Regional simulations of the Faroe Bank Channel overflow in a level model

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Abstract

The work presented in this paper is part of an effort to understand and improve the representation of overflows in large scale, coarse resolution ocean climate models. To this end we developed a regional model of the Faroe Bank Channel overflow using the MITgcm (Massachusetts Institute of Technology General Circulation Model), a typical global ocean model using discrete levels as the vertical co-ordinate. In order to isolate the numerical diffusion resulting from the advection of tracers, the model is run without any turbulence closure schemes, without convective adjustment or any other physically based parameterization of mixing. Comparison between the model results and recent observations of the Faroe Bank Channel plume allows assessment of the model performance, including its ability to correctly represent the mixing and the downslope transport in the plume. It is found that at the highest resolution used in this paper (2.5 km – horizontal and 25 m – vertical) the structure of the modeled plume and the magnitude of the entrainment is comparable to the observed plume.

The dependence of the mixing on various model parameters, such as vertical and horizontal resolution, vertical viscosity, drag coefficient and inflow forcing, is tested extensively. The numerical mixing in the model is found to be most sensitive to changes in the horizontal resolution, and to a lesser extent on vertical resolution and vertical viscosity. The inflow forcing and drag coefficient show only a very minor effect on the mixing.

The results presented in the paper identify the shortcomings of the model at coarser resolutions which need to be addressed when attempting to represent such overflows realistically in large scale climate and ocean models.

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1. Introduction

With the increasing effort to understand the intricacies of the Earth’s climate system, in particular its sensitivity and variability, there is also an increase in the effort to improve the models used to simulate...
the climate. A particular shortcoming of $z$-level climate models, which has received considerable attention in recent years, is the representation of dense overflows and their associated entrainment and mixing processes. Numerous modeling studies (Roberts et al., 1996; Roberts and Wood, 1997; Bernard et al., 1997; Griffies et al., 2000) of the nordic overflows (Denmark Strait overflow and the Faroe Bank Channel overflow) have clearly illustrated the problems associated with overflows. The way these overflows are represented in the models has far reaching effects on the basin scale circulation, especially in the North Atlantic (Roberts et al., 1996; Marsh et al., 1996; Bernard et al., 1997). Specifically, $z$-level models at coarse resolution have difficulty in moving dense fluid down the slope without generating excessive spurious mixing.

To correct the representation of overflow entrainment in coarse resolution $z$-level models a two-stage process is necessary. First, spurious mixing of the plume as it moves down the slope needs to be eliminated. To this end crude ‘plumbing’ models have been proposed (Beckmann and Doscher, 1997; Campin and Goose, 1999), connecting grid-cells above and below the topographic step, as well as full bottom-boundary layer models (Killworth and Edwards, 1999; Song and Chao, 1999) which resemble a terrain-following coordinate region appended to the $z$-coordinate model. Second, a parameterization of small-scale processes can be used to add the correct amount of entrainment back into the model. An example is Killworth and Edwards (1999), which employs a frictional bottom boundary layer model to determine the entrainment.

Another parametrization developed by Price and Yang (1998) avoids the need to move dense water down the slope by extracting it out of the domain at the source and injecting a mixed product water back into the domain downstream of the source. This is referred to as the Marginal Sea Boundary Condition (MSBC) and is currently being implemented in the CCSM (Community Climate System Model) at NCAR (Gokhan Danbasolgu and Wanli Wu, pers. com.) for the Nordic Sea overflows in an idealized setup. It has also been implemented into HYCOM (Hybrid Coordinate Ocean Model) by Halliwell et al. (2006).

This paper forms part of the effort of the Climate Process Team on Gravity Current Entrainment (CPT-GCE) to further our understanding of the short comings of current climate models when representing mixing in overflows. This understanding will help us to develop improved representations of the overflows and associated physical processes such as mixing. As part of the process of evaluating models, comparisons between regional simulations and observations are very useful, allowing the capabilities of the model to be tested in realistic topographic conditions and flow regimes, yet isolating the overflow from other aspects of the general circulation. To this end a regional model of the Faroe Bank Channel (FBC) overflow has been developed using the MITgcm (Massachusetts Institute of Technology General Circulation Model).

The MITgcm is a $z$-level model, like most of the commonly used models for climate simulations. The problems and shortcomings of the MITgcm in representing overflows are thus likely to transfer to other models of this kind. A range of idealized tests (Legg et al., 2006) have been run using the MITgcm and from these we have a basic understanding of how the model behaves. We are therefore motivated to examine the behavior of the model when representing a realistic overflow.

The FBC was chosen as a case study because of the availability of a new set of extensive observations in the channel system around the Faroe Islands presented by Mauritzen et al. (2005). These allow us to make extensive comparison between the model and the observations. Furthermore, aside from the idealized modeling study by Ezer (2006), using a sigma-coordinate ocean model, little modeling work has been done to shed light on the dynamics of the overflow. In contrast to Ezer (2006) we use realistic topography and stratification for the Faroe region.

The FBC overflow is one of the most important nordic overflows along with the Denmark Strait overflow and thus a source of significant amounts of deep waters in the region. Since Cooper (1955) linked the dynamics of the FBC overflow to the basin wide circulation of the North Atlantic Ocean and the climate in the region, it has attracted much attention in the 1980s, 1990s and 2000s. The most recent multinational observational campaigns in the region, include NANSSEN (Borenäs and Lundberg, 1988; Saunders, 1990) and VEINS (Hansen et al., 2001). In addition many individual cruises have been undertaken in recent years, such as RRS Discovery 242 in the fall of 1999 (Duncan et al., 2003), RRS Discovery 247 in the Summer of 2000 (Mauritzen et al., 2005; Prater and Rossby, 2005; Geyer et al., 2006), amongst many others. A comprehensive review of the North Atlantic and Nordic Sea Exchanges, including the FBC overflow, has been presented by Hansen and Østerhus (2000).

The overflow through the FBC is fed mainly from dense waters in the Norwegian Sea. The Norwegian Sea Deep Water (NSDW) and the Norwegian Sea Arctic Intermediate Water (NSAIW) enter the Faroe Shetland
Channel (FSC), south east of the Faroe Islands; inside the FSC these deep waters are now referred to as Faroe Shetland Channel Bottom Water (FSCBW). The channel system turns a sharp 90° angle entering the Wyville Thomson Basin south of the Islands. The deep waters are steered along this channel before they reach the FBC. Approximately 1.9 Sv of dense water overflows through the FBC as seen in a 5-year average, by Hansen and Østerhus (2000), while roughly 0.3 Sv overflow across the Wyville Thomson Ridge (WTR).

Mauritzen et al. (2005) observe gradual changes in the characteristics of the deep water masses on the way from the Norwegian Seas toward the FBC. Hence, by the time the Norwegian Sea Waters reach the channel they have already changed somewhat due to mixing in the FSC system.

At the sill the dense water is confined to a 15 km wide channel. The plume is observed to be on average 250 m thick and maximum velocities of up to 1 m/s are measured (Hansen and Østerhus, 2000).

Once the plume exits the channel it remains confined by topography for approximately a further 50 km, before the channel opens up to the Faroe Iceland slope. The plume then proceeds to spread across the slope and becomes a geostrophically adjusted current following the bathymetry.

Mauritzen et al. (2005) and previous studies find that most of the mixing happens within the first 100 km downstream of the sill. Here the majority of the water mass transformation takes place. By the time the plume has moved 150 km downstream it has become approximately 100 km wide and the transport in the plume has moved from dense water mass classes to distinctly lighter ones. None of the initial dense overflow water is present any longer and the hydrographic properties of the plume have changed significantly. 'The hydrographic properties are, for all intensive purposes, set within 100–150 km of the sill’ (Mauritzen et al., 2005). No further significant changes of the water mass occur between here and the Reykjanes Ridge.

The features at the sill as well as in the plume downstream of the sill are distinctly time variable (Mauritzen et al., 2005) and evidence of regular oscillations in the model will be discussed in the following sections of the paper.

The effect of tides on vertical mixing and bottom friction in overflows is increasingly believed to be of significant importance. They tend to generate large internal tides and non-linear internal waves in shelf regions where overflows occur. However, a study to investigate the importance of tides on the FBC overflow is beyond the scope of this paper and we will thus ignore them in this study.

Our goal here is to examine results from a series of regional simulations of the Faroe Bank Channel overflow with the MITgcm, and assess the ability of the model to reproduce the features of the observations and subsequently how the representation changes with a changing model setup. The outline of the paper is as follows: in Section 2 the setup, forcing and initial conditions, of the regional model are outlined. In Section 3.1, the results of the first control run are discussed in detail, considering the thickness and spread of the plume as well as its transports and time variability. In Section 3.2, the prominent changes in the plume’s characteristics at coarser horizontal resolutions will be discussed. In Section 4, a suite of sensitivity runs will be presented, investigating changes in the plume’s path, density changes and entrainment as a function of parameters such as vertical viscosity, $\nu_Z$, and the drag coefficient, $C_d$, as well as horizontal, $\Delta x$, and vertical, $\Delta z$, resolution and varying inflow forcing, $Q$. Finally in Section 5, conclusions will be drawn.

