative $u$ velocity) as elevations decrease following the time of surge elevation peak.

The contribution from the Coriolis term $fu$ is small [Fig. 5b(i)] as the $v$ component of velocity throughout the period was less than 5 cm s$^{-1}$ [Fig. 5e(i)], in agreement with observations (Howarth and Jones 1981). The contribution from the nonlinear term $S_{uv}$ (not shown) was close to zero at this location. Also, detiding the solution by subtracting a tide-only calculation from the total was very successful in this region. This shows that the assumption of a linear decomposition of the current field was valid. However, in very shallow (less than 10 m) near-coastal regions in the eastern Irish Sea, as shown by Davies and Lawrence (1994, 1995) and Davies and Jones (1996), the nonlinear terms are important and wind effects modify the tidal currents making it very difficult to remove tidal currents from the total solution. However, since our main focus here is in the deeper regions, in particular rigs D and B, the assumption of linearity is valid.
In the case of the terms in the $v$ equation of motion [Fig. 5c(i)] the Coriolis term [significantly larger than in Fig. 5b(i), due to the larger $u$ component of velocity] together with the elevation gradient term [smaller than in Fig. 5b(i) due to the slight increase in elevation along the coast of North Wales] are important in balancing the wind forcing. The contribution from $S_v$ was negligible.

The contribution of the various terms from wind forcing only [Fig. 5b(ii)] has a similar time variability to that found with the total surge [Fig. 5b(i)], although the acceleration term has been reduced and there are changes in the elevation gradient term. Although the various terms due to open boundary forcing [Fig. 5b(iii)] are smaller than from the wind forcing, it does contribute to the acceleration term and associated elevation gradient term. Also, the contribution to the $u$ velocity at position D is comparable and sometimes exceeds that due to the wind forcing.

Similar conclusions can be drawn from the time series of terms in the $v$ equation of motion [Figs. 5c(ii),(iii)] and the $v$ currents [Figs. 5e(ii),(iii)]. This examination of the various terms in the hydrodynamic equations justifies the linear separation in the region of the eastern Irish Sea into a contribution due to open boundary forcing and wind forcing. Advantage is taken of this in the next section where these are examined independently. Also, the fact that both open boundary forcing and wind forcing contribute to the current at rig D shows that each must be examined in some detail as discussed in the next section.

4. Idealized forcing

To understand the response of the currents off the west coast of Britain to meteorological forcing over the region, and outside the region (introduced as open boundary forcing), it is essential to perform calculations with each of these forcings. As shown, away from coastal regions such a separation is possible due to the linearity of the problem.

Initially, the steady-state response to open boundary forcing is examined in this section. Subsequently, the response to idealized meteorological forcing is considered. If linearity is assumed, then the response to a spatial constant wind stress from an arbitrary direction can be obtained from the flow due to two orthogonal winds, taken here as winds from the west and south. If the wind field is nonuniform, then the response due to wind stresses with north–south and west–east gradients is also required and given here. By this means, steady-state flows due to arbitrary boundary and wind stress forcing are obtained that may then be used to understand flow pathways due to a range of wind events and boundary forcings. Besides discussing the flow fields due to this range of forcings, the flows are used to understand the currents produced during the November 1977 surge. A complex interplay between wind-forced and boundary-forced flows is shown to occur. Idealized calculations with time-dependent forcing are also possible but outside the scope of this study.

a. Open boundary current forcing

In an initial series of calculations with idealized forcing (Table 1) designed to determine the influence of flow through the open boundaries upon currents in the eastern Irish Sea, the model was forced by inflow through the five sections of the open boundary given in Fig. 1a. The topography of the area (Fig. 2) is such that these boundaries are located in deep water, far removed from the shallower eastern Irish Sea.

An inflow through the northern part of the north boundary produces a flow to the south, part of which passes through the North Channel (Fig. 6a). The flow to the south is topographically steered (in the sense that the flow is parallel to contours of constant water depth) along the western side of the Irish Sea by the deeper water channel in this region. Some water enters the northern part of the eastern Irish Sea in the region to the north of the Isle of Man, leaving the area in the region to the south (Fig. 6a).

