

The deepening of the Atlantic water in the Lofoten Basin of the Norwegian Sea, demonstrated by using an active reduced gravity model

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[1] The Norwegian Atlantic Current (NwAC) as a poleward eastern boundary current is to be considered as the northern limb of the Meridional Overturning Circulation in the North Atlantic (MOC). It transports warm and saline Atlantic water (AW) northward toward the Arctic Ocean, before cooling and mixing with cold and low saline water masses return it to the North Atlantic to contribute in the MOC. In this study we focus on the AW in the Lofoten Basin (LB) of the Norwegian Sea (NS), where it occupies a wider and deeper domain than farther south and north. It comprises the major heat-reservoir in the NS with contact with the atmosphere, and as such is an important area for cooling and buoyancy loss. We show that the deepening of the AW is mainly caused by the reduced northward flow in the LB, manifested in the deep counter-current encountered by the Polar frontal jet. We demonstrate this effect by using an active reduced gravity model with a topographically-steered deeper flow. To achieve maintenance of volume transport, the model shows that observed differences must be balanced by an oppositely-directed deeper flow in the LB. The ocean appears to achieve maintenance of volume transport of AW due to buoyancy loss and varying deeper currents, by adjusting the vertical extension of AW, resulting in a substantial deepening in the LB. Based on the good fit between the model and observations, the suggested mechanism appears to be credible and feasible. **INDEX TERMS:** 1635 Global Change: Oceans (4203); 4512 Oceanography: Physical: Currents; 4516 Oceanography: Physical: Eastern boundary currents; 4528 Oceanography: Physical: Fronts and jets; 4536 Oceanography: Physical: Hydrography. **Citation:** Orvik, K. A. (2004), The deepening of the Atlantic water in the Lofoten Basin of the Norwegian Sea, demonstrated by using an active reduced gravity model, *Geophys. Res. Lett.*, 31, L01306, doi:10.1029/2003GL018687.

1. Introduction

[2] The Atlantic inflow to the Norwegian Sea (NS) manifested by the Norwegian Atlantic Current (NwAC) can be considered as the northern limb of the Meridional Overturning Circulation in the North Atlantic (MOC), and thus an important factor for climate and ecology. In a broad outline the NwAC is a poleward eastern boundary current [Veronis, 1973] transporting warm and saline water to the Barents Sea (BS) and Arctic Ocean (AO). Within the Arctic Mediterranean this warm and saline Atlantic water (AW) is

continuously modified, for the most part by (1) heat loss to the atmosphere before it subducts in the vicinity of Svalbard, (2) fresh water supply from the atmosphere and via rivers flowing into the BS and AO. The AW also partly recirculates and mixes in the Fram Strait with cold and low saline water before returning to the North Atlantic as a western boundary current and overflow [Mauritzen, 1996], to contribute in the MOC.

[3] In this study we focus on the AW in the Lofoten Basin (LB) of the NS (Figure 1), where the AW occupies a wider and deeper domain. The LB is an area of AW approximately 5×10^5 km² in size, with the two major branches of the NwAC on its eastern and western sides [Orvik and Niiler, 2002]. In the southern NS the AW appears as a slab, about 250 km wide and 500 m deep [Mork and Blindheim, 2000; Orvik et al., 2001]; in the LB the slab widens to 500 km and deepens to 800–900 m, before it again becomes more shallow and narrow and loses contact with the atmosphere farther north [Mauritzen, 1996]. Figure 2 shows the development of the AW (traditionally defined as the water mass with $S > 35.0$) as it flows through the NS toward Arctic.

[4] The LB comprises the major heat reservoir of AW in the NS which has contact with the atmosphere. In that sense its importance for the MOC is reflected through a net atmospheric heat loss of ~ 70 W/m² with a maximum above 100 W/m² [Mauritzen, 1996], resulting in strong surface cooling and buoyancy loss. The heat loss to the atmosphere is manifested through a temperature drop from roughly 5–10°C in Svinøy section (SS), to 3–6°C in LB, and further to 2–4°C west of Bear Island [Mauritzen, 1996]. The LB is also characterized by strong eddy fields, and long residence time for drifters, and presumably also for the AW [Poulain et al., 1996].