2. The model setup

For the study presented here we employ the MITgcm (Marshall et al., 1997a,b) in its hydrostatic mode with horizontal resolution $\Delta x = [2, 10, 20, 50]$ km and vertical resolution $\Delta z = [25, 50, 100]$ m. The model uses a partial step topography (Adcroft et al., 1997), allowing for a representation of the topography that does not restrict it to the depth of the levels. This implementation of topography has already made a significant improvement to the structural representation of overflows; however the spurious mixing resulting from the step topography remains a problem to be solved.

The transport of dense water down the slope by the overflow is represented relatively well if both the vertical and horizontal resolutions are fine enough to satisfy the following conditions:

$$\Delta z < h \quad \text{and} \quad \Delta x < h/\tan \alpha,$$

where $h$ is the thickness of the overflow and $\alpha$ is the topographic slope angle in the vicinity of the overflow (Winton et al., 1998). In the case of the Faroe Bank Channel, $h \sim 200$ m and $\alpha \sim 13 \times 10^{-3}$, hence the vertical
resolution $\Delta z$ has to be less than 200 m and the horizontal resolution $\Delta x$ should be less than about 15 km. We therefore expect to achieve a good representation of the plume when using 2 km and 10 km resolutions in the horizontal but not at the coarser resolutions; in terms of the vertical resolution all experiments lie within the limits of the criterion.

The advection scheme of the model employs a direct discretization method with flux limiting for tracers (Pietrazk, 1998), preventing the appearance of spurious oscillations in the tracer field and introducing diffusivity where needed for stability. The numerical diffusion thus introduced is not predictable and it is not possible to have an idea of its magnitude before actually running a specific experiment. It is this diffusion that we aim to quantify with the present study. Consequently the explicit diffusivity is set to zero. The current setup does not employ an explicit mixing scheme, such as convective adjustment or KPP (KPP is a mixing scheme developed by Large et al. (1994) to account for vertical mixing in the ocean). We emphasize that throughout this paper when we refer to mixing and entrainment in the model results, this is numerical mixing resulting from the advection scheme.

Legg et al. (2006) show a comparison between the mixing in a non-hydrostatic high resolution model run, in which the physical processes responsible for the mixing are partially resolved, and hydrostatic runs with a resolution of $\Delta x = 2.5$ km and $\Delta z = 30$ m. They find that the mixing is actually less in the hydrostatic 2.5 km resolution case compared to the non-hydrostatic case, suggesting that the numerical mixing at 2.5 km resolution underestimates the ‘true’ mixing. We might thus expect the numerical mixing in the experiments with $\Delta x = 2$ km to be comparable to the real mixing deduced from observations.

The model uses a quadratic drag which is set to $C_d = 2 \times 10^{-3}$ in most experiments. At fine resolution (2 km and 10 km) the horizontal viscosity is calculated using the Leith scheme for geostrophic turbulence (Leith, 1968, 1996) and it is thus variable. At the coarser horizontal resolutions (20 km and 50 km) the horizontal viscosity is set to a constant value, $A_h = 2.0 \times 10^3$ m$^2$/s, because the dominant geostrophic eddy scale is not resolved at these resolutions. In order for the Leith scheme to apply the grid-size needs to resolve the eddy scale. The vertical viscosity is $A_z = 2.0 \times 10^{-4}$ m$^2$/s in the control run, but is varied in some of the sensitivity runs.

The domain of the region is 750 km × 500 km in the East-West and North-South directions, respectively and a smoothed and gridded Smith and Sandwell (1997) bathymetry is used. The topography is shown in Fig. 1. Note that it is rotated anticlockwise by 45° with respect to real north; we will however refer to the boundaries using the cardinal directions, north, south, east and west, for simplicity. The Icelandic shelf lies at the western boundary of the domain, the Faroe Islands lie at the northern boundary and in the east the domain is bordered by the Scottish shelf. The domain is closed on all sides apart from a 170 km wide gap at the eastern end of the northern boundary where the opening of the Faroe Shetland Channel borders the domain. Here the forcing of the model is applied.

2.1. Initial conditions and inflow forcing

Fig. 2 illustrates the forcing conditions of the control run at the northern opening of the domain. In panel (a) the forcing velocity at the opening is plotted. At the bottom an inflow with uniform velocity $-0.0341$ m/s enters the domain (black) and at the surface an area of outflow (gray) is prescribed at a velocity $0.0433$ m/s that balances the inflow at the bottom in terms of volume transport. The forcing conditions of the sensitivity runs at this boundary are listed in Table 1. The in and outflow velocities are adjusted from run to run since the cross-sectional area of the forcing region changes somewhat between the different resolutions in the vertical and horizontal.

In panel (b) the boundary conditions of density are shown. Dense water (light gray) flows into the Faroe Shetland channel up to 600 m in all experiments, above that the ambient stratification applies. Most runs are setup with an inflow of 2.6 Sv at the northern boundary. The observations (Mauritzen et al., 2005) suggest an average transport of 2.9 Sv at section A, which is located about half way along the Faroe Shetland Channel. Some entrainment will have already taken place between section A and the northern boundary at which we prescribe the inflow and thus a slightly lower inflow transport was chosen. However we did also carry out one experiment with an inflow of 3.0 Sv. The density of the inflow: $\sigma_{oimf} = 28.07$ kg/m$^3$, is also taken from the observations published by Mauritzen et al. (2005).
Fig. 1. Topography of the domain with depth contours (m). Contour intervals (c.i.) are 100 m – thin contours and 500 m – thick contours. The main observational sections (B–H) by Mauritzen et al. (2005) are marked with stars and labeled with italic letters. The regions A and B, enclosed by dashed lines, are used to derive bulk entrainment estimates. The topographic abbreviations are as follows: FBC – Faroe Bank Channel, FSC – Faroe Shetland Channel, WTR – Wyville Thomson Ridge, and FIR – Faroe Iceland Ridge.

Fig. 2. Forcing and initial conditions of the control run. (a) Velocities at the northern boundary are shown: outflow in gray, \( v_{\text{out}} = 0.0433 \) m/s, and inflow in black, \( v_{\text{in}} = -0.0341 \) m/s. (b) Density conditions at the boundary: in light gray the level of the inflowing dense water in the FSC and above in dark gray the mildly stratified background. (c) Density profile at station 187 of section G. The dashed line is the linear fit to the waters between the surface mixed layer and the bottom plume used as a guide for the background stratification in the model. The horizontal lines mark the transition points between the ambient water masses, the interfacial water mass and the dense overflow water. The interfacial layer is denoted by \( \Delta h \) and the homogenous core of the plume by \( h_c \), the sum is the thickness of the plume: \( h = \Delta h + h_c \).
Panel (c) illustrates how the background stratification, set in the remainder of the domain, is derived. The density profile at station 187 in section G is shown. The top 200 m are characterized by the light mixed layer and the bottom 200 m by the dense plume. We chose the initial background stratification (dashed line) to be a linear fit to the profile in the intermediate waters between the surface mixed layer and the bottom plume. This gives a linear profile with $\rho_0 = 27.41 \text{ kg/m}^3$ at the surface and $\rho_{2000} = 27.58 \text{ kg/m}^3$ at the bottom.

The stratification in the model is prescribed by a linear equation of state. A passive tracer $\tau$ with a value of 1 is used to mark the overflow water, while ambient water masses have $\tau = 0$.

Initially the Faroe Shetland Channel system is filled with dense water ($\rho_{\text{ovf}} = 28.07 \text{ kg/m}^3$) from the northern boundary all the way to the FBC. In the high resolution experiments the level reaches up to 825 m. The sill depth is between 770 m and 800 m at all resolutions and so there is no dam break initially at the high resolution. In the coarser resolution cases the initial level of dense water reaches up to 700 m and thus the initial flow through the channel starts as a dam break. The difference in the setup at coarser resolution was necessary to ensure the numerical stability of the model at the northern boundary where the inflow is prescribed. There is no velocity anywhere in the domain initially, but as the model is integrated in time a flow is set up as a result of the continuous supply of dense water through the northern boundary and the outflow of ambient water at the surface.

Other setup specifics of the model include: non-slip boundary conditions; vertical levels: $\Delta z = 100 \text{ m}$ (levels 1–4), 50 m (levels 5–8), 25 m (levels 9–64) = total 2000 m. (In the run with $\Delta z = 50 \text{ m}$ levels 9–64 are reduced to levels 9–36 at 50 m each, and in the run with $\Delta z = 100 \text{ m}$ levels 5–64 are changed to levels 5–20 at 100 m each.)

A typical run at the finest resolution (2 km) is run for 25,000 time-steps with an integration time interval of $\Delta t = 300 \text{ s}$; this means a running time of about 87 days (3 months). Generally the overflow in the regions around the channel, reaches a steady state after about 50–55 days (12,000–15,000 time-steps). At around day 70 the plume reaches the southern boundary. At this point a mild change in the behavior of the plume is observed, possibly due to wall effects. Most of the analysis that follows is carried out on the results between days 50 and 70 (time-steps 14,400–20,200) in the 2 km resolution runs. This time-span was chosen as a compromise between having a steady plume near the sill, and at the same time having minimal effects of the wall included.