In the case of an inflow through the north west boundary (not shown), the majority of the current flows along the north coast of Ireland and through the North Channel. As previously, within the Irish Sea, the flow pathway is influenced by topography. These calculations demonstrate that in the absence of wind the flow paths within the Irish Sea are influenced by bottom topography and friction. The inflow through the North Channel and the subsequent flow to the south in the deep water channel of the Irish Sea is analogous to that found at 0600 UTC 15 November [Fig. 4e(i)]. The northern boundary forcing gives an inflow to the eastern Irish Sea to the north of the Isle of Man, similar to that shown in Figs. 4e(i),(ii). However, since Fig. 6a is a steady-state solution, this inflow has to be balanced by an outflow (otherwise surface elevations change with time; a non-steady-state solution), which occurs to the south of the Isle of Man. This outflow is not found in Figs. 4e(i),(ii), possibly due to the fact that this is not a steady-state solution, and also due to wind forcing, which will be considered later.

The inflow through the southwest open boundary produces a current to the north (Fig. 6b), although the majority of the flow leaves the area through the southern boundary. An inflow along the southern boundary is more effective in producing a flow to the north into the Irish Sea (Fig. 6c). This flow is topographically steered along the deep water channel in the western Irish Sea, leaving the area through the North Channel. Some water does, however, enter the eastern Irish Sea in the region to the south of the Isle of Man.

Although some of the inflow through the southeastern open boundary (Fig. 6d) does not go to the north the majority does and produces a current along the English coast that enters the eastern Irish Sea. In all cases, flow
from the southern boundary produces a flow to the north that is mainly confined within the deep water channel of the western Irish Sea. Some water does enter the eastern Irish Sea, with maximum inflow in the shallow region to the north of Anglesey. There is a flow to the north in the central part of the eastern Irish Sea, with an outflow in the region to the north of the Isle of Man. This flow field is very similar to that given in Figs. 4d(i),(ii), and suggests that topographic effects within the eastern Irish Sea influence the spatial distribution of flows from the south entering or leaving the region. However, as discussed previously the observations at D and B suggest an inflow in the region to the north of Anglesey that is not present in the model. The fact that a uniform inflow is found in both model and observations at 0000 UTC 14 November suggests that at 1800 UTC 11 November [Figs. 4a(i),(ii)] there is an imbalance in the north–south boundary forcing (which we will consider later) or the wind effect is incorrect. The influence of the wind is considered next.

b. Wind-induced forcing

Besides currents being influenced by open boundary effects they are also produced by wind events over the region. To examine this, steady-state depth mean currents due to a uniform wind stress from the west and south were examined. A uniform wind stress from the west produces a flow along the north coast of Ireland in the direction of the wind (Fig. 7a). This current is channeled to flow to the south in the North Channel. There is a small inflow to the eastern Irish Sea in the region north of the Isle of Man, with a uniform outflow to the south.

The flow pathways are similar to those due to a northern boundary inflow and appear to be influenced by topography. Within the eastern Irish Sea the surface current (not shown) moves water onshore in the wind direction, with offshore flow below this layer in the southern part of region, channeled in deep water where bottom frictional effects upon flow above the bottom boundary layer are a minimum. This vertical shear in the coastal regions of the eastern Irish Sea was also found in the simulation of the November 1977 surge at times of onshore winds. In the absence of current profile measurements in the region it was not possible to determine the accuracy of the computed shear. The outflow in deep water to the south of the Isle of Man (Fig. 7a) has a similar spatial distribution to the outflow given in Figs. 4d(i),(ii), suggesting that during such events the current pathway is determined by topography rather than the exact details of the wind field or the open boundary forcing. This in part explains the success of the model in reproducing the outflow currents at D and B.

In the case of a uniform wind from the south, there is a flow to the north in the western Irish Sea (Fig. 7b). Some of this water is driven to the east in the region between the Isle of Man and Anglesey and flows into the eastern Irish Sea. The strongest currents to the east occur close to the coast of Anglesey, with current magnitude decreasing on going northward as the Isle of Man is approached. To the north of the Isle of Man, there is an outflow from the eastern Irish Sea (Fig. 7b). The main features of this flow are similar to those found with inflow through the southern boundary (Figs. 6b–d). However, unlike in the boundary-forced case where current magnitudes did not vary significantly over the eastern Irish Sea, in the wind-forced case, current magnitudes are largest in shallow water. If complementary measurements to D and B had existed in shallow water, there would have been possible to distinguish between the effects of a wind from the south and southern boundary forcing. Without these it is difficult to determine the contribution of each at location D and B.

