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A model for the spreading and sinking of the Deep Ice Shelf Water in the Ross Sea

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Abstract: Spreading and sinking of the Deep Ice Shelf Water (DISW) in the Ross Sea are analysed using *in situ* observations and the results of a nonlinear, reduced gravity, layered numerical model, which is able to simulate the motion of a bottom trapped current over realistic topography. The model is forced by prescribing thickness and density of the DISW layer at the southern model boundary as well as ambient density stratification above the DISW layer. This density structure is imposed using hydrographic data acquired by the Italian PNRA-CLIMA project. In the model water of the quiescent ambient ocean is allowed to entrain in the active deep layer due to a simple entrainment parameterization. The importance of forcing the model with a realistic ambient density is demonstrated by carrying out a numerical simulation using an idealized ambient density. In a more realistic simulation the path and the density structure of the DISW vein flowing over the Challenger Basin are obtained and are found to be in good agreement with data. It is found that entrainment, which is particularly active in regions of strong topographic variation, significantly influences the pattern followed by the DISW layer. The evolution of the DISW layer beyond the continental shelf, i.e., in a region where the paucity of experimental data does not allow for a detailed description of the deep ocean dynamics, is also investigated.

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Introduction

The Ross Sea is a particular oceanographic environment where important water mass formation and transformation processes occur at several temporal and spatial scales. Interest in the processes occurring in the Ross Sea has increased in recent years largely due to an appreciation of the climatic relevance of this area, which plays a major role in the production of dense water at high southern latitudes (Jacobs *et al*. 1985, Jacobs & Comiso 1989, Locarnini 1994, Jacobs & Giulivi 1998, 1999, Hofmann & Klinck 1998). In fact, most of the dense deep water production in the Southern Ocean originates from the continental shelves of Antarctica. Flowing from oceanic regions next to the continental shelf they mix with the ambient water (typically less dense) as they flow down to the continental slope before reaching their equilibrium depth. If they reach the bottom of the slope they may become Antarctic Bottom Water (AABW) (Jacobs *et al*. 1970, Jacobs 1991, Hoffman & Klinck 1998, Bindoff *et al*. 2000). Many details about dense water formation on the Antarctic continental slope are still obscure, due to the difficulty in obtaining an adequate temporal and spatial resolution for atmospheric and oceanic phenomena related to the oceanic convection in the Southern Oceans from *in situ* data (a useful discussion on possible downslope mechanisms can be found in Baines & Condie 1998).

The Ross Sea, located in the Pacific sector of the Antarctic continental shelf, is characterized by a rather irregular topography (Fig. 1) with several reliefs and depressions, some of which are deeper than the shelf break depth (about 700 m deep). The Ross Ice Shelf (RIS), a broad ice cover which extends over nearly half of the continental shelf, marks the southern limit of the Ross Sea only for the near surface water layers, while deeper waters can flow freely between the bottom and the floating ice shelf (Jacobs *et al*. 1979). Reaching the continental shelf, the Circumpolar Deep Water (CDW) moves upward and mixes with the shelf waters forming the modified CDW (MCDW). The MCDW is the only water mass flowing over the continental shelf and therefore plays a fundamental role in the renewal of shelf waters and in the total heat budget of the Ross Sea (Budillon *et al*. 2000). The dense shelf waters in the Ross Sea are generally produced during winter, when the upper layers cool and freeze, releasing part of their saline content and, therefore, increasing the salinity of the subsurface waters. This process is frequently enhanced by the occurrence of strong katabatic winds in the Terra Nova Bay (TNB) Polynya, where a large amount of High Salinity Shelf Waters (HSSW) is therefore formed (Jacobs *et al*. 1985). Part of the HSSW is known to flow northward, towards the shelf break along the western sector of the Ross Sea (Budillon *et al*. 1999), and to take part in the formation of the AABW. Another branch moves southward under the