In Table 2 all runs carried out are listed. They differ in the inflow forcing, $Q$, the vertical viscosity, $\Lambda_z$, the drag coefficient, $C_d$, the vertical resolution, $\Delta z$, and the horizontal resolution, $\Delta x$ and $\Delta y$, used. In the first column an abbreviation is given to each experiment and from now on the experiments will in most cases be referred to by this name.

3. The plume structure

3.1. Control run

The control run (Contr) is the run against which all other runs will be gauged to ascertain the effect of changing certain parameters in the model. It has the highest horizontal and vertical resolution of all
runs, 2 km and 25 m, respectively, a vertical viscosity $A_z = 2.0 \times 10^{-4}$ m$^2$/s and a drag coefficient, $C_d = 2.0 \times 10^{-3}$.

To look at the structure and dynamics of the plume, that is, the extent of its spreading, its thickness and velocity of decent, we show snapshots of the plume thickness and the Froude numbers in Fig. 3a and b, respectively. The boundary of the plume is prescribed by a cutoff, $\tau = 0.1$, of the passive tracer. This limit applies for all results presented in this paper.

### 3.1.1. Thickness and spread

After exiting the narrowest and shallowest section of the FBC, the overflowing water mass first continues to be confined by the topography for about 50 km. Then the channel opens up onto the Faroe Iceland slope and a sloping bottom ($z \sim 13 \times 10^{-3}$) descends to 2000 m depth in the domain. The plume spreads to a width of about 200 km on the Faroe Iceland slope and its core remains confined between the isobaths 500 m and 1500 m.

The thickness plot (Fig. 3a) shows a patchiness of the plume. Generally the plume is about 150 m thick, however, several thicker patches up to 300 m can be seen along the slope. On reaching the Icelandic slope the plume makes a sharp 90° turn due to the topography and is thus steered in a southwesterly direction. Along this slope three distinct thick boluses of overflow water are apparent in the snapshot. However, we cannot say that the plume has reached a steady state in this region and so we cannot claim that these features are permanent.

An aspect of the flow that appears in the model but is absent in any observations is a current, which branches off from the plume once it has passed through the channel and wraps clockwise around the Faroe Islands. It seems that in the field a front of dense water over the FIR prevents any northward flow of plume water and in fact observations suggest a flow from the North over the ridge.

### 3.1.2. The Froude number

The Froude number is a good indication of the kind of dynamical regime the plume is in and it is often directly associated with the occurrence of entrainment. The entrainment parametrization developed by Turner (1986) based on experiments by Ellison and Turner (1959), prescribes entrainment as a function of the Froude number of the local flow. If the $Fr > 1.25$ entrainment occurs, for values $Fr < 1.25$ no entrainment occurs. This parametrization has been implemented and applied successfully in isopycnal models, such as the HIM and HYCOM (Hallberg, 2000), to prescribe diapycnal mixing. The MSBC also uses the Froude number to prescribe mixing and Price and Yang (1998) show a compilation of entrainment rates derived from laboratory experiments and observations as a function of the Froude number (Fig. 3, in Price and Yang (1998)). This

### Table 2

List of all sensitivity runs and their parameter settings

<table>
<thead>
<tr>
<th>Run</th>
<th>$Q$ (Sv)</th>
<th>$A_z$ (m$^2$/s)</th>
<th>$A_h$ (m$^2$/s)</th>
<th>$C_d$</th>
<th>$\Delta x$ (km)</th>
<th>$\Delta y$ (km)</th>
<th>$\Delta z$ (m)</th>
<th>$\Delta t$ (s)</th>
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<td>var</td>
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<td>2</td>
<td>2</td>
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<td>300</td>
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<tr>
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<td>$2.0 \times 10^{-4}$</td>
<td>var</td>
<td>$2.0 \times 10^{-3}$</td>
<td>2</td>
<td>2</td>
<td>25</td>
<td>300</td>
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<tr>
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<td>2</td>
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<td>300</td>
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<td>var</td>
<td>$2.0 \times 10^{-3}$</td>
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<td>2</td>
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<td>300</td>
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<td>2</td>
<td>25</td>
<td>300</td>
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<tr>
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<td>$2.0 \times 10^{-3}$</td>
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<td>2</td>
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<td>300</td>
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<tr>
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<td>var</td>
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<tr>
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<td>$2.0 \times 10^{3}$</td>
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<tr>
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<td>$2.0 \times 10^{-3}$</td>
<td>50</td>
<td>50</td>
<td>25</td>
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</tbody>
</table>

In the 2 km and 10 km runs the viscosity is variable (var) and set by the Leith scheme viscosity, for all other resolutions $A_h$ is constant.
picture has since been extended to include many other observations and laboratory experiments (Cenedese et al., 2004; Wåhlin and Cenedese, 2006).

Here we define the local Froude number of the flow as

$$Fr = \frac{U}{\sqrt{g' h}},$$

where $h$ is the thickness of the plume, $U$ is the vertically averaged velocity in the plume and $g'$ is the reduced gravity at each grid point of the plume defined as

Fig. 3. Snapshot of (a) the plume thickness (m) and (b) the local Froude number in the plume at days 64.58 of the control run. All average plume characteristics presented in this paper are computed considering tracer-laden water with $\tau > 0.1$ in the region bounded by the green lines, marked in (a) only.
Here $\rho_p(z)$ is the density in the plume, $\rho_{\text{int}}$ is the density at the interface taken from one grid point above the boundary of the plume, defined as $\tau = 0.1$, and $\rho_0$ is a reference density.

The Froude number is contoured in Fig. 3b. Regions of high Froude number are often associated with the edges of the flow where the plume is very thin. However, immediately downstream of the sill, away from the edge of the plume, the Froude number is large due to a more dynamic plume structure. Here velocities are particularly large (largest in the whole plume) and can reach up to 1 m/s; this is where most of the initial entrainment of the dense overflow happens – as we shall see.

Using the observational data from Mauritzen et al. (2005), Girton et al. (2006) compute the local Froude number at each station in a study to determine whether the overflow is hydraulically controlled. They find a patch of particularly high Froude numbers at the southern end (left-hand side) of sections F and G, where the overflow exits from the confinement of the channel and spreads out across the slope. The location of their patch of elevated Froude numbers agrees well with the location of high Froude numbers in the model.

Intermittently, a patch of large Froude numbers is observed in the corner between the FIR and the Icelandic slope, it was thought to be another mixing ‘hot-spot’. However, a run setup with an extended domain to the south (not shown) does not confirm this hypothesis. The patch of high Froude number here is not a consistent feature and occurs only occasionally.

The time averaged Froude number was computed along the plume by averaging each section of the plume’s Froude number in the $y$-direction, weighted by the area of the plume:

$$ Fr(x,t) = \frac{\int Fr(x,y,t)A(y,t)dy}{\int A(y,t)dy} , $$

where $A(y,t)$ is the area of a vertical slice of the plumes cross-sectional area at location $y$ and at time $t$. The dependence of $Fr(x,y,t)$ in Fig. 3b is thus reduced to $Fr(x,t)$. The results are then time averaged to get a mean profile of the Froude number in the along plume direction only ($\overline{Fr(x)}$). The time mean result is plotted in Fig. 4 as a thick white line, while the thin black lines are the results for individual time steps.

At the sill (0 km) the average Froude number is relatively high, though it never exceeds 1. A maximum in the Froude number is observed between about 50 m and 80 km downstream of the sill; here its value is about 0.77 on average. However, at individual time steps it can reach as high as 1.5 and more. Note that this is still an average value in $y$ and locally the Froude number can be even higher. This is a good indication of a highly
dynamic region, where most of the mixing occurs. After $x = -100$ km, $Fr$ remains fairly low and constant at around 0.5.

The region of high Froude number ($Fr > 1$), as seen in Fig. 3a downstream of the sill never extends all the way across the channel. It occurs periodically, then seems to be advected downstream and eventually disappears.

3.1.3. The transport

In Fig. 5 the transports at the sill, at a section 100 km downstream, and across the WTR are plotted as a function of time for $\tau > 0.1$. The horizontal lines in the figure indicate the mean flow at the sections. At the sill the mean transport is approximately 1.504 Sv and shows regular oscillations of amplitudes up to $\sim 0.5$ Sv. These oscillations peak about every 4–5 days. Further downstream ($x = -100$ km) the mean transport is 2.380 Sv suggesting an increase in transport of about 50%.

Regular time variability on scales of a few days is not easily discernable from observations. However, Mauritzen et al. (2005) repeated some of the observational sections several times thus allowing some insight into the variability of the transport at those sections. They find an average transport of 2.4 Sv at the sill, with a minimum observed value of 1.5 Sv and a maximum of 3.5 Sv. The somewhat lower transport in the model is most likely a result of the forcing at the northern boundary which is probably smaller than the actual transport in the field. Section G as presented by Mauritzen et al. (2005) lies about 100 km downstream of the sill; here they observe an average transport of 3.6 Sv. At section H, located 150 km downstream, the average transport is observed to be 2.7 Sv. These measurements clearly show that there is a large variability in the transport, the exact characteristics of which are not captured by the observations.

