Simulation of the Norwegian Coastal Current in the vicinity of the Halten Bank: comparison with observations and process study of bank-induced meanders *

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ABSTRACT


Results from a high-resolution, three-dimensional, real-time simulation of the Norwegian Coastal Current are compared at six current meter moorings deployed during March 1988 in the vicinity of the Halten Bank. The simulation was initialised from a set of fairly arbitrary velocity and density fields. The objectives are to examine (i) how accurately the results reproduce the observed means and variabilities, and (ii) how the dominant flow dynamics may be explained in term of the circulation produced by unstable baroclinic meanders over topography.

The simulation reproduces the means fairly well; the rms errors are less than 34% in five of the six moorings for the east/west velocity component, and are less than 32% in three moorings for the north/south component. The agreements in current variabilities are good only at three moorings, for which an averaged rms error of about 22% was obtained. At the other three moorings, the largest errors for the variabilities occur in the subsurface, where energetics are underestimated by as much as 60% or more in the simulation. The discrepancies are most likely due to insufficient vertical resolution, which results in a poor representation of the baroclinic structure, and also due to the smoothed topography used in the simulation. On the other hand, a meander upstream of the Halten Bank on March 26, 1988, is reproduced well by the model. The simulation suggests that the meander is a result of amplification of waves and eddies shed from a smaller bank upstream of the Halten Bank through dynamic instability. A process-oriented simulation has been conducted to support this hypothesis.

Introduction

The portion of the Norwegian Coastal Current (NCC) off the west northwestern coast of Norway is a good example of a current system influenced by both large-scale ocean dynamics—the North Atlantic inflow and the Norwegian Sea, and coastal discharges from rivers in the North Sea, fjords from the Scandinavian coast and the Baltic Sea (Sverdrup et al., 1946). In winter, the NCC can be distinguished in satellite imagery as a band of low-temperature water hugging the Norwegian coast from the Skagerrak to the Barents Sea shelf—a distance of more than 1500 km (Mork, 1981; Audunson et al., 1981). Satellite imagery also show the presence of meanders and eddies along the NCC (Mork, 1981; Audunson et al., 1981; Carstens et al., 1984; O.M. Johan-
nessen, 1986; Essen et al., 1989; J.A. Johannessen et al., 1989). These waves have typical wavelengths of about 100 km, a northward phase speed of about 0.15 m s\(^{-1}\), and a period of 5–7 days. Both theoretical and laboratory studies suggest that dynamic instabilities play a major role in the development of these waves and eddies (Mysak and Schott, 1977; Vinger et al., 1981; Ikeda et al., 1989; James, 1991). Off the mid-Norwegian northwestern coast, in the vicinity of the Halten Bank, there is evidence also that these mesoscale variabilities may be topographically induced (Eide, 1979; Audunson et al., 1981). Figure 1, taken from Eide (1979), shows the bathymetry in the vicinity of the Bank located at approximately 64°40′N, 08°30′E. West of the Bank is the continental shelfbreak adjoining the Norwegian Sea, and to the east the bank is separated from the coast by a trench the depths of which exceed 500 m in some places. There also exists other smaller banks: the Froya Bank to the southwest and the Sklinna Bank to the northeast (Fig. 1). This Halten Bank region is a site of intense field experiment in March of 1988 (Haugan et al., 1991), during which measurements of currents and horizontal maps of salinity and temperature at several depths were taken. It is also the site where a high-resolution three-dimensional simulation of currents forced by coastal discharge, North Atlantic inflow, and February/March 1988 atmospheric pressure/winds was conducted (Fig. 2: Oey and Chen, 1991). Both observation and simulation give detailed structure of the NCC meander and eddy fields, and support the notion that these are topographically induced. Both show a large current meander around the March 26, 1988, located approximately on the upstream side of the Halten Bank. Figure 3 shows the observed (Fig. 3a) and simulated (Fig. 3b) \(\sigma-t\) contours in the vicinity of the Halten Bank within the model’s

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**Fig. 1.** Bathymetry of the Halten Bank area (from Eide, 1979).
fine-grid nested region shown in Fig. 2, at \( z = -50\) m (left panel of each figure) and \( z = -100\) m (right panel), where "\( z\)" is the vertical coordinate with \( z = 0\) at the free surface. The corresponding simulated velocity vectors are shown in Fig. 3c. Figure 4 shows the bottom topography contours for the region shown in Fig. 3. The two figures together illustrate the influence of topography on the density and velocity fields. Oey and Chen (1991) also found that the simulated meander is not permanent, but rather undergoes periods of growth and decay depending on the strength of the NCC transport from the south. Thus the agreement, in time and location, of the observed and simulated meanders shown in Fig. 2 is quite remarkable, and suggests that a common underlying dynamics exists in both cases. Upon a more detailed examination of the development of the simulated meander over the two-months period, Oey and Chen (1991) further hypothesized that some resonant interaction between topography and baroclinic instability process occurs in order for the meander to grow, and that the topography which "triggers" this growth is the Froya Bank. A recent analytical study by Mitsu-dera and Grimshaw (1991) lends support to this hypothesis.