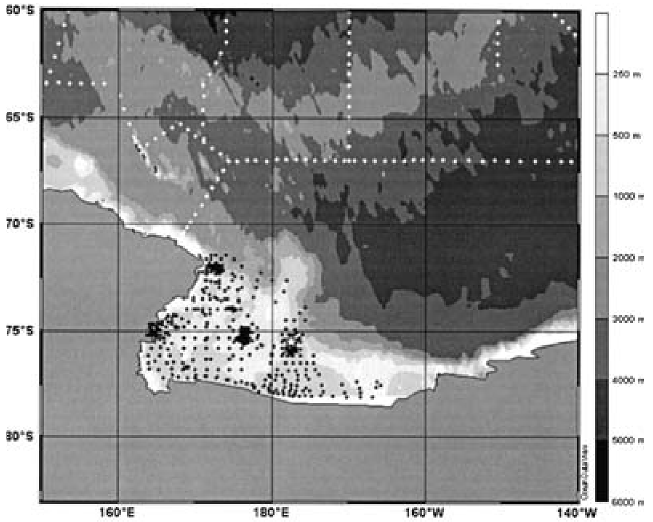

Fig. 1. Bathymetry of the integration domain. The stations used to determine the near-bottom ambient density are indicated. The location of the vertical section of Fig. 7 is also shown (white star).

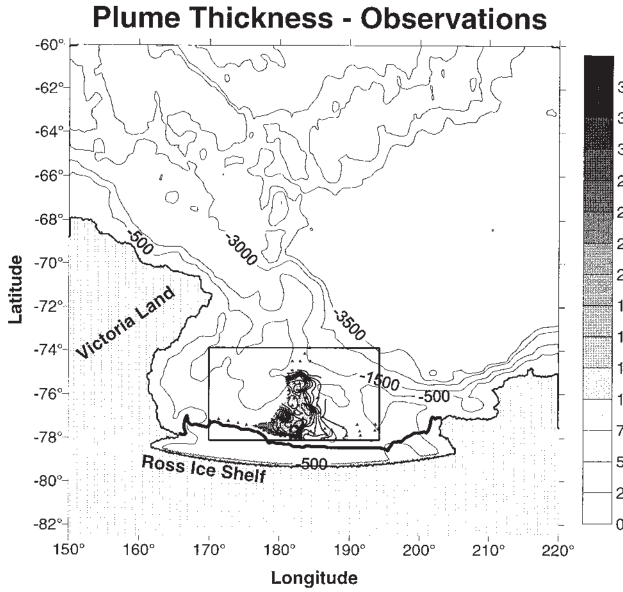

Fig. 2. Thickness of the DISW as measured during summer 1994/95. The values indicate the thickness of the water column where the measured potential temperature is colder than the corresponding freezing point at the sea surface. Triangles mark the locations wher hydrographical measurements used to trace the DISW vein were performed.

RIS, where it interacts with the basal ice giving rise to a different water named Deep Ice Shelf Water (DISW), which is characterized by a temperature lower than the freezing point at the sea surface. Cooling and ice melting at different depths under the RIS are the basic mechanisms for the formation of these waters. The DISW, which is located primarily on the central continental shelf, also moves northward (Budillon *et al*. 2002), giving a further contribution to the formation of the AABW (Jacobs *et al*. 1985, Jacobs & Comiso 1989).

In this paper we investigate aspects of the hydrodynamics of the Ross Sea related to the spreading and sinking of the DISW in two ways. First, by means of *in situ* observations and then by the results of a nonlinear, reduced gravity, frontal layered numerical model which is able to simulate the dynamics of isolated ("frontal") water masses and hence the dynamics of a bottom trapped current over realistic topography (Hessner *et al.* 2001, Rubino *et al.* 2002).

Observations

In the framework of the CLIMA project, as a part of the Italian National Research Program in Antarctica (PNRA), a detailed investigation of the thermohaline structure of the water mass in the central part of the Ross Sea was performed during summer 1994/95. At intermediate depths, the water column in this region is primarily constituted by the MCDW, which overlays the DISW and the HSSW (Jacobs *et al*. 1985, Trumbore *et al*. 1991). As the DISW is defined as the water with potential temperatures lower than the surface freezing point, its spatial distribution can easily be described by computing the layer thickness in locations where the measured potential temperature is colder than the one corresponding to the freezing point at the surface. The water mass distribution between 350 and 500 m depth traced according to this criterion is shown in Fig. 2. Close to the RIS, the DISW outflow is centred at about 177°W. It then moves northward and reaches the continental shelf break at about 75.5°S. At this location, evidence of the downslope flow of DISW was absent. This is probably due to the fact that, in this region, the bottom layer thickness decreases considerably while intense mixing takes place. This conjecture is confirmed by the fact that, during the high resolution measurements (with casts at 5 km) carried out in summer 1997/98, an active DISW downslope flow up to 1300 m depth was clearly identified.

