GEOPHYSICAL RESEARCH LETTERS, VOL. 28, NO. 18, PAGES 3433-3436, SEPTEMBER 15, 2001

On the Mid-Depth Circulation in the Labrador and Irminger Seas

R. H. Käse, [1] A. Biastoch, [2] and D. B. Stammer [2]

Abstract. A numerical circulation model with 1/6° resolution and an accurate topography formulation explains details of the observed circulation in the Irminger and Labrador Seas that were recently revealed by *Lavender et al.* [2000]. We show that the recirculation pattern is established through a locally wind induced flow controlled by the bottom topography and enhanced through remote baroclinic forcing by the dense plume of Denmark Strait overflow water. The basic circulation is a robust feature in a hierarchy of model setups. It exists in the purely barotropic case driven by steady winds and is even maintained when realistic daily forcing is added. The narrow recirculation zone is manifested by a sea level depression spanning from the Denmark Strait across the Irminger into the Labrador Sea.

Introduction

Recent float-based observations in the Labrador and Irminger Seas [*Lavender et al.*, 2000, henceforth LDO2000] have revealed a previously undescribed mid-depth circulation pattern that depicts the subpolar gyre of the North Atlantic as a narrow structure interconnecting these regions by a band of alternating flow just downslope of the 3000 m depth contour. This flow pattern provides a path of fast communication between the basins of the subpolar gyre in the North Atlantic. A rapid spreading of water masses between the Labrador and Irminger basin was already present in the classical picture of the circulation, as seen in *Dietrich et al.* [1975] (e.g. their Figs. 10.47 and 10.49) and recently supported by measurements of fluorocarbon tracers [*Sy et al.*, 1997].

Ocean circulation models traditionally have problems simulating the circulation in the vicinity of strong topographic gradients, especially those with z-coordinate formulations. Indications of a realistic circulation in the subpolar area of the Atlantic are seen in *Smith et al.* [2000] and in the isopycnal and sigma coordinate models of the DYNAMO intercomparison study [*Willebrand et al.*, 2001]. Suggestions that not just horizontal resolution, but especially the representation of the bottom topography has a substantial impact on the flow simulations are provided by *Griffies et al.* [2001]. We will demonstrate here that an improved topographic representation has a significant effect on the simulated Labrador Sea flow field and that the resulting barotropic subpolar circulation is significantly enhanced by the baroclinic flow associated with the dense water plume originating in the Denmark Strait. Spreading and adjustment rates of the baroclinic flow depend on vorticity dynamics related to sloping topography and lead to a simulated circulation at 700 m (Fig. 1) that essentially shows LDO2000 observed characteristics.

The Model

The model is based on the MIT z-coordinate ocean circulation model [*Marshall et al.*, 1997] with 31 vertical levels and resolves the subpolar North Atlantic and parts of the Greenland Iceland Norwegian (GIN) Sea with 1/6° × 1/6° (i.e. approximately 18 km meridional and 9 km zonal resolution). In contrast to "traditional" primitive equations models, the bottom-most box, however, is allowed to be partially filled [*Adcroft et al.*, 1997], leading to a significantly improved representation of the topography at the given resolution (Fig. 2, insert). Our strategy in the design of the model experiments was as follows:

(A) Explore the role of bathymetry in barotropic runs without (exp. A1) and with (exp. A2) partially filled bottom cells under steady wind forcing.

(B) Take into account the effect of stratification with an idealized initial two water mass setup (given by 2 mean temperature profiles north and south of the Greenland-Scotland ridge (GSR) system determined from the Levitus climatology). We thus include the additional important polar/-