We have two objectives in this paper. First, we wish to make a detailed comparison of the simulated and observed currents at the six current

![Fig. 2. (a) The model domain with detailed topography. Arrows denote inflows/outflows across open boundaries, and coastal runoffs. The fine-grid nested region is enclosed by dash-dot lines. (b) A schematic illustration of the fine-grid/coarse-grid system used in model. Both grids use "C-grid" and are shown here to have a coarse/fine ratio of 3. The "temperature" points are denoted by "dots", and the velocity points by "+" at half grid distances shifted from the dots. For clarity, the fine-grid velocity points are not shown. The coarse-grid points extend and overlay onto the fine-grid points, and coarse-grid variables inside and on (for velocities) the double-dashlined boundary are updated with tendencies computed from averaged sums of the fine-grid tendencies. The outer-most boundary points of the fine-grid are denoted by "x" and field values at these points are interpolated from those of the coarse grid. A schematic of the Flow Relaxation Scheme (FRS) zone is shown here to span over 6 fine-grid points. In effect, those fine-grid points lying within the FRS-zones are also interpolated from the coarse-grid values.](image)
meter moorings indicated in Haugan et al. (1991), and shown here on the model's topography in Fig. 5. We wish to examine how well the model reproduces the observed currents, both for the means as well as for the sub-tidal and tidal variabilities. The comparison will also bring forth a number of issues which need to be addressed in future simulations. Secondly, we wish to test the hypothesis that a combination of topography and baroclinic instability can indeed produce mean-ders of comparable amplitudes as those observed (and simulated) over the Froya/Halten Bank region.

The paper is organized as follows. The next section describes the model, and the third section compares the simulated currents with observations. The fourth presents the results of a process-oriented simulation designed for study of a coastal current flowing over a bottom bump approximating the Froya or Halten Bank. The paper ends with a concluding summary.

The model

We first describe briefly how the nested-grid calculation was carried out. Next we discuss the boundary conditions and other model details.

The nested-grid simulation

We used a nested-grid technique (Oey and Chen, 1991) applied to the three-dimensional, primitive-equation model of Blumberg and Mellor (1983, 1987) and Oey et al. (1985). The method uses two overlapping domains as shown in Fig. 2a, a large-scale coarse-grid domain defined approximately by the 51°–76°N latitudes, and 20°W–22°E longitudes, and a fine-grid nest defined by the 61°–66°N latitudes and 2°W–12°E longitudes. Horizontal grid spacings are 20 × 20 km in the coarse grid, and 4 × 4 km in the fine grid, and there are eleven levels in the vertical scaled by the local depth (the σ-grid system) as (0, −0.05, −0.1, −0.15, −0.2, −0.25, −0.3, −0.475, −0.65, −0.825, −1.0). The simulation was first performed on the coarse grid, initialized from an iterative spin-up procedure to ensure that the open-boundary sea-level distribution was fully consistent with the currents, and from Levitus' (1982) climatological temperature and salinity fields. This coarse-grid calculation was run for 345 days, and the result was used to initialize the nested grid, high-resolution simulation. The interfacing between the two grids is sketched in Fig. 2b, which shows as an example nine fine-grid cells for each coarse-grid cell, rather than the 25:1 ratio as actually used in the simulation. During each time step of the coupled integration, the coarse-grid values are interpolated onto the nest's boundaries to force the fine-grid calculation; the tendency terms for all fine-grid rectangles which make up each coarse-grid rectangle overlapping the nest are then summed over each coarse grid cell, and stored in sub-arrays with the coarse-grid indices. In Fig. 2b, the averaging is done for all cells inside and on (for velocities) the double-dashlined boundary. The tendency terms include the Coriolis acceleration, pressure gradient, and all mass and momentum fluxes due to advection and diffusion. Because of the conservative form of the finite-differencings used in the model, these summed tendencies become equivalent to coarse-grid tendencies calculated from fluxes across the coarse-grid interfaces within the nest, except of course they now include the sub-(coarse)grid scale eddies. These tendencies are now used to replace those coarse-grid tendencies which overlap the nest as the coarse-grid solution is also advanced one time step. In this way, internal "boundary conditions" for the coarse-grid calculation outside the nest are never needed, since the coarse-grid integration is, in practice, all done in one step including those grid points which overlap the nested grids. The nesting is two-way interactive in that flow in the nest can response to non-local, large-scale forcing in the coarse grid, while coarse-grid circulation in turn can "feel" the effects of eddies from within the nest.

The coupled calculation was carried out for 300 days, during which meteorological forcing for February and March of 1988 was applied cyclically five times, with temporal "ramping" used at the start of each cycle to reduce inertial oscillation. We think that this cyclic application of the forcing is the best that one can do in the absence of data prior to February 1988, as it gives a...
best-guessed initial field for the simulation. The underlying hypothesis is that the meander and eddy fields in the NCC are primarily forced by winds, topography and coastal runoffs, and that they have "memory" (within the nest) shorter than the simulation period of two months in which the forcing was applied. The entire simulation including both the coarse-grid spin-up and coupled calculation is therefore 645 days, and it is the last 60 days (for February/March 1988) which we analyze.

Boundary conditions

In addition to the February/March 1988 meteorological forcings, the model is also forced by inflows and outflows across the open boundaries, by tides, coastal and Baltic discharges, and also relaxation to wintertime climatology for model depths > 500 m. From Swift (1986), and others, we specify a net transport across the southwestern boundary of the model-inflow of North Atlantic water minus overflow of Norwegian Sea water, at 5 Sv, and a net inflow of the Greenland Sea water northeast of Iceland at 1 Sv. An inflow of 0.09 Sv is specified across the Dover Strait (Prandle, 1984). To balance inflows, outflows are specified along the northern boundaries (see Fig. 2a).

To account for winds and tides, and to ensure that waves from the model's interior can radiate out, the above inflows and outflows across the open boundary are used in conjunction with a radiation condition of the form:

\[
U_n = U_F n + C_i (u_n)_{\text{in}} - (U' n)_{\text{out}} / (c/H) \times [\eta - \eta_F (s, t)]
\]

where \(n\) is a unit outward normal to the boundary, \(\eta\) is the free-surface elevation, \(U_n\) the depth-averaged velocity normal to the boundary, \(U\) the depth-averaged velocity vector, subscript \(F\) denotes forcing (tides or the specified transports), \(s\) measures distance along the open boundary, and \(c = (gH)^{1/2}\). Since tides and transports are specified, the \(U_F\) and \(\eta_F\) consist of the sums of each forcing separately. Only \(M_2\) tides are used in the present simulation.