The numerical model

A possible approach for carrying out a numerical study on sinking and spreading of the DISW would be to use a high resolution three-dimensional model, which would be able to describe in details features of bottom trapped currents as well as their interactions with the upper water layers. However, the computational effort implied by such an approach could be prohibitive. We have used a simpler approach instead, using a model based on the reduced gravity approximation that assumes in this case an infinitely thick upper layer overlaying an active bottom layer. The model solves the nonlinear hydrostatic shallow water equations and is equipped with a technique for the treatment of movable lateral boundaries. An entrainment parameterization accounts for the one-way mixing of water masses between the active layer and the overlaying quiescent ambient ocean. This parameterization was first proposed by Jungclaus & Backhaus (1994). The entrainment is a function of the plume horizontal velocity, thickness, reduced gravity, and turbulent Schmidt number (and hence of the Richardson number). Our model resembles the one used by Jungclaus & Backhaus (1994) for the description of the Denmark Strait Overflow. However, the algorithms used to simulate the movable lateral boundaries and the pressure gradient and advection terms are identical to those implemented by Hessner *et al*. (2001) and Rubino *et al*. (2002) for the description of a buoyant surface layer in a stratified and homogeneous ocean. The model equations are solved on an Arakawa C-grid on a very high resolution grid with $\Delta x = \Delta y = 5$ km, this being allowed by the simplicity of the model used. As far as the movable later boundaries are concerned, the bottom layer can be visualized as being constituted by an active layer for those points where DISW is present (i.e. where the layer thickness exceeds a critical value, typically a few centimetres) and by an inactive membrane elsewhere. This membrane may be lifted, thus enabling the inflow of DISW. When the water leaves a certain region, a track of its passage remains, made of a thin layer of inactive points. Further details about the numerical model can be found in Jungclaus & Backhaus (1994), Hessner *et al.* (2001), and Rubino *et al*. (2002).

The model has been initialized using the ambient density provided by CTD data obtained by measurements carried out in the CLIMA project. In Fig. 1 the locations of data are indicated. The near bottom, ambient lateral density structure was determined by considering, for each available station, mean salinity and temperature values representative of the deepest station point. Moreover, from the measured density, we computed a vertical density gradient, which was used to introduce a correction accounting for the decrease in the density contrast encountered at the bottom layer interface as its thickness increases. At the southern model boundary we imposed a fixed bottom layer displacement that, after a transient phase, yields a water transport of about 1 Sv (Bergamasco *et al*. 2002). We also imposed a fixed density "inflow" value corresponding to a realistic DISW value near the RIS oceanic region.

Numerical results

In a first, basic numerical experiment, we investigate the importance of choosing a realistic ambient density structure. If an ambient density depending merely on the water depth is assumed, then the resulting plume would yield important differences with respect to observations, as shown in Fig. 3, where the result of a 20 year simulation is presented. In this case, in the absence of a realistic ambient density structure, the DISW plume is free to expand westward, occupying the western part of the Ross Sea. This pictures is clearly unrealistic, as it is well known that the deep layers in the western sector of the Ross Sea are occupied by the denser HSSW, which constitutes, locally, a sort of barrier to the spreading of DISW (Jacobs *et al*. 1985, Locarnini 1994, Budillon *et al*. 2002).

In a run aimed at simulating more realistically the DISW deep flow an ambient density stratification inferred from *in situ* data was imposed, as explained in the previous

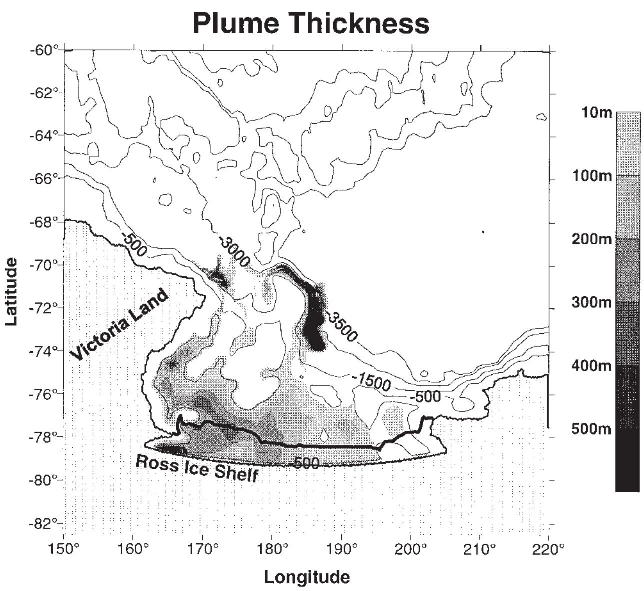

Fig. 3. Thickness of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using an idealized ambient density.