[5] In this study we seek to understand the dynamical basis for the deeper domain of AW in the LB. Our hypothesis is that this deepening is mainly caused by the deeper flow in the NS. In Orvik and Niiler [2002] we showed a topographic steering of the baroclinic jet in the Polar Front (PF) (the western branch of the NwAC), along significant bottom features such as the western slope of the Vøring Plateau (VP), the eastern slope of the Mohn Ridge (MR), and the western slope of the Knipowich Ridge (KR). For a near-surface baroclinic flow to be influenced by the shape of a deep ocean basin, a topographically-steered deeper flow is required. The along-isobath topographic steering of a near-surface baroclinic flow was demonstrated by Svendsen et al. [1991] using a simple two-layer geostrophic model. They expressed the topographic steering as $\vec{v}_1 \cdot \nabla h = 0$, where \vec{v}_1 is the current in the upper layer, the

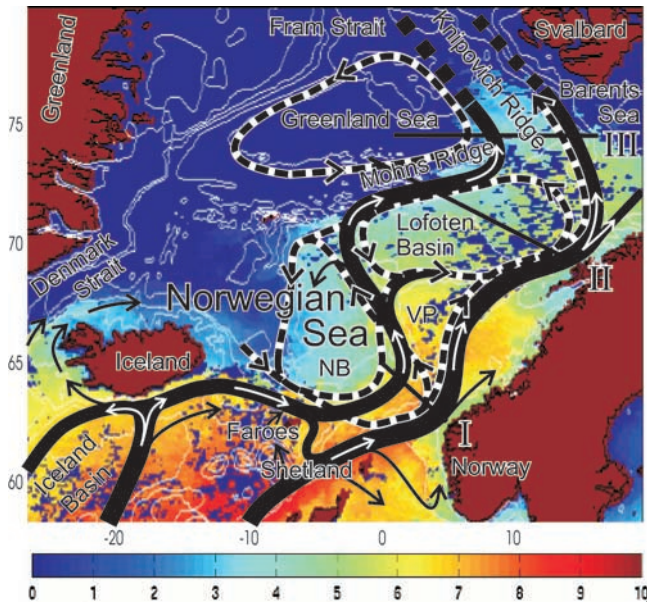


Figure 1. Schematic of the major pathways of near-surface Atlantic water through the Norwegian Sea (dark arrows), in the context of superimposed sea surface temperature from AVHRR image in March, 1991. The dashed arrows indicate the deeper flow. The straight lines show the Svinøy section (I), Gimsøy-NW section (II), and Bear Island-W section (III). Abbreviations explained in the text. [Based on Poulain *et al.*, 1996; Orvik *et al.*, 2001; Orvik and Niiler, 2002; Jakobsen *et al.*, 2003; Nøst and Isachsen, 2003].

total depth being $h = h_1 + h_2$. In fact, the pressure field attributed to a topographically-steered deeper flow acts as an artificial bottom for the baroclinic flow in the upper layer.

[6] To date, deeper current records from the central NS are rather sparse. According to model results [e.g., Nøst and Isachsen, 2003], in its broad features the deeper and near bottom flow consists of strongly topographically-steered cyclonic gyres in the Norwegian Basin (NB), the LB and the Greenland Sea (GS) (Figure 1). In consequence, the slab of AW flows with a deeper northward flow in the SS and Bear Island-W section, while it encounters recirculation with a reduced northward flow in the LB. Since the deep current along the shelf slope appears to be uniform northward all through the eastern NS (Figure 1), it cannot account for the reduced northward transport and deepening of the AW in the LB. In that sense, we concentrate on the deeper flow in connection with the western branch of the NwAC. This is a northward deeper flow in the eastern NB including the SS and the western slope of the VP, a similar southward deeper flow along the eastern MR in the LB, and then again a northward deeper flow farther north along the KR, in association with the GS gyre. This is in accordance with the pathways of the jet in the PF steered along the 2000 m isobath through the SS and along the slope of the VP, before it turns northeastward along the MR, and then again north along KR [Orvik and Niiler, 2002]. In that sense the frontal jet is flowing with a deeper current in the eastern NB and VP before it faces a deeper counter-current when it switches over to the LB along the eastern MR, before it again enjoys the advantage of a deeper current in the same

direction farther north toward the Fram Strait. In this study we will substantiate the hypothesis that the reduced northward transport in the LB, manifested in the deeper counter-current the frontal jet faces along the eastern MR, is an important factor for the deepening of the AW in the LB. We will demonstrate that effect by applying a simple two-layer reduced gravity-model with an active deeper layer, comparable with the classical reduced gravity model introduced by Stommel [1965] for the Gulf Stream.