A radiation condition is used for the three-dimensional velocities along the open boundary:

\[
(n \cdot u)_{\text{in}} + C_i (n \cdot u)_{\text{out}} = 0,
\]

where \(C_i\) is a constant internal phase speed equal to \((Hg \cdot 10^{-3})^{1/2}\). This value is chosen to correspond to a \(\sigma\) difference of approximately 1 to 3 \(\text{kg m}^{-3}\) across the main pycnocline depending on its depth. Our past experience with calculations using simpler model configuration suggests that the result is relatively insensitive to the exact \(C_i\) used, provided that it is not too far off from the internal phase speed corresponding to the simulated density stratification. A forward time, one-sided spatial differencing is used to solve eqn. (2).

Coastal runoffs are specified along five runoff zones (Fig. 2a): 615 \(\text{m}^3\ \text{s}^{-1}\) from the east coast of Britain, 4150 \(\text{m}^3\ \text{s}^{-1}\) from Europe continent, 1.5 \(\times 10^4\) \(\text{m}^3\ \text{s}^{-1}\) from the Baltic Sea, 1900 \(\text{m}^3\ \text{s}^{-1}\) from south of Norway and Sweden into the Skagerrak, and 1600 \(\text{m}^3\ \text{s}^{-1}\) from the Norwegian west coast. The British and European runoffs are from Taylor et al. (1983), the Baltic from Ehlin (1981), and the Scandinavian fjords from Wold (1989). At each runoff grid, a velocity profile, \(u_R(z)\) linear in \(z\) is specified:

\[
u_R (z) = u_0 + 0.1 \text{sign}(u_0) \times [1 + 2(z - \eta)/(H + \eta)]
\]

where \(u_0\) is the barotropic velocity obtained from dividing the total discharge along each of the five runoff zones by the summed width of runoff grids in that zone, and by the local depth. This expression in general gives, at each runoff grid, a surface seaward velocity a little greater (by \(|u_0|\)) than 0.1 \(\text{m s}^{-1}\), and a bottom landward velocity a little less than 0.1 \(\text{m s}^{-1}\).

For salinity and temperature, we use

\[
(S, T)_t + |n \cdot u_n| (S, T)_n = 0,
\]

where \(n\) is a unit normal along the coast. For surface seaward flow, \((S, T)_n\) is approximated by \([S, T] - (S_F, T_F)\)/\(\Delta n\), and for bottom landward flow \((S_F, T_F)\) are replaced by the model's interior \((S, T)\) values at the seaward grid next to the runoff grid. Here, \((S_F, T_F)\) are the values of \(S\) and \(T\) at discharge. We equate \(T_F\) to \(T_{cs}\), where \(T_{cs}\) is the climatological temperature at surface,
and $S_F$ to $\int S_v \cdot dz/H - \Delta S_0$, i.e., to a value $\Delta S_0$ less than the vertically averaged climatological salinity ($S_v$) value at the runoff grid. We set, rather arbitrarily, $\Delta S_0 = 3$ ppt.

The same expression (4) is also used for the temperature and salinity along the open boundary, where now the $T_F$ and $S_F$ are specified, during inflows, from Levitus' climatology.

The meteorological forcing, wind velocities and pressure, were provided to us by Dr. Martinsen and colleagues of the Norwegian Meteorological Institute (DNMI). The pressure fields were obtained from the European Centre for Medium Range Weather Forecast, and wind vectors were calculated from these pressure fields as detailed in Engedahl and Martinsen (1988). We calculated (kinematic) windstress vectors from wind velocities $U_w$ using the following formula:

$$\tau_0 = C_D |U_w| U_w$$

(5a)

where the drag coefficient $C_D$ (already multiplied by the ratio of air density to water density) is given by:

$$C_D = 1.275 \times 10^{-6}$$

$$|U_w| \leq 5 \text{ m s}^{-1},$$

$$= (0.95 + 0.065 |U_w|) \times 10^{-6}$$

$$5 \text{ ms}^{-1} < |U_w| \leq 20 \text{ m s}^{-1},$$

$$= (1.48 + 0.038 |U_w|) \times 10^{-6}$$

$$20 \text{ ms}^{-1} < |U_w| \leq 50 \text{ m s}^{-1},$$

$$= 3.4 \times 10^{-6}$$

$$50 \text{ ms}^{-1} < |U_w|$$

(5b)

This empirical formula is used to fit the $C_D$ values given in table 2 of Bunker (1976), for the case $-1^\circ C < T_{air} - T_{sea} < -0.3^\circ C$. This air/sea temperature difference is appropriate for the modelled region where heat transfer from the ocean to the air appears to be a prevalent condition (Gathman, 1986).

The total upward heat flux consists of the long-wave back radiation minus incident solar radiation, plus sensible and latent heat losses from the ocean to the atmosphere. The sensible and latent heat losses were provided to us by DNMI. The net (back-incident) radiation is taken to be a linear function of latitude, from Bunker (1976), and varies from $-75 \text{ W m}^{-2}$ at 50$^\circ$N to $-55 \text{ W m}^{-2}$ at 73$^\circ$N.

**The deep ocean density field**

For long-term simulation, the density structures in the deeper layers of the model domain may slowly deviate from climatology. This may be caused by our inadequate modeling of the diapycnal mixing process, or by errors introduced at the open boundary. To avoid this, we add an adjustment term, $-G(z)[e - e_c]$, where $e$ represents either temperature or salinity, to the prognostic equation, i.e.,

$$e_t + \ldots = -G(z)[e - e_c],$$

(6)

where "..." denote advection and diffusion terms, omitted for simplicity.