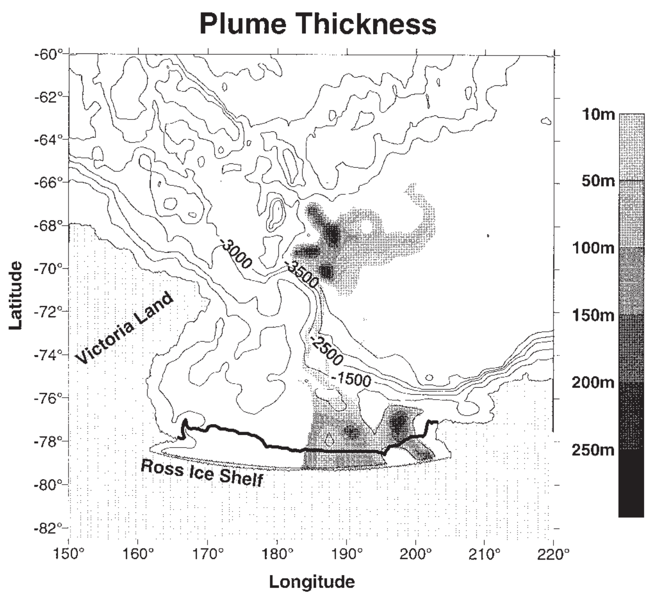

Fig. 4. Thickness of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using a realistic ambient density.

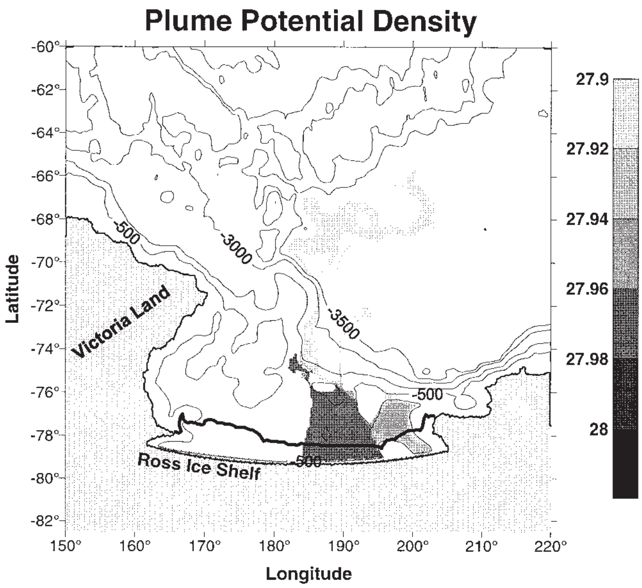

Fig. 5. Potential density of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using a realistic ambient density.

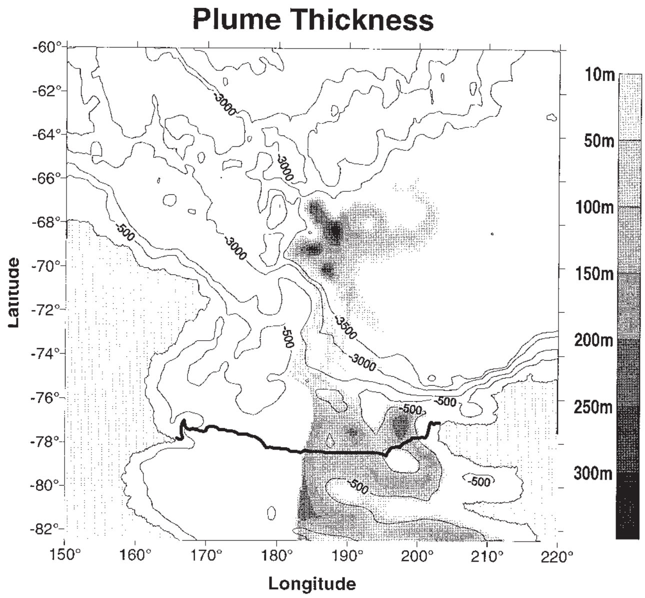

Fig. 6. Thickness of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using a realistic ambient density, and assuming that the southern model boundary is located at 82.5°S.