2. Methodology

[7] We approximate the AW current associated with the baroclinic part of the PF jet, by using a simple two-layer geostrophic model with an active lower layer (Figure 3). The densities of the lower and upper layer are ρ_0 , and ρ respectively, the upper layer having thickness H . The interface H outcrops at the surface along the y -axis ($H = 0$), forming a front in accordance with the PF and its accompanying baroclinic jet. Assuming an along-isobathic geostrophic flow confined along steep slopes in both layers [Svendsen *et al.*, 1991], the current v_0 in the lower layer becomes,

$$v_0 = (\rho_0 f)^{-1} \frac{\partial p}{\partial x}. \quad (1)$$

The x and y -axes are in across- and along isobathic (frontal) directions, respectively.

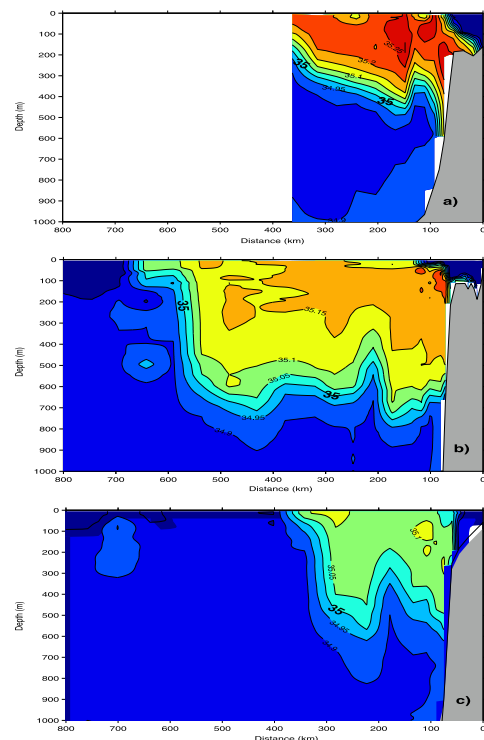


Figure 2. Development of the AW from Svinøy section (a) through the Gimsøy-NW section (b) to Bear Island-W section (c), presented in terms of salinity transects along sections (a), (b), and (c). Atlantic water is traditionally defined as water with $S > 35.0$.

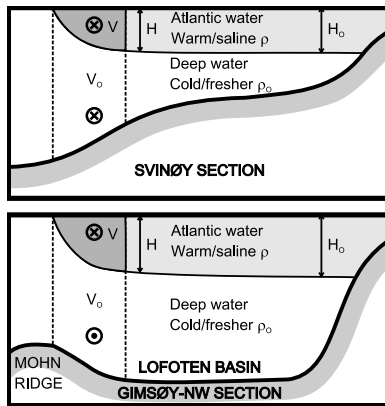


Figure 3. Schematics of the active reduced gravity model, with a deeper along current in the “Svinøy section” and a countercurrent over the Mohn Ridge in the Lofoten Basin.

[8] Using (1) as boundary condition at the interface, the current in the upper layer becomes

$$v = \frac{g'}{f} \frac{\partial H}{\partial x} + v_0, \quad (2)$$

where $g' = g \frac{\rho_0 - \rho}{\rho_0}$ is the reduced gravity, and f is the Coriolis parameter.

[9] The total volume transport in the upper layer associated with the frontal jet ($V_{Vol} = \int v ds$), where s is the across-stream area, is then

$$V_{Vol} = \frac{g'}{2f} H_0^2 + \int v_0 ds, \quad (3)$$

where H_0 is the maximum thickness of upper layer.

3. Application and Discussion

[10] The active reduced gravity model (Figure 3) coincides fairly well with the slab-like extension of the AW both in the SS [Orvik *et al.*, 2001] and the LB (Figure 4), and appears feasible for demonstrating interfacial changes. Concerning a classical reduced gravity model with a passive lower layer [Stommel, 1965], the cross-stream profiles of both upper-layer thickness (H) and current (v) turn out to be exponential structures decaying away from the front on a length scale of a Rossby deformation radius. With an active lower layer, the cross-stream profiles will be affected by the deep flow, but in any case trapped along the steep topography [Svendsen *et al.*, 1991]. However, a specification of the flow field is outside the scope of this study, since we are focusing on volume fluxes in accordance with (3). The hydrographic transect across the LB (Figure 4) shows a wider front, with a westward excursion of the surface layer of AW. This is because of the meandering and eddy structure of the PF, resulting in mixing on several scales [Blindheim, 1990; Allen *et al.*, 1994; Orvik *et al.*, 2001]. Since configuration in section 2 does not specify total transport (3), use of this model is justifiable because the volume transport is related to advection and not details of the frontal structure. In the SS the strong deeper flow coincides with

the PF [Orvik *et al.*, 2001], and from models [Nøst and Isachsen, 2003] it appears to be the case over the MR, elucidating the basis of this model.