Here, $G(z)$ is a time constant given by

$$G(z) = G_d (1 - \exp[G_h - z])$$

(7)

where $G_d = 4.63 \times 10^{-8} \text{ s}^{-1}$ and $G_h = 2 \times 10^{-3} \text{ m}^{-1}$. Thus for depths less than about $G_h = 500 \text{ m}$, there is essentially no adjustment; while in the deepest layers, adjustment is made at time scales of the $O(G_d^{-1})$, or about 250 days. This adjustment method is due to Sarmiento and Bryan (1982).

**Sub-grid scale processes**

To represent sub-grid scale processes, the model utilizes the Smagorinsky diffusion formulation in which the horizontal viscosity and diffusivity coefficients, $A_M$ and $A_H$ are modeled by:

$$A_M \text{ or } A_H$$

$$= C \Delta x \Delta y \left[ u_x^2 + u_y^2 + (u_x + v_y)^2 / 2 \right]^{1/2},$$

(8)

where $C$ is a constant taken to be 0.2 in the coarse grid, and 0.05 in the fine grid. The values of $"A_{M,H}"$ in the region of the North Sea are about 50 to 300 m$^2$ s$^{-1}$, which are in the range of values used in other models with horizontal grid sizes of about 20 km (see, e.g., Davies, 1983). In the nested region, the values are about 5 to 20 m$^2$ s$^{-1}$, with an average over the nest and over the two-months period of about 10 m$^2$ s$^{-1}$. 

The different values of viscosity and diffusivity used in the coarse- and fine-grid regions are necessary. The use of a coarse-grid value for both grids would obliterate the eddy field in the nest, while fine-grid viscosity and diffusivity values would cause excessive numerical oscillations in the coarse grid. The simulation is therefore limited in that it cannot predict propagation, into the nest, of observed mesoscale features (meanders or eddies) generated in the coarse grid, since the latter in general does not resolve these features.

Comparison with observation

In Oey and Chen (1991), we made a preliminary comparison of the simulated and observed density fields, as shown for example in Fig. 3. This was the only data available to us at that time. We also compared some gross features of the salinity, temperature and currents with observations published in the literature. These comparisons indicate that the simulation was fairly successful in reproducing a number of important observed features and responses, including the strength of the NCC transport, the cross-current salinity change, and wind-induced responses in the North Sea. The published NORCSEX’88 (Norwegian Continental Shelf EXperiment, 1988) current meter data by Haugan et al. (1991) now provides a unique opportunity to more rigorously verify the simulated results in the Halten Bank region.

Detailed descriptions of NORCSEX’88 have been given in the special June 1991 issue of the Journal of Geophysical Research, Ocean. Apart from the paper by Haugan et al. (1991), there also exists a technical report by Johannessen and

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Fig. 3. (a) The observed “March 1988” $\sigma$-$t$ contours at $z = -50$ and $-100$ m, plotted in the model’s fine-grid nested region (see Fig. 2a) which includes the Froya and Halten Banks. (Data courtesy of the Marine Research Institute, Bergen). (b) Same as (a) for the simulated $\sigma$-$t$ contours on March 27, 1988. (c) Same as (a) for the simulated velocity vectors (plotted at every third grid point) on March 27, 1988.
Haugan (1991) in which a more detailed description of the observational results are given. We will compare the simulated currents at the six moorings shown in Fig. 5. The moorings consisted of four to six current meters at depths from \( z = -25 \) m to 25 m above the bottom, and for periods encompassed by March 5 through 25, 1988; the shortest record was for current meter "CMTI", which was from March 5 through 22. Before we present the detailed results, it must be pointed out that there are discrepancies in the model and real topographies, as a comparison between Figs. 1 and 5 shows. The model topography was derived from a gridded depth array of \( 20 \times 20 \) km resolution, and fine details of trenches and ridges are smoothed out. Although the discrepancies are slight, they are important in the present case in which flow is significantly controlled by topography. Another difficulty in the comparison process is that although the latitudes/longitudes of the moorings are known, some accuracies are lost when these are transferred to the model grid since the latter is on a Cartesian plane obtained from a polar stereographic map with map factor of 1 at 60°N. Ideally, an interpolative optimization scheme can be applied in which a number of grid points neighboring the observation mooring are used so that some measure of the error (the root mean square error, for example) between the observed and simulated currents is minimized. This has not been done here, and grid points nearest to the observed mooring stations are used in the comparison.

Figure 6, taken from Haugan et al. (1991), shows the observed mean currents from the six mooring locations. Currents are temporal means over the respective deployment period of each mooring, and vertically averaged over 4–6 current meters at each mooring. Note that for ease of comparison, the plot is rotated 42° anticlockwise from that shown in Haugan et al. to conform to the orientation of the model domain. These mean currents can be compared with the corresponding

Fig. 3 (continued).
simulated currents in Fig. 5. The agreement is fairly good in that both currents show intensification for the three nearshore moorings CMT1, CM1 and CM2, an offshore (fairly intense) current deflection on the upstream (i.e. south) side of the Halten Bank at mooring CMT2, and weaker onshore currents on the downstream side of the Bank at moorings CMT3 and CM3. This circulation pattern clearly indicates the strong constraints which bottom topography has on the flow. It is also clear from “snapshots” of the simulated current and density fields presented in Oey and Chen (1991), of which one example was given in Fig. 3, that the mean circulation pattern is a cumulative effect of time-dependent meanders which develop as a result of flow interaction with topography. Such a meander would grow downstream of the Froya Bank, and is modified further as it encounters the Halten Bank.