section. In Fig. 4 the plume thickness is shown after 20 years of integration. The plume horizontal distribution resembles the one of Fig. 2 inferred from the observed potential temperature. However, a spreading of the bottom trapped DISW plume is evident in an area east of 177°E over the shelf, which does not correspond to observed features. This may be due to inaccuracies in the bottom topography representation, leading to east–west basin connections that are not realistic, and/or to the absence, in the model, of the upper-ocean dynamics that may locally exert a deep influence on the spreading of the DISW bottom layer. It should also be noticed that, next to the RIS region, the DISW flows as an intermediate layer rather than as a bottom trapped layer. As the model is only able to simulate the dynamics of a bottom trapped layer, we refer necessarily, in this area, to a "hybrid" deep water mass that is the sum of the intermediate DISW and the denser (saltier and warmer) layer flowing beneath.

As the bottom layer reaches the shelf break, regions of very small layer thicknesses are encountered, where it appears that different bifurcations take place contributing to the DISW transport toward the abyssal region further north. Here, influenced by the local topography, there are different areas of water accumulation. The DISW layer spreads here zonally and then turns northward. Figure 5 shows the corresponding potential density of the plume at the same integration time. In general, very weak entrainment is simulated, with the exception of the shelf break area, where it occurs strongly in a very localized region that contributes substantially to the formation of the dense water of the abyssal plain located further north.

A third numerical experiment is aimed at investigating possible paths formed by the DISW under the RIS. To achieve this the integration domain was extended further south (Fig. 6). Thus we assumed, for this test, that DISW is present under the RIS, at about 183°W, as south as 82.5°S. In this case the results obtained agree only partially with the results inferred from the *in situ* observations. In fact, while the DISW northward spreading seems to be reproduced correctly, a relevant, complex eastward spreading is also simulated (Fig. 6) that does not correspond to observations. Again, this could be the result of inaccuracies in the bottom topography, or of the absence of an active upper layer in the model, but, more probably, it indicates that the assumed presence of DISW as south as 82.5°S is not realistic.

We now consider a comparison between model results and observations which refers to the behaviour of the DISW in the shelf break area. As discussed in the introduction, during the summer of 1997–98 a "mesoscale" experiment was carried out in the shaded region of Fig. 1 in order to investigate the flow in this area. Reaching the continental shelf break, the DISW (identified here by a potential temperature colder than -1.92°C) occupies a bottom layer of about 150 m thick (Fig. 7). A clear signature of the dense water overflow over the continental slope is found down to 1200 m, where a DISW layer with a thickness of about 70 m is present. Similar evidence was not detected in the other sections eastward of this transect. From these observations, and considering the absence of DISW at depths greater than 1300 m in section of Fig. 7, it seems likely that the overflow of this dense water may experience a strong westward deviation here as a consequence of its interaction with the overlaying westward slope current. However, the

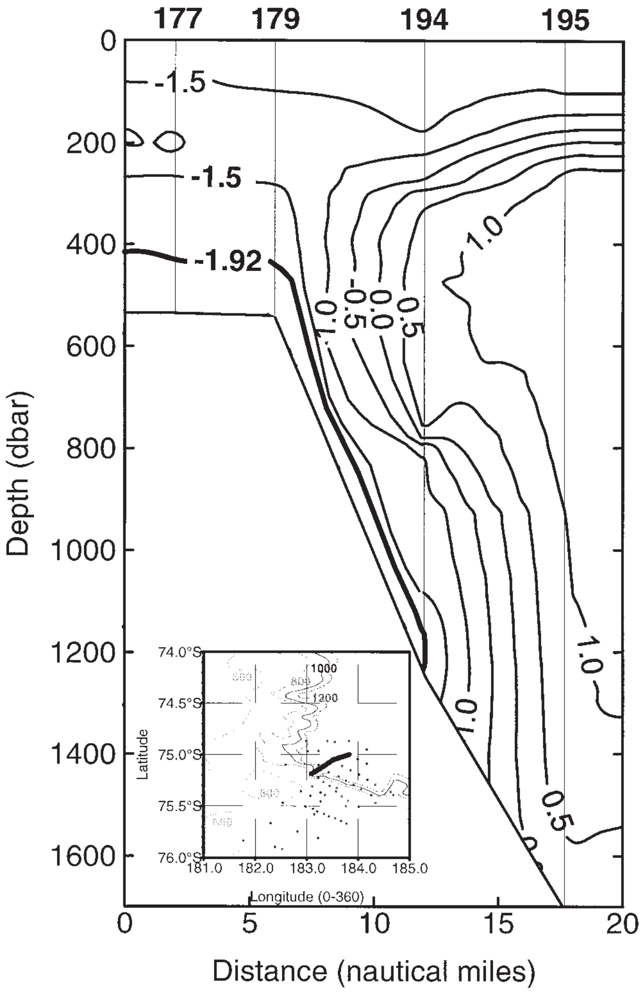

Fig. 7. Vertical section of potential temperature [°C] at the shelf break as measured during the "mesoscale" experiment in summer 1997/98. The sampling points and the section are reported in the small box (depths in metres).

hydrographical stations made on the western side of the survey region are too coarse to resolve this proposed circulation pattern.