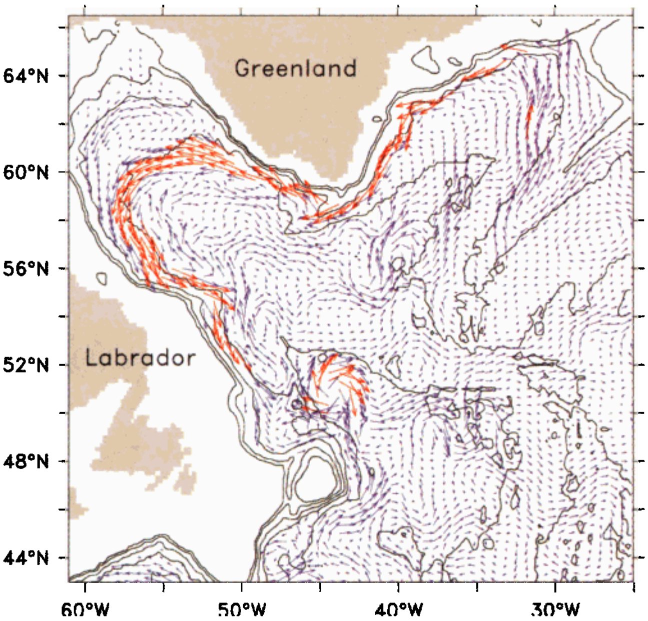

Figure 1. 1995 mean velocity vectors at 700 m depth in the model run with optimized bottom topography representation and daily surface forcing. Vectors are smoothed over the same horizontal scale as in *Lavender et al.* [2000], vectors greater than 10 cm/s scaled to equal length and marked red.

---

[1] Institut für Meereskunde Kiel, Germany.

[2] Scripps Institution of Oceanography, La Jolla, U.S.A.

Paper number 2001GL013192.
0094-8276/01/2001GL013192$05.00

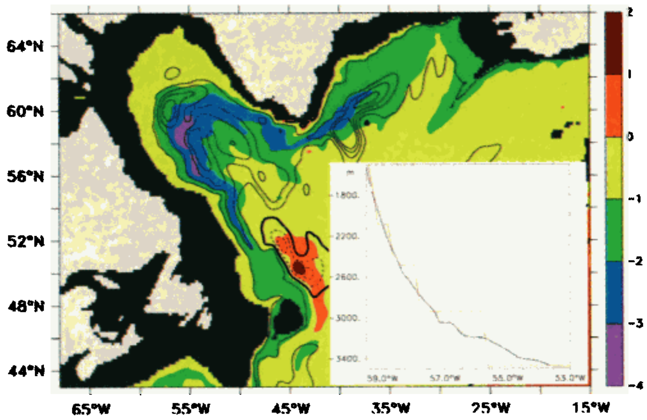

Figure 2. Barotropic case: sea surface height, in cm, averaged over one month of iteration for partially filled (exp. A2, color) and full bottom cells (exp. A1, isolines). The picture insert shows bathymetry profiles at 58° N for NDGC 5 arc-minute data (green), exp A1 (red) and exp. A2 (black).

subpolar exchange of cold water masses. A time mean wind forcing was applied and the bottom topography was represented without (exp. B1) and with (exp. B2) partially filled bottom cells. Because temperature is the controlling parameter that mainly determines the density in the subpolar gyre, we use a temperature - density relation as in *Käse and Oschlies* [2000].

(C) Apply full time dependent surface fluxes. The model is forced with the daily surface wind stress and heat and freshwater flux fields for the period 1992 - 1997 that emerged from an ocean assimilation approach [*Stammer et al.*, 2001]. The initial conditions were taken from climatology [*Levitus et al.*, 1994], against which the solution was restored in a 3° wide strip along the southern boundary.

Results

To compare the model results with LDO2000 we will focus mainly on the geostrophic pressure at the 700 m level. As expected, the barotropic wind driven circulation is established rather quickly, and within 3 months the western boundary current reaches a quasi-steady state (Fig. 2) which with partially filled cells looks strikingly similar to LDO2000.

The staircase-like bottom topography (Fig. 2, insert) leads to a broad-brushed subpolar gyre (contours in Fig. 2). With partially filled cells, a faster spreading of information along the topographic slope is evident. It confirms the analysis of *Pacanowski and Gnanadesikan* [1998] who improved topographic wave propagation by partially filled cells. The improved dynamics leads to a reduction of the cross-slope flow and thus reduces the cyclonic spin-up of the deeper parts of the basins. The similarity with the LDO2000 geostrophic pressure field estimation is obvious. Its magnitude, however, is too small.