Table 1 shows observed and simulated values at all current meters. The mean and standard deviation for both the east/west and north/south component velocities, as well as the analyzed M2 tidal current amplitudes and rotations, are given. As a measure of difference between the observed and simulated values, we also give the root mean square percentage error defined for each mooring as

\[
\text{rms \% error} = 100 \times \left[ \frac{\sum (\text{Model-Observed})^2}{\sum (\text{Observed})^2} \right]^{1/2} \%
\]

where \( N \) is the number of current meters at each mooring (note that this definition has the effect of inflating the error when the observed mean square average, the denominator on the right hand side of eqn. (1), is small, as is the case for mooring CM3, for example). Another way of displaying the information in Table 1 is given in Fig. 7, in which the observed values are plotted against the simulated values for the east/west (a) and north/south (b) mean velocities, the standard deviations of east/west (c) and north/south (d) velocities, and the M2 tidal current amplitude (e).
For the mean currents, large rms % errors (> 45%) are found for the north/south velocity component at CMT1, CMT3 and CM3, and for the east/west component at CM1. At CM3 the error is inflated by the small magnitude of the currents there. The simulation gives a weak southward flow, as observed, and the vertical shear is also predicted well, of the order of $2 \times 10^{-4}$ m s$^{-1}$ (so that the error is barotropic). The region is where a cyclonic flow is predicted in the simulation (see, e.g., Fig. 3c) and a slight north-eastward shift of the position of CM3 on the model grid (as we have experimented) can actually eliminate the error (to within 5%). Such a shift is indeed suggested by the position of the CM3 on the real map (Fig. 6), as compared with that on the model map (Fig. 5), and points out the previously mentioned difficulty associated with exact positioning of the mooring on a Cartesian-based, smoothed-out model topography. Similar "position error" can be expected also for other moorings.

At CMT1 and CMT3, the simulated vertical shears of the mean north/south velocities are of opposite signs to those observed (except for the surface layer meter at CMT1, which has the same sign), and this accounts for the relatively large errors at these stations. The discrepancies are probably caused by opposite signs in the cross-isobath density gradients in the subsurface, but this cannot be readily confirmed by the highly aliased observed hydrographic maps (similar to Fig. 3a). Similar discrepancies in the baroclinic structure at CM1 may also account for the large error there, although in this case position error may also be large because of the complex underlying topography not resolved in the model in this region.

For the standard deviation (s.d.) taken here to be a measure of the eddy kinetic energy, the simulation in general underestimates the values especially at moorings CMT3, CM1 and CM3. Except for mooring CM3, for which part of the
s.d. error can again be attributed to position error, the errors near the surface are in general small, with a maximum in the rms % error equal to 25% for the north/south component at CM1. There appears to be a substantial portion of flow variability (about 50% from Fig. 7c, d) in the subsurface which is not resolved in the simulation. The source of discrepancy is most likely because of insufficient vertical resolution used in the simulation, especially in the subsurface where coarser σ grid spacings are used. A high vertical resolution is necessary in the NCC, in which strong stratification (both in the horizontal and vertical) is expected and baroclinic instability process is important.

For the M_2 currents, the rotation directions in general agree with those observed: clockwise upstream (CMT1 and CMT2) and anticlockwise downstream (CM3) of the Halten Bank. At CMT3, the near-bottom rotation is simulated correctly (anticlockwise) but those in the near surface are clockwise instead of anticlockwise. How-

Fig. 6. The depth and temporal means of currents at the six mooring locations during the NORCSEX'88 (from Haugan et al., 1991).
ever, at a few grid points further downstream and shoreward, the simulated near-surface rotation changes to anticlockwise direction. This is illustrated in Fig. 8, in which the M\(_2\) ellipses at the first \(\sigma\) grid point near the surface are plotted. A similar change from clockwise to anticlockwise rotation at grid points further downstream of the mooring CM3 in the near-surface can also be seen in Fig. 8. For the M\(_2\) current amplitudes, largest discrepancies are found at moorings CMT1 and CM1, at CMT2 in the subsurface, and at CM2 in the near-surface. Errors near the surface at CMT3 and CM1 are also large. Seven of the ten points lying well above the 45° line in Fig. 7e are from CMT1 and CM1, one is from CMT2 and two are from CM2. The errors at CMT1 and CMT2 are most likely caused by the large topographic change at the shelfbreak, across which large variation in the tidal amplitude occurs (see Fig. 8) and some accuracy is lost due to insufficient vertical resolution. At CMT3, the rather large simulated amplitude near the surface ap-

### TABLE I
Model/observed currents comparison at the six NORCSEX'88 moorings

<table>
<thead>
<tr>
<th>Mooring</th>
<th>Depth (m)</th>
<th>East velocity (cm s(^{-1}))</th>
<th>North velocity (cm s(^{-1}))</th>
<th>M(_2) Amp.* (cm(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>CMT1</td>
<td>25</td>
<td>18.89</td>
<td>13.70</td>
<td>11.02</td>
</tr>
<tr>
<td></td>
<td>80</td>
<td>16.27</td>
<td>14.63</td>
<td>8.06</td>
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<td></td>
<td>150</td>
<td>11.96</td>
<td>14.71</td>
<td>7.73</td>
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<td></td>
<td>175</td>
<td>10.38</td>
<td>11.27</td>
<td>7.21</td>
</tr>
<tr>
<td></td>
<td>rms % error:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CMT2</td>
<td>25</td>
<td>2.83</td>
<td>2.98</td>
<td>14.34</td>
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<td></td>
<td>80</td>
<td>2.20</td>
<td>3.80</td>
<td>11.07</td>
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<tr>
<td></td>
<td>rms % error:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CMT3</td>
<td>25</td>
<td>6.95</td>
<td>5.47</td>
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* The "c" and "a" denote clockwise and anticlockwise rotation directions, respectively, of the tidal current ellipses.
pears to be due to position error, since a few grid points further downstream the amplitude decreases to approximately 5 cm s\(^{-1}\), closer to that observed. Errors at CM1 and CM2 are most likely because of discrepancies in bottom topography, mentioned previously.