In addition to the FBC overflow plume, there is a mild overflow ($\sim 0.268$ Sv) over the Wyville Thomson Ridge. This overflow has been studied extensively in the field by Sherwin and Turrell (2005). In the past it was believed that the transport of this overflow was quite small in the range of 0.2–0.35 Sv; however, the recent observations show that it exhibits intermittent burst of transport as large as 2.1 Sv. One would think that for such a burst to take place a distinct and impulsive forcing mechanism has to occur somewhere upstream in the FSC regions. The forcing in the model presented here is uniform and steady in time; thus no such burst of a strong WTR overflow is observed.

3.1.4. Comparison with observational sections

For qualitative comparison of the plume structure in the model and the observations, model sections and observational sections are put side by side in Fig. 6a–f.

![Fig. 5. Transport (Sv) of overflow water through the sill section, $x = 0$ km, (open circles), at location $x = -100$ km (open squares) and over the WTR (closed circles). The tracer-limit used is $\tau > 0.1$. The dash-dotted, dashed and dotted lines illustrate mean transport between days 50 and 70 at the sill, at $x = -100$ km and over the WTR, respectively. The means are: 1.504 Sv at the sill, 2.380 Sv at $x = -100$ km, and 0.268 Sv across the WTR.](image-url)
The model sections are obtained by establishing the CTD station locations within the model domain and linearly interpolating the model results at the four nearest grid points to that location. In many of the sections, the observational stations are as much as 6 km apart. To make use of the much higher resolution (2 km) of the model output, \( n \) number of artificial stations are introduced between any two observational stations so as to preserve a horizontal resolution of about 2 km in the model sections. For example, if station 1 has

Fig. 6. Observed density at (a) section D3, (c) section F2, and (e) section G2 (courtesy of J. Price), compared with the time-averaged density of the model at (b) section D3 with \( n = 1 \), (d) section F2 with \( n = 2 \), and (f) section G2 with \( n = 3 \).
co-ordinates \((x_1, y_1)\) and station 2 \((x_2, y_2)\) which are approximately 6 km apart then \(n = 2\) extra stations are introduced between stations 1 and 2 with coordinates:

\[
\left( \frac{x_1 + x_2 - x_1}{n + 1}, \frac{y_2 - y_1}{n + 1} \right), \quad i = 1, 2, \ldots, n
\]

(provided \(x_1 < x_2\) and \(y_1 < y_2\)). Then the density at these artificial stations was again found by interpolating the model results at the four nearest grid points.

In the model we also have the added advantage of having a long time series of sections to look at, whereas observations tend to have a very limited temporal resolution. In what follows three observational sections (D–G) are compared with their model counterpart. The model sections used for comparison are time averaged over days 50–70.

The density distribution at the sill (Section D3) is shown in Fig. 6 a for the observations and in Fig. 6 b for the model section. Both the model and the observations show an increase in plume thickness from the left-hand side to the right-hand side of the channel; this is due to the effects of rotation.

The model and the observations show an increase in plume thickness from the left-hand side to the right-hand side of the channel; this is due to the effects of rotation.

The dense core in the model seems somewhat thinner over the whole section than the observations suggest. This is probably a result of the inflow of dense water into the domain prescribed at the northern boundary being too small and thus leading to a smaller transport through the sill than observed in the field. This suggests that the average observed transport used as a guide to prescribe the inflow is lower than the actual transport needed to supply the observed transport of dense water at the sill. This suggests that the average observed transport used as a guide to prescribe the inflow is lower than the actual transport needed to supply the observed transport of dense water at the sill. The interface slope bounding the densest overflow water is about 0.01 in both the data section and the model section.

Fig. 6c illustrates the density distribution at section F2 in the observations and Fig. 6d in the model. The core of the plume is confined to the slope between about 600–800 m depth both in the data and the model and has a width of about 30 km. The dense layer has approximately equal thickness in the model and the observations.

The time variability of the plume structure at this section is illustrated in Fig. 7 over a 5-day period. For most of the time the spread of the plume is not significantly different from the mean state, however occasionally the plume will recede up the slope as it does on day 68.06. At various times it will also become slightly narrower or thinner.

Fig. 6 e and f illustrates the density distribution of the plume at section G2 in the observations and the model, respectively. The spread of the plume in the observations is quite uniform over the entire slope covered by the section with little change in its thickness across the section. When looking at a time average plume structure in the model (Fig. 6 f) it appears significantly more diluted than the observed plume. This is due to large changes in the plume distribution in time which lead to a misleading picture of the average density at the section.

The details of this time variability are revealed in Fig. 8, where a series of section snapshots are plotted over a 5-day period. At day 63.89 the plume is thick and wide with a dense core that compares well to the structure of the observations, while only 1.5 days later on day 65.28, the plume has almost vanished and all that remains is a thin (~50–70 m) bottom current that then gradually builds up again. Comparing the observational section with the first snapshot in Fig. 8 it is clear that the model does quite well in representing the vertical thickness and horizontal extent of the plume.

Recent work by Geyer et al. (2006) on the time variability of the overflow has revealed a highly regular oscillation of the overflow downstream of the sill with a period of 88 ± 4 h (3.7 days); this is remarkably close to the oscillation observed in the model. Geyer et al. (2006) find that the oscillations are most prominent between 75 km and 140 km downstream of the sill and are best observed at mooring array A. Array A is located in roughly the same location as section H by Mauritzen et al. (2005), which lies about 50 km downstream of section G where we observe the clear structural changes of the plume presented in Fig. 8.

Both Geyer et al. (2006) and Ezer (2006) compare these oscillations to the supercritical wave regime observed in laboratory experiments by Cenedese et al. (2004). Furthermore, the analysis of idealized numerical two-layer experiments of an overflow by Kida (2006) have suggested that the oscillations are the results of baroclinic instabilities.
3.1.5. Interfacial layer thickness

To determine the interfacial layer thickness the observational density profiles were first filtered using a low-pass Blackman filter (Harris, 1978) with a halfwidth\(^1\) of 25. In this way the profile is smoothed and low frequency signals are reduced. The local density gradient was then computed along the filtered profile. A cutoff of \( \frac{\text{d}q}{\text{d}z} > 0.001 \) for the gradient was assigned to determine the top and bottom of the interfacial layer. In Fig. 9a the filtered and unfiltered profiles are shown along with the linear background profile used in the model. The absolute value of the gradient is plotted in Fig. 9b for the filtered and the unfiltered profiles. The locations where the gradient cutoff value applies are indicated by the horizontal lines.

For comparison the model data are interpolated to the station locations and the density profiles at these location are used to determine the bottom layer and interface layer thickness, which are then time averaged. The density gradient cutoff in the model is chosen to be \( \frac{\text{d}q}{\text{d}z} > 0.001 \), as for the observations.

It turns out that this cutoff is somewhat large and underestimates the interfacial layer thickness somewhat, but in tests it was found to give more consistent and less ambiguous results than lower cutoff values.

Fig. 10a shows a comparison between the density in the bottom 25 m (approximately) in the observations and the bottom level density in the model. There is a clear indication that the densest waters are probably not preserved as well in the model (though again it should be noted that the model values are averaged in time).

\(^1\) The ‘halfwidth’ is the half the number of data points to which the filter is applied.
In Fig. 10b the interface and plume thickness of both the model and the observations are plotted for each section. Both the homogeneous core layer, $h_c$, of the plume and the interfacial layer, $\Delta h$, characterized by a strong density gradient, are of similar thickness in both the observations and the model. The bottom layer thins significantly from $\sim 200$ m at the sill to $\sim 50$ m downstream as it spreads out across the slope. The interface thickness changes little as the plume makes its way downstream, it lies consistently at approximately 150 m. This means a total layer thickness, $h$, of the plume of about 200 m.

3.1.6. Summary

Overall, the control run with $\Delta x = 2$ km grid size represents the spreading and the vertical structure of the plume well. The interfacial layer thickness ($\Delta h$) and the core thickness ($h_c$) are very similar in both the model and observations. However, the bottom densities are more diluted in the model pointing to excessive mixing. A patch of high Froude number occurs about 50–80 km downstream of the sill; this patch has been seen in a recent analysis of the observational data (Girton et al., 2006). This is where most of the mixing is taking place in both the model and the observations as will be discussed in the next section.

The patch of high Froude number is an indicator of very high velocities in the plume. A comparison between the observed and modeled velocities in the region suggests that they are comparable with maximum velocities of 0.8–1.0 m/s. However, in the model there are occasional bursts of velocities that are very large (up to 1.5 m/s). We cannot say that such large velocities never occur in the field, given the small sample size of the
observations. The model velocities are, however, larger than the largest velocities usually recorded and mentioned in the context of observations.

The time variability of the plume is a clear feature in the model, which has also been described by Ezer (2006) in a recent idealized modeling study of the region. Clear observational evidence of this oscillation

Fig. 9. (a) Density profile and (b) density gradient profile of station 92 (section D3); thin black lines – unfiltered profile, thick black lines – filtered profile, horizontal black lines mark the cutoff bounding the surface layer, the interfacial layers and the plume core.

Fig. 10. (a) Bottom density, \( \rho_{\text{bot}} \) (open circles) at each section in the model (dashed line) and in the observations (solid line). (b) Interfacial layer thickness \( \Delta h \) (open triangles) and core thickness \( h_c \) below the interface (open squares) at each section in the model (dashed lines) and in the observations (solid lines).
was also presented by Geyer et al. (2006), showing a very good agreement of the timescales in the model and the observations. A plausible cause of the oscillations as a result of baroclinic instabilities was recently presented by Kida (2006).