[11] In this study we assume that the volume transport (3) of AW along the pathway of the PF jet is constant. A justification for this assumption is given in Orvik and Niiler [2002], who substantiated the continuity of the frontal jet from Iceland to Svalbard noting a volume transport of 3.5 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) in the Iceland-Faroe Front [Perkins *et al.*, 1998], 3.4 Sv in the SS [Orvik *et al.*, 2001], and 3.0 Sv farther north over the KR [van Aken *et al.*, 1995]. These estimates are based on standard zero reference depths i.e., no deeper flow. We use the same methodology to estimate the volume transport associated with the PF in the LB, where Figure 4 shows a baroclinic jet about 100 km wide and 500 m deep with a maximum speed of over 15 cm/s. The volume transport of this jet turns out to be about 5.0 Sv, which does not fit with maintenance of volume transport northward from the SS (3.4 Sv). To justify this discrepancy, the zero reference level methodology is extended by taking into account the deeper flow. This is demonstrated using the active reduced gravity model (Figure 3), and by use of observations.

[12] According to (3), the maintenance of the volume transport of the northward flowing AW will mainly be disrupted by (1) Gradual buoyancy loss due to heat loss to the atmosphere, resulting in decreasing reduced gravity g' (first term $\frac{g'}{2f} H_0^2$), or (2) changes of the deeper flow v_0 along the pathway of the baroclinic flow (second term $\int v_0 ds$).

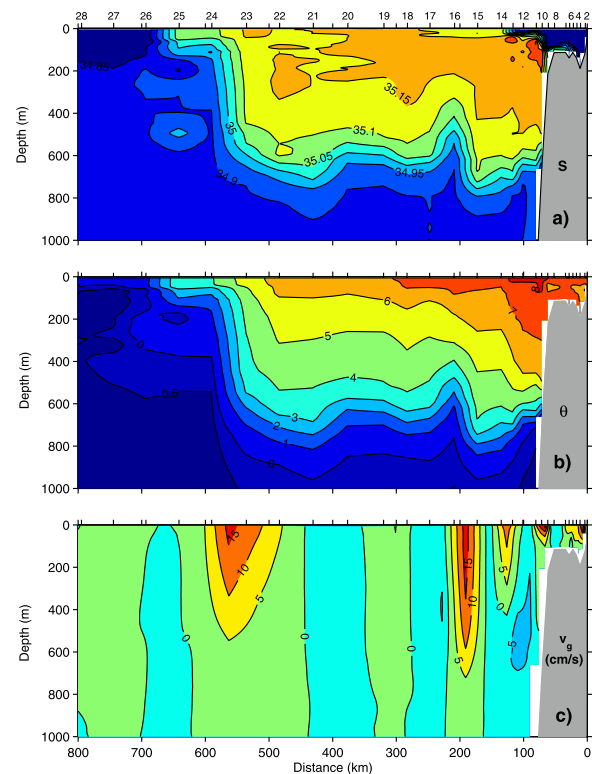


Figure 4. Gimsøy-NW section. A transect of temperature, salinity and current across the Lofoten Basin. The transect shows the wide, slab-like extension of AW, with the PF and associated jet over the eastern slope of the Mohn Ridge.

[13] Using $\Delta\rho = \rho_0 - \rho = 0.4 \text{ kg m}^{-3}$, and $H_0 = 500 \text{ m}$ in the SS [Mork and Blindheim, 2000] and $\Delta\rho = 0.25 \text{ kg m}^{-3}$, and $H_0 = 800 \text{ m}$ in the LB [Mauritzen, 1996; K.A. Mork, personal communication], the baroclinic transport $\frac{g'}{2\gamma} H_0^2$ becomes 3.8 Sv in the SS and 5.7 Sv in the LB. These transports are comparable with the estimates from hydrography, associated with the PF, i.e., 3.4 Sv in the SS and 5.0 Sv in the LB. To achieve maintenance of the transport, (3) shows that the difference has to be balanced by an oppositely-directed deeper flow v_0 , manifested through the second term volume flux of about 1 Sv; a positive contribution in the SS and a negative contribution in the LB.