The relatively large discrepancies in the \(M_2\) current amplitudes are in contrast to comparison study in the North Sea where Davies and Furnes (1980) found better agreements between the modeled and observed \(M_2\) currents. There are two reasons for the larger discrepancy. First, tidal currents along the Norwegian coast are much weaker (about ten times weaker) than those found in the North Sea, and relative errors are therefore large even if the observed and modeled currents agree to within a few cm s\(^{-1}\). Secondly, the topographic gradients along the Norwegian

![Graphs showing observed vs. simulated data](image)

Fig. 7. Plots of observed vs. simulated for (a) the east/west mean velocity, (b) the north/south mean velocity, (c) east/west standard deviation velocity, (d) north/south standard deviation velocity and (e) tidal current amplitude.
coast are in many places large because of the existence of trenches and ridges, and because of the close proximity of the shelfbreak. This requires very high model resolution both in the horizontal and vertical for accurate tidal current prediction. Indeed, the comparison results of Davies and Furnes (1980) for stations along the southwestern Norwegian coast (south of 62°N) also show relatively large discrepancies [errors of 0(25%) and at a few stations greater than 100%], similar to what we have found here. However, since tidal currents are weak, the discrepancies are not expected to significantly affect the subtidal circulation induced by winds and NCC meanders and eddies.

To see how well subtidal events are accounted for in the model, we compare in Fig. 9 (a) simulated and (b) observed (taken from fig. 5 of Haugan et al., 1991) velocity stick plots from all moorings at $z = -25$ m. Since only the $M_2$ tidal component was used in the simulation, this was removed by averaging the model currents over two tidal cycles. As can be expected from Table 1, the best agreements between model and observation are at moorings CMT1, CMT2 and CM2, where the simulation reproduces events indicated by the "closing" and "opening" fans of the vector sticks. Assuming a generally northeastward flow, "open fan" structure reflects passage of a cyclonic eddy, and "closed fan" an anticyclone (Haugan et al., 1991). While there are differences between the model and observation in the intensity of these events, the general pattern agrees fairly well. At moorings CMT3, CM1 and CM3, the agreements are poor, as had also been noted previously in conjunction with Table 1. Similar conclusions are also obtained when subsurface current meters are examined (not shown).
A process study of bank-induced meanders

Both the simulation and observation (see, e.g., Fig. 3) suggest that meanders upstream of the Halten Bank are produced as a result of amplification of bank-induced eddies through dynamic (predominantly baroclinic) instability process. This was hypothesized in Oey and Chen (1991) and an essential argument in support of it is that in the absence of a cross-shelf density gradient, i.e., in the absence of a baroclinic coastal current, no such meanders would be found (Huppert and Bryan, 1976; Boyer and Zhang, 1990). We proceed here to conduct model experiments designed to isolate the relevant flow physics. We use the same primitive equation model as that used for the simulation. The differences are in the model geometry and boundary conditions as described in the followings.

Model setup

The model domain is a channel on an f-plane \((f = 1.3 \times 10^{-4} \, \text{s}^{-1})\), 200 m deep, 94 km wide (east/west) and 306 km long (north/south). The channel is open at its northern and southern ends. Along the northern end, an Orlanski’s (1976) type of radiation condition similar to eq. (2) is used, in which the corresponding phase speeds (both barotropic and baroclinic) are set constant. The barotropic phase speed is set to \((gH)^{1/2}\), where \(H = 200\) m, and the baroclinic phase speed is \((H_1 g \Delta \rho / \rho)^{1/2}\), where \(H_1\) is a measure of the upper-layer depth (Fig. 10), and

Fig. 9. A comparison between the (a) simulated and (b) observed (from Haugan et al., 1991) current vector stick plots at the six NORCSEX’88 moorings, at \(z = -25\) m.
\[ \Delta \rho/\rho \] is taken to be \( 2.4 \times 10^{-3} \) corresponding to a salinity difference of about 3 ppt between the upper and bottom layer waters (see below). The value of \( H_1 \) is chosen to be 85 m, corresponding approximately to the depth of the upper layer after the geostrophic adjustment of the flow, but other values ranging from 50 to 100 m were tested, with little changes in the solutions. The solutions are also insensitive when Orlanski’s formula for computing the phase speeds from the upstream values was used. The radiation condition is applied to all the prognostic variables except the salinity, for which an advection scheme given by eq. (4) is used.

At the southern boundary, a baroclinic inflow superimposed on a barotropic transport of magnitude \( 9 \times 10^5 \) m\(^3\) s\(^{-1}\) is specified. The baroclinic

--

Fig. 9 (continued).

Fig. 10. A cross-channel/vertical section view of the initial salinity distribution \((S_1 = 32 \text{ ppt}, S_2 = 35 \text{ ppt})\) used in the process-oriented model experiments. This distribution is also used for the inflow boundary. The plot also gives the bottom bump (or seamount) used, shown here at a section which cuts across its maximum height.
inflow is in geostrophic balance with a cross-shelf salinity front as shown in Fig. 10, in which $S_1 = 32$ ppt is linearly changed to $S_2 = 35$ ppt further to the west. The front is located at approximately 30 km from the eastern wall (the coast) and maximum depth of the less-saline upper layer is 100 m.

Experiment I: Salinity at $z = -15$ m, $t = 1$-5 days

Experiment I: Salinity at $z = -15$ m, $t = 6$-10 days

Fig. 11. Experiment 1: salinity contours (CI = 0.4 ppt) at $z = -15$ m for $t = 1$ through 10 days at 1 day interval.
located at the coast. These values for the transport and cross-frontal salinity change are fairly typical of those for the NCC. Previous studies have shown that baroclinic instability plays an important role in the development of meanders and eddies (see, e.g., Mysak and Schott, 1977; Mork, 1981; Vinger et al., 1981). A flat bottom is therefore used in the present study to maximize the baroclinic response. The baroclinic Rossby radius of deformation, \( R_0 \), is estimated to be about 11 km. Grid spacings in the \( x \) and \( y \) directions are therefore chosen to be 2 km and there are ten, evenly distributed \( \sigma \) levels in the vertical.