3.2. Coarse resolution runs

The aim of discussing model runs with coarser resolution is to illustrate and understand how the model solutions degrade as the resolution of current climate models is approached.

Since the Faroe Bank Channel is barely 15 km wide at mid-depth one major effects of increasing the grid size is the gradual obscuring of the topographic characteristics of the channel and its surroundings. This in turn determines how the flow develops, whether the dense water can move down the slope and how much numerical mixing occurs. Furthermore at coarser resolutions the ability of the model to resolve small scale processes such as small eddies is affected.

3.2.1. Horizontal grid size 10 km

Fig. 11a shows the plume thickness on day 66.84 of run Hr10, which is considerably different from the control run. In the control run the plume is typically about 150–200 m thick while in the 10 km resolution case, large regions of plume thickness up to 400 m and above occur. However the horizontal extent of the plume is similar in both cases. It seems therefore that a coarser horizontal resolution promotes an enhanced mixing in the vertical, thus making the plume thicker and more diluted (as we shall see in the next section).

Fig. 11 b shows a snapshot of the Froude-number on the same day. As in the control run, a region of relatively high Froude number is observed just downstream of the sill between $-50$ and $-100$ km, however the numbers are not nearly as high as in the control run (hence the different color scale). This is a result of a much thicker plume as illustrated in Fig. 11a and smaller average velocities, as will be seen in the next section.

The mean transport through the channel and its variability are listed in Table 3. At the sill the transport is $2.071 \pm 0.176$ Sv, and downstream increases to $3.606 \pm 0.586$ Sv. The transport over the WTR is small at $0.300 \pm 0.050$ Sv. Compared to the control run more diluted water makes its way through the channel and there is a significant increase in transport downstream indicating a noticeable increase in mixing due to the increased grid size. The variability at the sill and downstream is reduced compared to the control run but a period of about 4 days is maintained.

3.2.2. Horizontal grid size 20 km

In Fig. 11c and d the plume thickness and the Froude number are shown respectively for a run with 20 km resolution and at day 110.24. In order to ensure the numerical stability of the run the horizontal viscosity, $\eta_h$, was changed from being determined by the Leith scheme to a constant value $2 \times 10^3$ m$^2$/s. In this run the topography of the Wyville Thomson Ridge was adjusted to a height of 500 m below the surface, to prevent excessive loss of dense water over the ridge.

The increase of the plume thickness from the 10 km run to the 20 km run does not seem as drastic as from 2 km to 10 km resolution. However, as in the 10 km resolution case there is a noticeable difference in the Froude numbers between the 20 km and the control run. Again at the coarser resolution the Froude numbers are much lower. It is therefore likely that the velocities are somewhat lower in the 20 km run; this will be confirmed in the analysis on the sensitivity runs presented in the next section. A lower velocity and an increased layer thickness conspire to lead to drastic changes in the Froude number. The lower velocities might be a direct result of introducing a larger horizontal viscosity.

The transports in the 20 km case are comparable to the 10 km transports: $2.150 \pm 0.069$ Sv at the sill, $4.477 \pm 0.571$ Sv 100 km downstream and $0.150 \pm 0.014$ Sv over the WTR. The variability, especially at the sill, has decreased compared to the higher resolutions runs which illustrates the decreased ability of the model to resolve meso-scale processes.

3.2.3. Horizontal grid size 50 km

In Fig. 11e and f the plume thickness and Froude number are shown for a run with 50 km horizontal resolution. The parameters for vertical and horizontal viscosity are the same as in the 20 km run, the only
difference here is the grid size. As in the 20 km run we also adjust the height of the WTR to 500 m below the surface to prevent excessive loss of overflow water across the ridge.

Fig. 11. Snapshot of the plume thickness (m): (a) at day 66.84 of the 10 km resolution run, (c) at day 110.24 for the 20 km resolutions run (with $A_z = 2 \times 10^{-4} \text{m}^2/\text{s}$ and $A_h = 2 \times 10^3 \text{m}^2/\text{s}$), and (e) at day 250 for the 50 km resolution run. In panels (b), (d), and (f) the local Froude number is shown for the same experiments at the same day, respectively. For all plots $\tau > 0.1$.

Table 3
Mean transport and variance at the sill, at $x = -100$ km and over the WTR, for runs with different horizontal resolutions: $\Delta x = 2, 10, 20,$ and 50 km

<table>
<thead>
<tr>
<th>$\Delta x$ (km)</th>
<th>$Q$ (Sv) sill</th>
<th>$Q$ (Sv) $-100$ km</th>
<th>$Q$ (Sv) WTR</th>
</tr>
</thead>
<tbody>
<tr>
<td>2</td>
<td>1.504 ± 0.267</td>
<td>2.380 ± 1.063</td>
<td>0.268 ± 0.164</td>
</tr>
<tr>
<td>10</td>
<td>2.071 ± 0.176</td>
<td>3.606 ± 0.586</td>
<td>0.300 ± 0.050</td>
</tr>
<tr>
<td>20</td>
<td>2.150 ± 0.069</td>
<td>4.477 ± 0.572</td>
<td>0.150 ± 0.014</td>
</tr>
<tr>
<td>50</td>
<td>1.923 ± 0.162</td>
<td>2.112 ± 0.165</td>
<td>0.210 ± 0.158</td>
</tr>
</tbody>
</table>
The plume thickness (Fig. 11e) is again very large compared with the control run. The Froude numbers (Fig. 11f) in the plume on the other hand are patchy and small with indication of a maximum in the channel. This leads to the speculation that the velocities in the plume have decreased even further when increasing the resolution. More evidence of the smaller velocities will be presented in the next section.

The mean transport (listed in Table 3) through the channel is $1.9 \pm 0.16$ Sv at the sill, and $2.1 \pm 0.17$ Sv 100 km downstream. The flow over the WTR remains small at $0.21 \pm 0.16$ Sv. The change in transport back to values that are more like in the 2 km resolution case is due to the highly diluted water mass that exits through the FBC onto the slope. The flow is already so diluted that very little extra mixing takes place. No regular oscillations are present in the flow at this resolution.

3.3. Summary

By changing the resolution (and the way the horizontal viscosity is set) in the model significant changes are observed in the structure of the plume.

At the highest resolution (2 km) in the model the structure of the plume in terms of its spreading along the slope, its thickness and dynamics agree well with the observations. There is some indication that the mixing may be a little excessive compared with the observations.

The major effect of a decreased resolution is the thickening of the plume in the vertical which points to an increase in mixing. In addition the velocities drop and the plume therefore changes from a fast flowing thin bottom current to a thick sluggish water mass with a large, diluted transport. At the coarsest resolution (50 km) so much mixing has happened before the plume reaches the sill, so that comparison with observations or the higher resolutions models becomes difficult. Velocities are so low now that the plume takes much longer to develop the same distance as in previous runs.

4. Results of the sensitivity study

To facilitate the improvement of the representation of mixing processes and the resulting formation of deep overflows in the ocean, it is vital to understand how the model behaves in different configurations. This allows us to understand where the main problems lie and how we might go about improving the model. Therefore a study of the sensitivity of the model to a variety of parameters such as the vertical viscosity, the drag coefficient, inflow forcing, and vertical as well as horizontal resolution is carried out. Subsequently the dynamical and tracer characteristics of the FBC plume are compared and contrasted for the different runs. In particular, we look at the plume path in the domain, as well as the change of the density of the plume along its path.

There are several regions in the domain where overflow water branches off from the main plume. These are for example the WTR overflow and the branch of flow around the Faroe Islands. These may distort the results of the main plume which we want to focus on here. Hence a region within the domain is defined, shown in Fig. 3a, and the analysis is restricted to tracer laden water contained in this region only. All water outside the region that may also contain tracer is ignored.

4.1. Plume path

In Fig. 12a the tracer-weighted average of the plume path, defined by equation:

$$Y(x) = \frac{\int y \tau(x, y, z) dy dz}{\int \tau(x, y, z) dy dz}, \quad (5)$$

is shown for all sensitivity runs superimposed on the topography. These paths are computed as time averages of the model outputs from day 50 to 70 for the 2 km resolution runs, from days 50 to 70 in the 10 km run, from days 60 to 80 in the 20 km run and from days 200 to 250 in the 50 km run.

The paths of runs with the coarser resolutions in the vertical and horizontal show the most noticeable deviation from the high resolution runs. Run Hr20 follows a path higher up the slope than the control run. This is
an indication of a more diluted plume as was also observed by Legg et al. (2006). The path of run Vr50 follows the majority of runs closely but lies somewhat lower down along the slope as it gets close to Iceland, this generally points to a denser, heavier plume; the same applies to run Hr50. It should be pointed out here that the topography in the plot is that for the 2 km resolution case. The plume path of Hr50 seems to move up the slope as the exit of the channel, the reason for this is that the topography in this run is so coarse that the mount south of the Faroe Islands has a much smaller horizontal extent and the channel opens up earlier.