Besides considering flows due to a uniform wind, currents produced by a spatially variable wind were examined. Initially, a wind pattern in which there is a wind from the north over the western part of the region, changing linearly to a wind from the south over the eastern area, was considered. The depth mean currents (Fig. 7c) in the eastern Irish Sea have a similar distribution to that found in Fig. 7b. In this area the winds are comparable, namely, a wind from the south driving flows in shallow water, namely, the eastern Irish Sea in the wind direction. Over the western Irish Sea where there is a transition from a wind from the south to one from the north, the wind field is weak, as are the currents. This wind distribution gives similar currents in the eastern Irish Sea to those found with a wind from the south, without any appreciable flow in the North Channel. This suggests that such a component of the wind field, combined with an inflow through the North Channel forced by the northern open boundary (Fig. 6a) could produce a similar flow field to that in Fig. 4c(i).

In the case of a north–south gradient in the wind field, namely, a wind from the east in the north of the model to one from the west in the south of the model, the currents in the eastern Irish Sea (Fig. 7d) have a similar distribution to those found with a wind from the south (Fig. 7b) as described previously. In this case there is a strong wind from the east off the north coast of Ireland, which produces a flow in the wind direction in the region. This wind gives rise to the strong flow toward the north through the North Channel. In the southern part

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Fig. 6. Depth mean currents due to an inflow through (a) northern section of northern boundary, (b) western section of southern boundary, (c) southern section of southern boundary, and (d) eastern section of southern boundary.
of the domain (the Celtic Sea) the wind is toward the east, giving a flow to the east (Fig. 7d).

Although these steady-state flow fields cannot explain all the details of a storm surge event in which time variability is important, they give some insight into the role of topography, open boundary, and wind forcing. As discussed above, without more comprehensive current measurements over the entire west coast region it is difficult to determine if the failure of the model to reproduce the inflow observed at locations D and B on 11 November was due to errors in the west coast model boundary forcing or inaccuracies in the wind field.

5. Concluding remarks

Three-dimensional models of the west coast of Britain and the eastern Irish Sea have been used to investigate the storm surges of November 1977. Calculations have shown that by using a high-resolution model of the eastern Irish Sea that incorporates wave–current interaction effects with open boundary input taken from the west coast model, it is possible to accurately reproduce elevations in the region. However, comparison with current measurements made in the eastern Irish Sea have shown that the models failed to reproduce the observed inflow at 1800 UTC 11 November. Major inflows at other times were reproduced, as were the two outflows during the period. A detailed examination of the current fields showed that the first inflow to the eastern Irish Sea was associated with a flow from the south that produced an inflow close to the coast of Anglesey, and failed to reproduce the flow to the north of that inflow. The second inflow was associated with currents from the north and south that produced a spatially uniform inflow in the region to the south of the Isle of Man, which was reproduced by the model. The final inflow that was also reproduced was associated with a flow from the north. Outflows events that followed major surge elevation peaks gave maximum currents in the deep water region where bottom frictional effects were a minimum. For a uniform wind from the south, the flow pattern in the eastern Irish Sea was comparable to that produced by forcing at the southern boundary of the west coast model. However, current magnitudes varied over the region, with strongest currents in shallow water where the wind forcing effect was largest.

Flow patterns due to spatially variable winds showed similar circulations in the eastern Irish Sea to those found with uniform winds. However, in the western Irish Sea where these wind fields were small the flows were significantly different from those found with uniform winds.

The calculations presented here show that although a storm surge model can compute changes in elevation at coastal gauges within the eastern Irish Sea, this does not guarantee that it can reproduce the associated currents. At times of outflow following major storm events, where the major driving force is the local sea surface elevation gradient and where currents are parallel to contours of constant water depth (viz. follow f/h contours), the model can reproduce the currents. However, as shown previously (Jones and Davies 1998) flows entering the eastern Irish Sea to the north and south of the Isle of Man are more important in determining the flow field in the region than the local wind. In such cases the west coast model must correctly reproduce these inputs. As shown here these inputs are determined by a delicate balance between northern and southern open boundary forcing to the west coast model and wind fields over the area. To determine if the model has the correct contribution from each and, hence, can reproduce flows in the region, a more comprehensive dataset of current measurements than that available here is required. Based on the flow fields discussed here, measurements both to the north and south of the Isle of Man, with complementary deployments in the Celtic Sea and North Channel for far-field inflows are required. Results from recent measurements in the North Channel.
(Knight and Howarth 1999; Davies et al. 2001) have also demonstrated the importance of this region in determining the circulation of the Irish Sea.

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