[14] Current records in the SS show an average deeper flow associated with the frontal jet of the order of 5–10 cm/s [Orvik *et al.*, 2001]. Over the eastern MR, the average deeper countercurrent appears to be about 5 cm/s or less [Nøst and Isachsen, 2003]. Roughly, since the AW associated with the PF jet in the SS is 50 km wide and 400 m deep, an average deep flow of 5 cm/s [Orvik *et al.*, 2001] transports a volume of AW of about 1 Sv. In the LB over the MR where we have no current records, an oppositely-directed volume flux of order 1 Sv or even larger seems reasonable (Figure 4), and thus in accordance with the need to balance the model.

[15] As indicated by (3), the ocean appears to achieve maintenance of the volume flux through adjustment of the thickness (H_0) of the AW. The baroclinic term 1) can in principle account for the deepening of the AW northward from the SS ($\sim 500 \text{ m}$) to the LB ($\sim 800 \text{ m}$), as it gradually loses buoyancy. In fact, this is the trend for a poleward flow, in general. However, buoyancy loss cannot alone account for a deepening of the AW from 500 m to 800 m from the SS to the LB, and cannot possibly account for the shallowing farther north. This disagreement, again indicates the existence of a deeper, oppositely-directed flow, contributing to the second term in (3).

[16] Concerning the vertical extension of the AW farther northward toward Svalbard, where the buoyancy loss also is substantial, there is no deepening of the AW, quite the contrary with a shallowing (Figure 2). The peculiar variability of the hydrography with a shallowing of the AW and a decreasing baroclinic transport from LB (5 Sv) to the KR ($\sim 3 \text{ Sv}$), also justifies the necessity of an along current deeper flow along eastern KR to achieve balance in volume transport. In fact, this is the case, with an even stronger northward deeper current (above 5 cm/s) along the KR [Nøst and Isachsen, 2003].

[17] In a broad sense, for the PF jet, the ocean appears to achieve maintenance of volume transport of AW in spite of buoyancy loss and varying deeper currents, by adjusting the vertical extension (interface) of AW. When the AW flows from the SS toward the MR in the LB, it encounters a deeper countercurrent along the eastern MR, and appears to respond with a deepening of the AW (H_0 increases). This deepening of the AW flow can be interpreted as the jet in facing a deep countercurrent, slows down. Due to conservation of volume, a slowdown will cause piling up of water and thus a deepening of the AW (H_0 increases). This results in a new balance with a deeper AW, where maintenance of volume flux is achieved, in accordance with (3). As the AW flows farther northward along the MR, it will join a deeper along current, even stronger over the KR. An along deeper

current will ease the resistance to push the AW through the system, and the ocean seems again to respond in a similar way by shallowing the AW (H_0 decreases). The substantial shallowing of the AW according to Figure 2, and Mauritzen [1996], indicates a fairly strong deeper flow along the KR, also in agreement with [Nøst and Isachsen, 2003]. This switch over to the second term in (3), is also indicated by a decrease in the first term in (3), caused by both the loss of buoyancy and shallowing of the AW (decreasing H_0).

4. Concluding Remarks

[18] In this study we have shown how the topographically-steered deeper flow may affect the extension of AW, with emphasis on the deepening in the LB. We have demonstrated the effect by using an active reduced gravity model and analysis of data. Based on the good fit between the model and observations, the suggested mechanism appears to be credible and feasible. However, it must be noted that the data set is sparse, particularly in the interior of the NS, and small adjustments of the parameters may influence the fitting substantially. Thus the apparent coincidence between model and observations must be considered as suggestive. A more comprehensive study will be required included additional observations, to better understand the processes related to the PF and the varying vertical extension of the AW.

[19] In Orvik and Niiler [2003] the western branch of the NwAC was shown to be a continuous jet in the PF, guided along striking topographic features from Iceland to Svalbard. In this study we substantiate the importance of the deep topographically-steered flows, by demonstrating how they may affect the depth of the AW in the NS. In particular, we have concentrated on the deepening of the AW in the LB, presumably an important area both for climate and ecology. Referring to Svendsen *et al.* [1991], the deeper flow accounts for the topographic steering of the PF and for the extension of AW. Indeed the importance of the bottom topography cannot be exaggerated.

[20] **Acknowledgments.** This study is a contribution to the Svinøy section monitoring program, partly supported by NOClm. Thanks are due to Kjell Arne Mork and the Institute of Marine Research, for providing hydrographic data and performing data analysis, and for suggestions and careful reading of the manuscript by Martin Mork and Alastair Jenkins, and two anonymous reviewers.

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