All of the high resolution paths are tightly packed along the slope. The only exception is run Az2e2, which follows a somewhat lower path. We will see that in this case the plume is noticeably less diluted and hence denser.

The experiments by Legg et al. (2006) show that in the non-hydrostatic run with $\Delta x = 500$ m, the plume stays much further up the slope compared to the run with $\Delta x = 2.5$ km. This is consistent with more vigorous entrainment at the higher resolution due to partially resolved non-hydrostatic mixing processes. As the resolution is decreased further the plume paths lie again further up the slope, suggesting an increase in entrainment. These results are in agreement with what we find in this regional simulation, with the exception of run Hr50, which exhibits a very low path along the slope.

Fig. 12. (a) Average tracer-weighted plume path, defined by (5) for all sensitivity runs. (b) Time averaged tracer-weighted plume density, defined by (6).
In Fig. 12b the tracer-weighted average buoyancy of the plume, defined by equation:

\[ B(x) = \frac{\int b \tau(x,y,z) dy \, dz}{\int \tau(x,y,z) dy \, dz}, \]  

is plotted as function of \( x \), and using the same time averaging as before.

Again the coarser horizontal and vertical resolution experiments stand out because they reach a lower final density. All of them, with the exception of run Vr50 which follows the control run density closely, are already more diluted when the overflow water reaches the sill. Run Az2e2 stands out for having mixed significantly less than the plumes of the other runs.

The tracer-weighted density still decreases gradually beyond \( x = -100 \) km but the rate of decrease is much smaller than the initial rapid entrainment during the first 100 km immediately downstream of the sill.

If the coarser resolution experiments are already more diluted when they reach the channel, where does this dilution occur? In Fig. 13a the tracer-averaged density is plotted from the northern wall (\( y = 190 \) km for Contr, Vr100, Hr10 and Hr20 and \( y = 250 \) km for Hr50) to \( y = 50 \) km along the FSC and in Fig. 13b from the eastern boundary (\( x \approx 200 \) km) to \( x = 40 \) km from the sill. Most noticeably there is a decrease in the average density in the first grid point as soon as the dense water enters the domain. The control run and run Hr10 decrease by approximately 0.04 kg/m\(^3\), while run Vr100 and Hr20 decrease by approximately 0.05 kg/m\(^3\) and Hr50 decreases by 0.06 kg/m\(^3\). After that all five experiments continue to gradually dilute within the FSC (Fig. 13a) as well as on the approach to the FBC (Fig. 13b). The average density of the control run drops by 0.07 kg/m\(^3\) from the inflow to the sill, Vr100 drops by 0.14 kg/m\(^3\), Hr10 by 0.12 kg/m\(^3\), Hr20 by 0.17 kg/m\(^3\) and Hr50 by 0.33 kg/m\(^3\). These estimates illustrate that there is significant mixing and entrainment upstream of the FBC that critically affects the density of the plume after it passes through the channel.

A different way of looking at the changes in density along the plume is to estimate density averages of certain regions of the plume. We can then assess the density changes as a results of parameter changes more easily.

![Graphs showing density loss in FSC and density loss north of WTR](image-url)
Fig. 14 illustrates the sensitivity of the source and final density of the overflow to various parameters. The source density is taken to be the tracer-averaged density at the sill, while the final density is estimated as the average from the tracer averaged densities in the range $x = [-100, -200]$ km and $x = [-200, -300]$ km, in Fig. 12b.

Our findings suggest a clear decrease in the source density with the coarsening of the grid. The trend of the final density as a function of horizontal resolution is less clear; all final densities in Fig. 14a are similar in magnitude. However the difference between the source and final densities decreases with coarser resolutions, indicating a decrease in mixing. This trend is a direct result of the decreased source densities and the increased mixing upstream of the sill. It should be noted that direct comparison amongst the runs of different resolutions can be ambiguous since different horizontal viscosities are used at the various resolutions.

In Fig. 14b the changes in density are plotted as a function of the vertical resolution, $\Delta z$. There is a clear decrease in the initial density as the resolution decreases from 25 m to 50 m and 100 m. This decreasing trend is mirrored in the final densities downstream.

The result presented in Fig. 14c again confirms the surprising behavior of the model as the vertical viscosity is changed. At $A_z = [2 \times 10^{-9}, 2 \times 10^{-8}, 2 \times 10^{-7}]$ m$^2$/s, the source density and final buoyancy are very similar, however, the case with $A_z = 2 \times 10^{-2}$ m$^2$/s shows a markedly different behavior. The final density is much closer to the initial density of the overflow suggesting less entrainment.

Visual inspection of snapshots of velocity sections have shown that there is significant grid scale noise in the velocity fields for experiments $A_2e3$, Contr (with $A_z = 2 \times 10^{-4}$ m$^2$/s) and $A_2e5$. The velocity field was significantly smoother in experiment $A_2e2$, with almost no grid scale noise. We therefore concluded that the decreased mixing in $A_2e2$ is due to a smoother velocity field and thus smaller gradients in the velocity fields. When tracers are advected around by this smoother velocity field, less numerical diffusion is needed to eliminate small-scale noise in the tracer field. So even though the vertical viscosity does not enter directly into the tracer equation it has a significant implicit impact on it through the tracer advection.

![Fig. 14. Average density at the sill section (circles), average (final) density between $x = [-100, -200]$ km (squares) and $x = [-200, -300]$ km (triangles) of the tracer-weighted density (Fig. 12b) are plotted as a function of (a) horizontal resolution, $\Delta x$ (km), (b) vertical resolution $\Delta z$ (m), (c) vertical viscosity, $A_z$ (m$^2$/s), and (d) inflow forcing, $Q$ (Sv).](image-url)
Legg et al. (2006) use much higher vertical viscosities in their experiments: $A_z = 5 \times 10^{-2} \text{m}^2/\text{s}$ at $\Delta x = 2.5 \text{ km}$, and $A_z = 5 \times 10^{-1} \text{m}^2/\text{s}$ at $\Delta x = 10 \text{ km}$ and 50 km, making direct comparison somewhat difficult. Legg et al. (2006) show a comparison between the final density in a non-hydrostatic model run, in which the physical processes responsible for the mixing are partially resolved, and cases with a resolution of 2.5 km, 10 km and 50 km. They find that the final buoyancy computed for the same experiments does not show a clear monotonic behavior. In the reference case the mixing in the non-hydrostatic case and the 2.5 km case are similar suggesting comparable amounts of entrainment. At 10 km the buoyancy increases which is in agreement with the decrease of mixing, however, the vertical viscosity used is an order of magnitude larger and the decreased entrainment may be as a result of this change, as indicated by our results. At 50 km the final density is again close to the final buoyancy of the 2.5 km and the 500 m case indicating similar levels of entrainment. At the very high resolution the mixing is due to partially resolved shear-driven overturning and at the coarser resolution it is purely due to unconstrained numerical mixing.

Finally Fig. 14d illustrates that the change in source densities with increased inflow forcing is quite small but a gradual increase in final densities is observed with increased inflow forcing. This suggests that with increased transport of dense water through the FBC more of the dense water is conserved downstream.

Not shown in Fig. 14 is the dependence of the source and final densities on the drag coefficient. It was found that the drag coefficient has practically no effect on the density of the plume either at the source or further downstream. The range of $C_d$ was limited to $C_d = [2 \times 10^{-3}, 3 \times 10^{-3} \text{ and } 5 \times 10^{-3}]$.

Besides the changes in density of the plume we are also interested in changes of velocity and the thickness of the plume on average. The rms velocity, $u_{\text{rms}}$, and the average plume thickness, $h$, have been computed as a function of $x$ and are plotted in Fig. 15a and b, respectively. By $u_{\text{rms}}$ we mean the full root mean square velocity including the average speed. This measure of the velocity is closely related to the kinetic energy of the flow, which is available to do the mixing that we are interested in. In general a peak in the velocity, of up to 0.7 m/s, is observed in the first 50 km after the sill for all high resolution experiments. This peak tapers off downstream and the average velocity settles at about 0.2 m/s. Coarsening the resolution has the effect of eroding this velocity peak, both Vr100 and Hr10 show a peak but the values are somewhat lower (0.50–0.55 m/s) than those in the high resolution runs. The runs Hr20 and Hr50, do not exhibit a maximum in velocities after the sill, and the Hr50 velocities become very small explaining the very slow evolution of the plume in this run.

![Fig. 15. (a) Time-averaged root mean square velocity, $u_{\text{rms}}$ (m/s), along the plume, and (b) time-averaged plume thickness, $h$ (m), for all sensitivity runs (line styles are as marked in the legend of Fig. 12a).](image-url)
The thickness of the plume is remarkably constant along the path of the overflow for all high resolution runs. During the approach to the sill all runs have a very similar overflow water thickness but the Hr50, Hr20, Hr10 and Vr100 runs branch off near the sill giving rise to significantly thicker plumes. The Az2e2 run exhibits a marginally thinner plume, which is evidence for less entrainment.

4.2. Watermass transformation

The transformation of dense water in the plume is estimated in two regions marked A and B in Fig. 1. From looking at the Froude numbers and at the average changes of density along the plume in the previous section, we speculate that most of the dense water is transformed within region A. This region encompasses most of the observational sections downstream of the sill. Region B covers an area further downstream, where it is assumed that most of the overflow water has been transformed and less mixing occurs. The observational section H lies in this region.