No-slip and zero salt-flux conditions are used along the side walls. Zero salt-flux condition is used at the bottom and the free-surface. The model solves only the salinity equation and temperature is set constant at 10°C. Momentum flux at the surface is zero (no windstress) and at the bottom it is balanced by quadratic stress computed using the velocity nearest the bottom as detailed in Oey et al. (1985). The model uses the Smagorinsky type of horizontal diffusivity and viscosity, eq. (8), with \( C = 0.05 \) and a resulting maximum coefficient of about 5 m² s⁻¹. Three experiments were conducted: (1) flat-bottom experiment, (2) bump experiment and (3) Huppert-Bryan experiment. In experiment 2 a Gaussian bottom bump centered at approximately 27 km from the coast and 150 km from the inflow boundary (i.e. at the mid \( y \)-position) is specified:

\[
\begin{align*}
h = h_{\text{max}} \exp\left\{ - \left[ \left( x - x_0 \right)^2 + \left( y - y_0 \right)^2 \right]/R^2 \right\}
\end{align*}
\]

where \( h_{\text{max}} = 50 \) m, \( x_0 = 68 \) km, \( y_0 = 150 \) km and \( R = 10 \) km (note that \( x = 0 \) is at the western wall). A section view of this depth distribution is plotted in Fig. 10. and plan-view contours will be shown below when results from experiment 2 are presented. Except for the presence of the bump, all other aspects of experiment 2 are identical to those of experiment 1. In experiment 3, we essentially repeated Huppert and Bryan’s (1976) experiment (except of course that we now have coastal walls and inflow/outflow conditions) in that only vertical stratification is used (i.e. barotropic transport but no coastal front). This was done merely to affirm that no instability occurs and

![Fig. 12. Experiment 1. stream function contours (CI = 0.1 Sv) for \( t = 6 \) through 10 days at 1 day interval.](image)
that the results are similar to those found by Huppert and Bryan: formation of anticyclone over the bump and shed cyclone downstream of the bump. We will omit showing the results of this experiment.

Model results

There are two types of meander development which we wish to study: (i) the effect of the bump on a pre-existing, propagating meander; and (ii) how finite-amplitude meanders can be produced by flow interaction with the bump. Previous studies by Chao and Kao (1987) and Oey (1988) indicate that an initially quiescent front (e.g., the salinity front in Fig. 10 with initial velocities set to zero) along the entire length of the channel will eventually develop waves and meanders, first of short wavelength and then of long wavelength typical of that produced by baroclinic instability. The front is perturbed in this case by small-scale ageostrophic flow adjustment through vertical mixing (Chao and Kao, 1987), and waves and meanders would still develop even if the initial front is in geostrophic balance with the velocity field. However, explicit perturbations introduced initially hasten the development of the meander (Oey, 1988). In view of the results of these previous studies, we initialize the calculation with the salinity distribution shown in Fig. 10 specified along the entire length of the channel, while velocity components are set to zero. However, the along-front velocity at the southern boundary is specified to be in geostrophic balance with the salinity field (see previous section). The purpose is to initiate a perturbation near the southern boundary where the flow is forced to adjust to the interior as it crosses the boundary, and to generate the desired northward propagating meander. In the presence of a bump located in the mid-channel, one can track how the meander propagates past the bump. Moreover, perturbation will also be generated over the bump and its history can be studied.

Fig. 13. Experiment 1: velocity vectors (plotted at every third grid point in x-direction, and every fourth grid point in the y-direction) at z = -15 m for t = 6 through 10 days at 1 day interval.
Fig. 14. Experiment 2: salinity contours (CI = 0.4 ppt) at $z = -15$ m for $t = 1$ through 10 days at 1 day interval. The depth contours over the bump are 190 (outer most contour), 175 and 160 m.
Results from experiment 1

Figure 11 shows contours of salinity at $z = -15$ m for experiment 1 from $t = 1$ through 10 days at 1 day interval, and Figs. 12 and 13 the stream function and velocity vectors at $z = -15$ m from $t = 6$ through 10 days, also at 1 day interval. An examination of the detailed time series (omitted) indicates a flow adjustment phase which is nearly complete with the development of an along-front flow by $t \approx 5$ days. The initial perturbation at the southern boundary can be seen to amplify as it propagates downstream through dynamic instability. A series of backward breaking waves are formed, similar to those found in the laboratory experiments of Vinger et al. (1981) and Griffiths and Linden (1982), as well as in other numerical studies (see, e.g., Chao and Kao, 1987; James, 1987; Wood, 1988; Oey, 1988). The wavelength of the resulting meander at its matured state ($t \geq 7$ days) is approximately 70 to 80 km, or $2\pi R_0$, so that the instability is likely to be baroclinic (see, e.g., Griffiths and Linden, 1982). The backward breaking is sufficiently intense that a strong cyclone develops on the seaward side of the front, followed by a somewhat weaker anticyclone downstream, and the pattern repeats downstream. Wave-breaking results in a northward flow at the outer (i.e. western-most) rim of the cyclone. This is confined to the near-surface layers (Fig. 13), however, and does not actually join the main flow in the south, although the surface salinity contours may suggest such flow continuity (see, in particular, contours at $t = 9$ and 10 in Fig. 11).