A substantial part of the subpolar cyclonic gyre water enters as dense water flowing through the GSR system. We used exp. B2 with partially filled cells to established a first order estimate of the real circulation in the presence of stratification and dense water supply of North Atlantic Deep Water (NADW). This idealized baroclinic experiment establishes a preliminary quasi-steady state in the western boundary again after 3 months. But contrary to the previous pure barotropic simulation, the Ekman divergences and the GSR outflow now modulate the density field and the strength of the circulation on longer time scales. After one year, the subpolar gyre north of 53° N is practically established (Fig. 3a). In Fig. 3c we have sketched successive phases of the propagating overflow plume characterized by the plume arrival, while the insert figure shows a section of the barotropic stream function at 58° N with an equivalent color coding. Even after 5 months, the overflow plume has not yet reached the latitude of 58° N, and the western boundary current transport is only about 10 Sv. But after

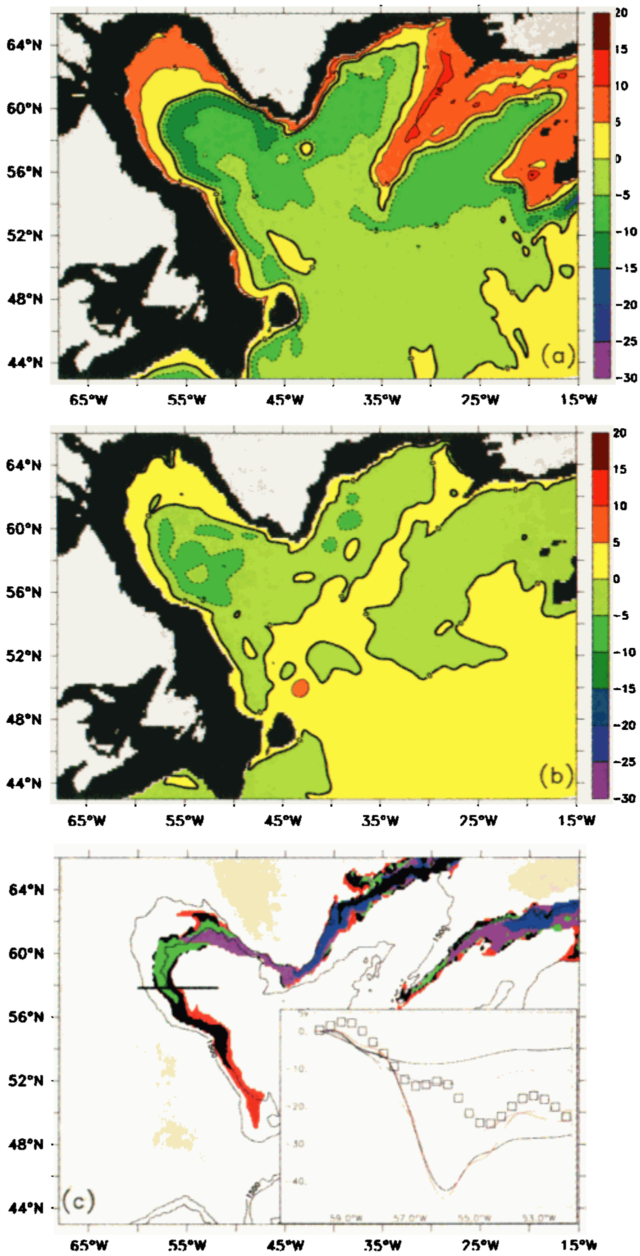

Figure 3. Baroclinic cases: (a) Geostrophic pressure at 700 m depth, in cm of water, averaged over month 13 of exp. B2. (b) As (a) but averaged over month 24 of exp. B1. (c) GIN Sea tracer tongue of exp. B2 after 3 (black), 5 (blue), 7 (purple), 9 (green), 11 (black) and 13 (red) months of integration. The smaller picture shows the transport profiles in Sv at 58° N at the same time steps. Shown is also the transport profile after 24 months of exp. B1 (black symbols).