Fig. 16a illustrates the transport of water masses in and out of region A. The density range is divided into 16 classes from 28.07 kg/m³ to 27.39 kg/m³ in intervals of 0.04 kg/m³. The densest class (28.03–28.07 kg/m³) thus corresponds roughly to the definition of FSCBW by Mauritzen et al. (2005), of which they observe a transport range between 0.4 Sv and 1.5 Sv at section D. This corresponds well to our mean of about 0.9 Sv of dense water transport into region A.

The next four bars (27.87–28.03 kg/m³) correspond to the definition of NSAIW (Norwegian Sea Arctic Intermediate Water). At the sill Mauritzen et al. (2005) find a transport of about 0.75 Sv in this density class, the model results suggests an average transport of about half of that value (0.35 Sv) into the entrainment region. More of this density water is exported, i.e. produced, in the region than is imported.

The following six bars (27.63–27.87 kg/m³) correspond to the approximate density range of AI/NIW (Arctic Intermediate/Norwegian Intermediate Water). In the model very little transport of this water mass into the entrainment region is observed, but the majority of transport out of the region lies in these density classes. This means that all the initial dense water is transformed to water with these density characteristics. This agrees with the observations, which suggest the largest increase in transport in this density range between the sill section and the section furthest downstream (see Fig. 16 in Mauritzen et al. (2005)).

![Graph](image-url)

**Fig. 16.** (a) Transport in $T_{in}$ in gray, and out $T_{out}$, in white, of region A divided into density classes. The black line marks the net transport for each density class. (b) Same as (a) for region B.
Fig. 16b shows the transport in and out of region B. Again a net transport of higher density classes into the region is observed with a dilution towards lighter classes. In general there is a lot of advection in and out of the region and a relatively small amount of water is being transformed.

The observations presented by Mauritzen et al. (2005) show that by the time the overflow has reached section H, none of the very densest original water mass prevails, which agrees with the water transformation observed in the model. None of the densest water entering the primary entrainment region A leaves this region intact.

Fig. 17 is a compilation of the net transport in and out of region A for most runs listed in Table 1. All of the high resolution runs show a very similar transport structure in the spectrum of density classes. As expected run Az2e2 shows much less water transformation than the other high resolution runs.

As the resolution coarsens the spread of density classes entering the region increases and the transport out of the region shifts to lower density classes. In the Hr50 run no water of the densest class remains by the time the plume enters region A; it has all been diluted upstream of the sill.

4.3. Entrainment

Fig. 16a and b have already indicated that most entrainment takes place in region A, within the first 100 km of the sill. We will now quantify the entrainment using two different methods: A bulk entrainment estimate for regions A and B and an estimate using density gradients derived and presented in Fig. 12b.

4.3.1. Transport method

To estimate the entrainment we quantify the change of the transport of overflow waters in and out of a given region. This method has been used by Legg et al. (2006) to estimate the entrainment immediately after the dense water emerges from the channel onto the slope. The entrainment is given by the following expression:

\[ E_t = T_{out} - T_{in} + \frac{dV}{dt}, \]

where \( T_{out} \) is the combined outward transport through all boundaries of the region, \( T_{in} \) is the transport into the region, which in the case of region A enters mainly through the FBC, at the eastern boundary, and \( \frac{dV}{dt} \) is
the rate of change in volume of the plume within the regions at a given time step. The entrainment velocity is then given by

$$ W_e = E_t / S, $$

(8)

where $S$ is the surface of the plume in the region. It is determined at each time step to be the surface area of the box that is covered by the plume. Finally the entrainment coefficient is defined as

$$ a_E = W_e / U_{rms}, $$

(9)

where $U_{rms}$ is the average root mean square of the horizontal velocities in the region.

In this way the entrainment coefficients have been estimated for regions A and B, and the results are plotted as a function of the parameters $\Delta x$, $\Delta z$, $A_z$, and $Q$ in Fig. 18a–d, respectively. Also shown in the figures is the entrainment coefficient derived from observations, which was estimated to be $\sim 5.5 \pm 1.7 \times 10^{-4}$ by J. Price (pers. com.) and Price et al. (2005).

The estimates confirm that most mixing occurs in the region encompassed by region A, where the $a_E \sim 3-5 \times 10^{-4}$ for most sensitivity runs. Markedly less mixing occurs in region B, here $a_E \sim 1-2 \times 10^{-4}$. The most notable changes in entrainment occur with changes in the horizontal resolution. As the grid size coarsens there is a marked increase in entrainment, which at $\Delta x = 20$ km exceeds the observed value. It drops down below the observed value at $\Delta x = 50$ km because the plume is already very diluted by the time it enters the region.

The vertical resolution, $\Delta z$, and vertical viscosity, $A_z$, also have a noticeable effect on the entrainment rates. Entrainment increases somewhat with increasing vertical grid size and it decreases with increasing values of $A_z$. These trends have already been observed when looking at along plume changes in density (see Fig. 12b). The magnitude of the inflow forcing, $Q$, has little effect on the mixing of the plume. We also investigated the sensitivity of the runs to changes in the drag coefficient, $C_d$, but due to space constraints the results are not shown here. In region B the entrainment coefficient is practically the same for all coefficients used; in

![Fig. 18. Bulk entrainment coefficients are estimated in region A (circles) and region B (squares) and plotted as a function of: (a) horizontal resolution $\Delta x$ (km), (b) vertical resolution, $\Delta z$ (m), (c) vertical viscosity $A_z$ (m$^2$/s), and (d) inflow forcing, $Q$ (Sv). The dashed black line marks the value determined from the observations by J. Price (pers. com.).](image-url)
region A, a very slight decrease in the entrainment coefficient is observed from \( \alpha_E = 4.42 \times 10^{-4} \) at \( C_d = 2.0 \times 10^{-3} \) to \( \alpha_E = 4.19 \times 10^{-4} \) at \( C_d = 3.0 \times 10^{-3} \) and then to \( \alpha_E = 4.02 \times 10^{-4} \) at \( C_d = 5.0 \times 10^{-3} \).

The increased mixing with increasing horizontal resolution observed here is in contrast with the results presented by Legg et al. (2006), who consistently find that as \( \Delta x \) is increased the entrainment coefficient shows a slight decrease. They speculate that the decrease is due in part to a decrease in the dense flow velocities at coarser resolution. We observe a clear decrease in velocities in the region (Fig. 15) with increased grid size, however, the entrainment coefficient increases at the same time for all resolutions with the exception of \( D_x \) = 50 km. Some of the decrease in entrainment observed by Legg et al. (2006) may be due to the increased vertical viscosity used as the grid size is coarsened.

Another way to estimate the entrainment using the transport in and out of region A can be derived from the net transport curve in Fig. 16. If the curve is integrated for each timestep and then divided by the surface area of the plume at that timestep, the entrainment velocity is derived as a function of time. Averaging that velocity we find \( w_e = 2.137 \times 10^{-4} \) m/s for run Contr plotted in Fig. 16. If \( w_e \) is scaled using \( u_{rms} \) we find the entrainment coefficient \( \alpha_E = 4.126 \times 10^{-4} \). Note that these values are slightly smaller than those derived using the transport method above (\( w_e = 2.289 \times 10^{-4} \) and \( \alpha_E = 4.417 \times 10^{-4} \)), this difference is due to the factor \( \frac{dy}{dt} \) which is included in the calculation of entrainment from Eq. (7), but not when it is calculated from the transports shown in Fig. 16.

### 4.3.2. Tracer method

Instead of using the conservation of volume to derive the entrainment, we can also use conservation of mass. The expression of mass conservation formulated for a control volume, \( V \), is

\[
\frac{D(V \rho_p)}{Dt} + \nabla \cdot (V \rho_p) = S w_e \rho_{int},
\]

where \( \rho_{int} \) is the density at the interface taken from one grid point above the boundary of the plume, defined as \( \tau = 0.1 \), and \( \rho_p \) is the density of the plume. Analogous to Eq. (7) mass conservation can be rewritten as

\[
\frac{d(V \rho_p)}{dt} + (A u_n \rho_p)_{out} - (A u_n \rho_p)_{in} = S w_e \rho_{int},
\]

where \( A \) is the surface area of in and out flow (they can be different) and \( u_n \) the velocity normal to the area of in and outflow.

In observational studies, such as those by J. Price in the FBC (pers. com.) and Price et al. (2005) and Girton and Sandford (2003) for the Denmark strait, this expression is simplified to a pointwise expression (as opposed to an expression that uses a finite control volume). Furthermore, the plume is assumed to be steady. Under these considerations and using various averaged quantities, the entrainment rate is determined simply from the density gradient along the plume:

\[
w_e = \frac{u \bar{h} \, d \rho}{d \rho / d x}.
\]

Here \( \bar{u} \) is the average rms velocity in a given stretch of the plume. As in the previous computations for the along plume density changes, the along plume rms velocity is computed at each time step and then averaged in time. The rms velocity is then averaged over 100 km long sections of the plume.