Results from experiment 2

Figures 14, 15 and 16 show the salinity, stream function and velocity vector plots, respectively, for experiment 2, and can be compared with the corresponding plots for experiment 1 shown previously in Figs. 11, 12 and 13. Three topography contours over the bump are also shown: at 190 m (the outer-most contour), 175 m and 160 m (the shallowest depth over the bump's center is 150 m). The plots show that in addition to meanders formed by perturbation at the southern bound-

![Experiment 2: Stream Function, t= 6-10 days](image-url)
ary, there are now a series of three meanders downstream of the bump, denoted in Fig. 14 by A, B and C. The meander A exits the northern boundary at $t = 6$ days before it has a chance to mature (see Fig. 14 at $t = 6$ days). The perturbations B and C develop into finite-amplitude meanders downstream of the bump. The meander C, in particular, initiates around $t = 3$ days near the end of geostrophic adjustment, when a small cyclone is shed downstream of the bump and flow over the bump is anticyclonic (compare Figs. 11 and 14 at $t = 3$ to 5 days for experiments 1 and 2). The anticyclonic turn over the bump, and the shedding of a cyclone downstream of the bump, have also been noted in the prototype simulation over the Froya Bank (Oey and Chen, 1991; Fig. 3b). In the latter, however, the Halten Bank downstream of the cyclone clearly has a non-negligible effect on the subsequent meander development. Nevertheless, the present study shows that over a bump of width and height similar to those found for the Froya Bank, a coastal current with significant cross-shore as well as vertical density gradients can develop instabilities that evolve into finite-amplitude meanders downstream. This conclusion is different from that obtained with the Huppert-Bryan experiment, for which no meander amplification is found.

Since experiments 1 and 2 are identical except for the bump in “2”, we can assess the effects of the bump by subtracting the results of “1” from those of “2” for all water grid points shared by both experiments (i.e., grid points above and outside the bump). The results are shown here in Fig. 17 for the salinity difference at $z = -15$ m (note that for $t = 1$ to 5 days, contour interval is 0.2 ppt, while for $t = 5$ to 10 days, it is 0.4 ppt). Stream function and vector differences show the same general features and are not included here. The amplification of cyclones, where the difference is positive, (solid contours) and anticyclones (dashed contours) downstream of the bump are clearly seen in this figure.

Experiment 2 also shows effects of the bump on meanders which propagate over the bump from some upstream location. The effects are
Exp. 2 - Exp. 1: Salinity at z=-15m, t=1-5 days (CI= 0.2ppt)

Fig. 17. Contours of the salinity difference between experiment 2 and experiment 1 ("exp 2" - "exp 1"; CI = 0.2 ppt for t = 1 to 5 days, and = 0.4 ppt for t = 6 to 10 days) at z = -15 m for t = 1 through 10 days at 1 day interval. The depth contours over the bump are 190 (outer most contour), 175 and 160 m.
illustrated here by comparing the stream function plots for experiments 1 and 2, in Figs. 12 and 15, respectively. The meander of interest here is the crest (western-most extension of a meander) marked as "M" in Figs. 12 and 15. For experiment 1, the crest amplifies through dynamic instability and at $t = 8$ days a total of 0.5 Sv constitutes the meander. For experiment 2, an additional anticyclonic vortex is induced over the bump and at $t = 8$ days a total of 0.7 Sv constitutes the meander. The crest also extends some 20 km farther to the west than that for experiment 1. For $t > 8$ days, there is significant interaction of the crest "M" with the crest upstream of it, and a clear identification of its amplitude is not possible. On the other hand, while a propagating crest is further amplified over the bump, a trough with cyclonic vorticity is damped. This can be seen by comparing Figs. 12 and 15 for the trough upstream of "M" from $t = 6$ through 10 days (in particular $t = 7$ and 8 days). Amplification of the cyclone inside the trough in experiment 2 is seen to be less (by about 0.1 Sv) than that for experiment 1.

**Summary**

We conduct here a model/observation comparison of currents at six moorings deployed during March 1988 off the mid-Norwegian coast in the vicinity of the Halten Bank. The simulation was initialized from a set of fairly arbitrary initial fields which ensure only the reproduction of a realistic climatology. In other words, the initial field was not derived from the observed March 1988 hydrography, nor was the NCC current meander inferred from, say, satellite imagery. The simulation was fully prognostic except for water depths greater than 500 m where a nudging (with time scale of 250 days) of the density field toward Levitus' climatology was used. The simulation assumes, therefore, that the NCC and its meanders are deterministic in the sense that they are predominantly controlled by (1) large-scale mixing of the warm and saline North Atlantic inflow and less-saline, generally cooler (during winter) coastal discharge; (2) topography and (3) atmospheric forcings. To the extent that this assumption is valid, the agreements between the simulated and observed currents are quite good. The simulation reproduces the means fairly well; the rms errors are less than 34% in five of the six moorings for the east/west velocity component, and are less than 32% in three moorings for the north/south component. The agreements in current variabilities are good only at three moorings, for which an averaged rms error of about 22% was obtained. At the other three moorings, the largest errors for the variabilities occur in the subsurface, where energetics are underestimated by as much as 60% or more in the simulation. The discrepancies are probably caused by insufficient vertical resolution, as well as by the rather smooth topography used in the model.

The simulation also suggests that the NCC meanders during the NORCESEX'88 are produced as a result of amplification by dynamic instability of waves and eddies shed off by flow interaction with the bottom topography. To further examine the process, a channel model is used to calculate the meanders and eddies which result when a coastal front interacts with an underlying bump. The bump generates waves which amplify as they propagate downstream. The bump also amplifies a pre-existing anticyclonic meander crest which has propagated over the bump from upstream, and damps a cyclonic trough. The process may be related to the theory of resonant interaction between a baroclinic current and localized topography given by Mitsudera and Grimshaw (1991), but this is as yet to be investigated with more sophisticated analyses.

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References