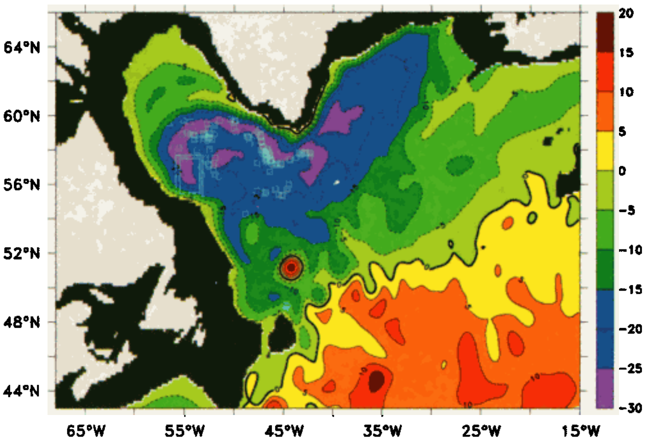

Figure 4. Full baroclinic case (exp. C): As (Fig. 3a) but averaged over Jan - Jun 1995. Positions after 6 months (light blue) of 700 m floats, initialized in a cluster (box).

7 to 9 months a drastic increase to 30 and 40 Sv occurs synchronously with arrival of the overflow plume. Finally, after the plume has reached Flemish Cap after 11 months, a substantial fraction of the boundary flow (about 20 Sv) is returned northward (red and black curves in insert of Fig. 3c). The overall appearance is that of the barotropic flow (exp. A1) but enhanced significantly in its amplitude due to the baroclinic forcing. With traditional staircase topography (exp. B1, Fig. 3b) the baroclinic overflow plume has significantly slower propagation. Moreover the enhanced eddy activity leads to a more diffusive flow and the plume is found at greater depth, thus at smaller topographic gradients. Note that the narrow recirculation is completely missing (squares in insert of Fig. 3c).

The fundamental flow pattern found under simplified conditions (Figs. 2 (color) and 3a) are not changed with realistic stratification and daily forcing. Results of this exp. C mean 700 m geostrophic pressure is shown in Fig. 4. Model floats have been included at several depth levels. The positions of simulated floats 6 month after they have been released in a cluster at 700 m depth near the most prominent site of winter time deep convection reveal that the cluster is dispersed in large part into the recirculation west and east of Greenland. Only a small fraction has taken the southern route in the fast western boundary current with cross-isobathal flow north of Flemish Cap, as described in LDO2000. These floats arrived 6 months later east of the Mid Atlantic Ridge or in the southeastern Irminger Sea west of the Reykjanes Ridge. Major sites of cross-slope flow are the region north of Flemish Cap at 48° N and also the area north of 55° N, both connected with sharp directional changes of the continental slope. The induced downslope flow initiates stretching of the water column and produces additional cyclonic vorticity, that leads to enhanced northward return flow characterizing the downslope flank of the geostrophic pressure anomaly.

The structure of the Labrador Sea boundary current of exp. C (Fig. 5a) reveals the doming of the isopycnals required for the observed reversal in a predominantly geostrophic flow [*Fischer and Schott*, 2001]. The geostrophic flow results from the combined pressure gradients due to surface elevation and the internal stratification. The latter accounts for 40% of the external part, but with opposite sign. Combined they lead to the total transport shown in Fig. 5b.