Similarly, the average plume thickness \( \bar{h} \) is defined by

\[
\bar{h} = \frac{\int h \, dy}{\int dy},
\]

where the interval of integration from one side of the plume to the other is defined by the tracer cut off point \( (\tau > 0.1) \). The \( \bar{h} \) we use is then simply defined as the mean plume thickness (from Fig. 15) averaged over 100 km long segments of the plume, as before. We define \( \delta \rho \) by

\[
\delta \rho = \frac{1}{\bar{h}} \int_0^\bar{h} \rho_p(z) \, dz - \rho_{int},
\]
where \( \rho_p(z) \) is the density in the plume and \( \rho_{int} \) the density at the interface. This is different from the definition of the density difference (denoted \( \rho' \)) by Girton and Sandford (2003) who take it to be the contrast between the plumes density and the density of the background stratification. The plume is in more immediate contact, however, with waters above it and so a contrast between the plume’s density and the interfacial density might be somewhat more appropriate to use. The two estimates are only marginally different. In the first 100 km there is no difference in \( \Delta \rho \) between the two estimates. Beyond 100 km the average density difference is somewhat lower when the interfacial density is used as opposed to the background density. The entrainment coefficient is then defined by

\[
\alpha_E = \frac{w_E}{\bar{u}}.
\]

So technically \( \bar{u} \) is not needed to find the entrainment coefficient, as it cancels out.

We shall use the density gradients derived using a linear fit to the tracer weighted densities in Fig. 12b in three section along the plume: (a) between the sill and \(-100\) km, (b) between \(-100\) km and \(-200\) km, and (c) between \(-200\) km and \(-300\) km.

In Fig. 19a and b the sensitivity of the gradient entrainment coefficient derived using expression (12) to the horizontal and vertical resolution is plotted, respectively. Clearly, the sensitivity to \( D_z \) reveals that most of the entrainment happens in the first 100 km with less in the mid and end section of the plume. The trends in the experiments with varying \( D_x \) are much less conclusive. At the finer resolutions 2 km and 10 km there is significantly more mixing in the first 100 km than further along the plume. As the horizontal resolution decreases, however, the entrainment rates become closer in magnitude in all regions along the plume. Compared to the observations which are derived in a similar fashion, the entrainment estimates in the region close to the sill are up to a factor of five higher.

There is a noticeable difference in the results from the two methods. The bulk entrainment estimates are consistently lower than the estimates using density gradient expression (12). An estimate computed using the full mass conservation Eq. (11) applied to the same control volume that we used for the transport method, gave estimates of the entrainment coefficient that are very similar to those derived from the transport method: \( \alpha_E = 4.417 \times 10^{-4} \) from the conservation of transport and \( \alpha_E = 4.047 \times 10^{-4} \) from the conservation of mass. Both of these estimates are for run Base26. The estimate using the density gradients in the first 100 km section of the plume is \( \alpha_E = 1.685 \times 10^{-3} \). Most of this discrepancy is undoubtedly due to the fact that the density gradient method utilized highly simplified and averaged plume quantities.
It should be noted that in both estimates the diffusive fluxes are not explicitly accounted for. However, they will directly affect the density difference and density gradients in the second method. In the transport method it is the viscosity that will affect the estimate and is again not explicitly accounted for.

The estimates all cluster around the observed value with the bulk estimates mostly somewhat lower and the density gradient estimates somewhat larger. Both estimates are clearly most sensitive to the changes in $\Delta \chi$, but no clear monotonic trend is observed.

5. Conclusions

We have described an extensive comparison between model simulations of the Faroe Bank Channel overflow and observations. A principal conclusion of this comparison is that at the highest horizontal resolution of 2 km the representation of the plume in the model compares well with the observations. A visual comparison of the model with the observational sections has shown that the plume’s thickness and spread agree well. This agreement is confirmed by the quantitative estimates of the interfacial layer thickness ($\Delta h$) which remains constant at around 150 m in the model as well as the observations and the thickness of the core ($h_c$) of the plume, which decreases significantly from around 150–200 m at the sill to 50 m downstream. Estimates of the entrainment rate have shown that the amount of mixing in the model at highest resolutions is comparable to the observed mixing.

The model results have shown clearly the time variability of the overflow. At the highest resolutions of 2 km and 10 km a clear oscillation of the overflow with a period of about 4 days is seen. This oscillation is absent at coarser resolutions since the model loses its ability to resolve mesoscale processes.

Model runs at progressively coarser resolutions (10 km, 20 km, and 50 km) have been used to illustrate the degeneration of the structure and dynamics of the plume as climate model resolutions are approached. The horizontal spread of the plume is very similar in all experiments but two main problems have been identified, which lead to a distortion of the behavior of the plume: (1) The plume becomes too thick. For all resolutions coarser than 2 km the plume was observed to be significantly thicker on average suggesting a larger entrainment into the plume from above. (2) The plume becomes much more sluggish. At 10 km resolution a peak in velocity is still observed at and just beyond the sill, however, at 20 km and 50 km this peak in high velocities in the entrainment region is absent. Downstream of the sill the velocities in all runs are comparable, apart from the 50 km run in which the velocities in the plume are only about 25% of what they are in all other runs causing the plume to evolve approximately four times slower. The setup in terms of horizontal and vertical viscosity is identical in the 20 km and 50 km experiments; thus it is difficult to ascertain why exactly the 50 km run develops so differently.

The thicker plume structure is of course an indicator of enhanced mixing in the coarse resolution runs. Estimates of density changes along the plume, the entrainment coefficient at various regions along the plume and the final density of the product water have confirmed the enhanced entrainment at coarser resolutions. However a large amount of this enhanced mixing happens upstream of the FBC before the overflow waters reach the sill. The entrainment in the plume itself actually decreases slightly with a coarsening grid because the source waters are already very diluted.

The high resolution runs confirm what has recently been observed in the field: that the best conditions for entrainment into the plume, i.e. high velocities and steep slopes, are to be found approximately 70 ± 20 km downstream of the sill. The final product water conditions are essentially set in these first 100 km of the plume.

Sensitivity studies presented have shown that the entrainment in the model is most affected by changes in the horizontal resolution as already discussed, showing an overall increase in entrainment with increase in grid size most of which occurs upstream of the sill. Another parameter that has a noticeable effect on the structure and entrainment of the plume is the vertical viscosity, $A_z$. For the highest $A_z = 2 \times 10^{-2}$ m$^2$/s the plume remains thinner and less entrainment takes place downstream, resulting in a denser plume on average. Visual inspection of snapshots of velocity sections have shown conclusively that there is significant grid scale noise in the velocity fields for experiments Az2e3, Contr (with $A_z = 2 \times 10^{-4}$ m$^2$/s) and Az2e5. The velocity field was significantly smoother in experiment Az2e2, with almost no grid scale noise. We therefore concluded that the reason for the decreased mixing is due to the smoothing of the velocity field as a result of the increased vertical viscosity. The vertical viscosity does not directly affect the tracer field, but a change in the gradients of the velocity field has a direct impact on the tracer field through the tracer advection.
The nature of the advection scheme is such that numerical diffusion is introduced to eliminate small-scale noise in the tracer field. Higher viscosities lead to smoother velocity fields (seen as a tendency for less energy near the grid scale in both snapshots and spectra). When the tracer is advected by a smoother velocity the tendency to generate grid-scale noise in the tracer field is reduced, and so less numerical diffusion is needed. Hence higher viscosities lead to less mixing.

In the model most of the mixing occurs in locations where it is necessary to preserve the smoothness of the advected tracer field. These locations coincide with strong velocity gradients, i.e. large shear, and small Richardson numbers (large Froude numbers). Hence the mixing occurs in the right place in the high resolution model, despite the fact that there is no explicit mixing parameterization. However, the fact that the amount of numerical mixing is sensitive to the vertical viscosity coefficient is a reminder that this numerical diffusion is not a physically based parameterization of mixing, and the approximate agreement between the magnitude of mixing and observations is probably fortuitous. In order to more precisely control the amount of mixing, the numerical mixing would have to be reduced greatly, so that it was significantly less than the observed mixing, and then a physically based mixing parameterization applied. New, less diffusive advection schemes (e.g. Prather, 1986) are currently being implemented in several ocean models, and future work will examine the numerical mixing produced when these schemes are used to simulate overflows. A possible candidate for a physically based parameterization of the mixing is the new nonlocal shear-driven mixing scheme under development by Jackson et al. (submitted for publication). At very coarse resolutions results may be improved by a reduction in spurious mixing upstream of the sill (by use of a less diffusive advection scheme) combined with a better representation of the narrow channels. Such a representation of channels smaller than the grid size, the partially open barrier algorithm, is currently being pursued by Adcroft and Hallberg (pers. com.), and we plan to test this new scheme in this configuration for the Faroe Bank Channel.

In summary, our study has evaluated the plume dynamics and mixing in a regional simulation of the Faroe Bank channel using the z-level MITgcm. We have found that the model reproduces the results obtained from observations well at the highest resolutions, while at coarser resolutions results are degraded. Coarse resolution leads to simulated plumes which are thicker and more sluggish. Plume entrainment has been shown to decrease for increased vertical viscosity. These results are now motivating ongoing research into the use of less diffusive advection schemes and new algorithms for representing narrow channels as part of the Gravity Current Entrainment Climate Process Team.

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