Conclusion

We have demonstrated that the observed flow structure in the Labrador Sea can be explained as a locally wind-driven circulation in the presence of topography interacting with the baroclinic flow field due to the Greenland-Scotland overflow. A necessary element in a realistic simulation of the Labrador Sea appears to be the proper representation of the bottom topography and the path of dense NADW water flow. Although the precise vertical extent of the dense water plume is not met in our model, the net vertically integrated density anomaly connected with it follows the correct path along the topography (Fig. 3c) as do perturbations in transport in the pure barotropic run. The strong topographic slopes surrounding the subpolar gyre have a significant im pact on the vorticity balance of the circulation. The classical theories of Stommel and Munk assumed rectangular basins with vertical walls. The vorticity acquired

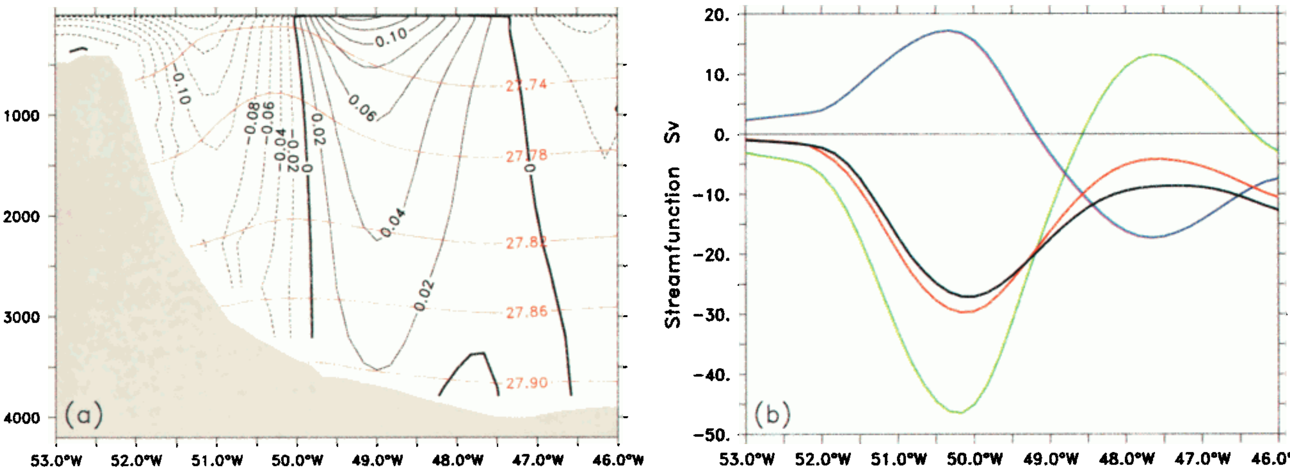

Figure 5. (a) 4-year mean meridional velocity and density section at 53° N. (b) Barotropic stream function at 53° N. Total transport (black), geostrophic (red), splitted in external (green) and internal (blue) parts.

in the Sverdrup regime needed to be dissipated in a narrow boundary current. More realistic theoretical approaches do not necessarily require frictional boundary layers [*Salmon*, 1994] but instead sloping shelfs. A further analysis in *Biastoch et al.* [2001] indeed verifies that the flow across f/H contours is balanced by the joint effect of baroclinicity and relief (JEBAR). Our results suggest a mechanism that provides a lagged response of the subpolar gyre to changes in overflow and results in a varying intensity of the recirculation cell. Since the recirculation of the western boundary current occurs at higher temperature, it would transport additional heat into the Labrador Sea and could influence the strength of the deep convection as well as the position of the North Atlantic current axis. As pointed out by a reviewer, a feedback loop in which the control of the Gulf Stream position by the deep western boundary transport at the crossover point produces decadal oscillations has recently been suggested by *Joyce et al.* [2000]. Modeling of these phenomena would clearly require an exact propagation along the topography which is achieved in our examples via partial bottom cells.

Acknowledgments. We thank A. Adcroft for support in setting up the model. RHK was supported through the German Research Society (SFB-460) and spent time at SDSC with partial support from ONR (NOPP ECCO) grant N00014-99-1-1049. AB and DS were supported by NASA grant NAG5-8623. Support by NCAR and NPACI computer resources is gratefully acknowledged.

References

Adcroft, A., C. Hill, and J. Marshall, Representation of topography by shaved cells in a height coordinate ocean model, *Mon. Wea. Rev.*, *125*, 2293 - 2315, 1997.

Biastoch, A., R. H. Käse, and D. Stammer, Exchange processes over the Greenland Scotland Ridge, 2001, in preparation.

Dietrich, G., K. Kalle, W. Krauß, and G. Siedler, *Allgemeine Meereskunde - Eine Einführung in die Ozeanographie*, third ed., Gebr. Bornträger, Berlin, Stuttgart, 1975.

Fischer, J., and F. A. Schott, Labrador Sea Water tracked by profiling floats from the boundary current into the open North Atlantic, *J. Phys. Oceanogr.*, *31*, 2001, in press.

Griffies, S. M., C. Böning, F. O. Bryan, E. P. Chassignet, R. Gerdes, H. Hasumi, A. Hirst, A. Treguier, and D. Webb, Developments in ocean climate modelling, *Ocean Modelling*, *2*, 123 - 192, 2001.

Joyce, T. M., C. Deser, and M. Spall, The relation between decadal variability of subtropical mode water and the North Atlantic Oscillation, *J. Climate*, *13*, 2550–2569, 2000.

Käse, R. H., and A. Oschlies, Flow through Denmark Strait, *J. Geophys. Res.*, *105*, 28,527 - 28,546, 2000.

Lavender, K. L., R. E. Davis, and W. B. Owens, Mid-depth recirculation observed on the interior Labrador and Irminger seas by direct velocity measurements, *Nature*, *407*, 66 - 69, 2000.

Levitus, S., R. Burgett, and T. Boyer, *Salinity. Vol. 3. World Ocean Atlas 1994, NOAA Atlas NESDIS*, U.S. Dep. of Comm., Washington, D.C., 1994.

Marshall, J., A. Adcroft, C. Hill, L. Perelman, and C. Heisey, A finite-volume, incompressible Navier Stokes model for studies of the ocean on parallel computers, *J. Geophys. Res.*, *102*, 5753 - 5766, 1997.

Pacanowski, R. C., and A. Gnanadesikan, Transient response in a z-level ocean model with bottom topography resolved using the method of partial cells, *Mon. Wea. Rev.*, *12*, 3248–3270, 1998.

Salmon, R., Generalized two-layer models of ocean circulation, *J. Mar. Res.*, *52*, 865 - 908, 1994.

Smith, R. D., M. E. Maltrud, F. O. Bryan, and M. W. Hecht, Numerical simulation of the North Atlantic Ocean at 1/10°, *J. Phys. Oceanogr.*, *30*, 1532 - 1561, 2000.

Stammer, D., C. Wunsch, R. Giering, C. Eckert, P. Heimbach, J. Marotzke, A. Adcroft, C. N. Hill, and J. Marshall, The global ocean circulation during 1992 - 1997, estimated from ocean observations and a general circulation model, 2001, submitted.

Sy, A., M. Rhein, J. R. N. Lazier, K. P. Koltermann, J. Meincke, A. Putzka, and M. Bersch, Surprisingly rapid spreading of newly formed intermediate waters across the North Atlantic Ocean, *Nature*, *386*, 675 - 679, 1997.

Willebrand, J., B. Barnier, C. Böning, C. Dieterich, P. D. Killworth, C. LeProvost, Y. Jia, J.-M. Molines, and A. L. New, Circulation characteristics in three eddy-permitting models of the North Atlantic, *Progress in Oceanography*, 2001, in press.

R. H. Käse, Institut für Meereskunde Kiel, Düsternbrooker Weg 20, 24105 Kiel, Germany (rkaese@ifm.uni-kiel.de)

A. Biastoch and D. B. Stammer, Scripps Institution of Oceanography, 9500 Gillman Dr., La Jolla, CA 92093-0230, U.S.A.

(Received March 16, 2001; revised June 6, 2001;
accepted June 28, 2001.)