The model results (Fig. 8) shows evidence of a large area where no bottom layer is present and also an area of strong entrainment, indicated by the strong attenuation of the bottom layer density. This behaviour can be further clarified by considering the simulated plume thickness. Figure 9 shows that while the dominant thickness is between 1 and 20 m at some locations it may reach 60 m. It must also be stressed that the depth of 1300 m, where the model yields the larger DISW layer thickness, is the same depth where the spilling of the DISW was identified during the "mesoscale" experiment. Therefore the model is able to reproduce the main features that have been observed to characterize the distribution of the DISW vein of bottom water over the shelf break. The abrupt disappearance of such a flow of DISW may be controlled by several mechanisms and physical constraints, such as the presence

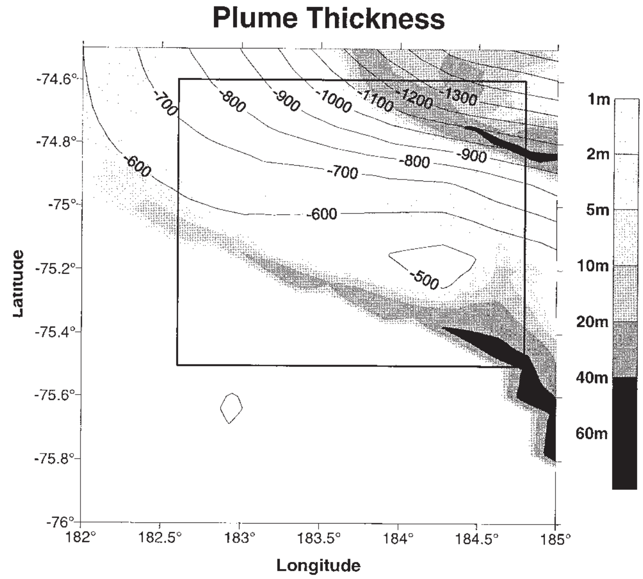

Fig. 8. Thickness of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using a realistic ambient density in the region where the "mesoscale" experiment (the box identifies a subsection of the region where this experiment was performed) was carried out during summer 1997/98.

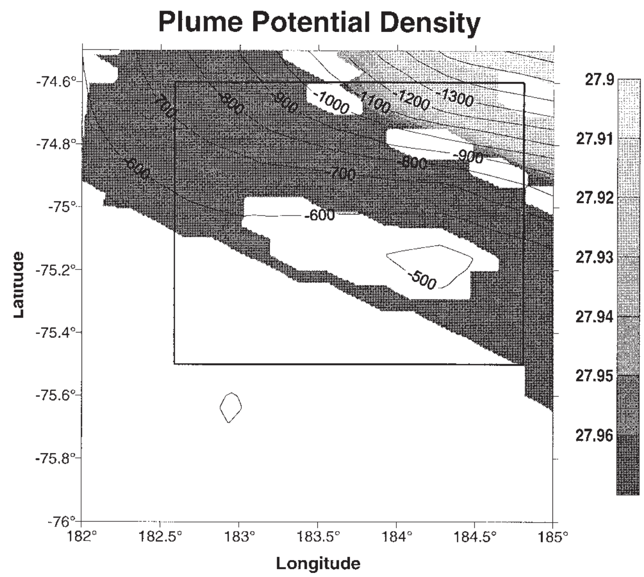

Fig. 9. Potential density of the DISW plume (after 20 years of integration) as a result of the numerical simulation carried out using a realistic ambient density in the region where the "mesoscale" experiment (the box identifies a subsection of the region where this experiment was performed) was carried out during summer 1997/98.

of canyons (the presence of rather irregular bathymetry can be deduced by the depth data acquired during the "mesoscale" experiment), vigorous upper layer currents, and interactions with the bottom particulate (Fohrmann *et al*. 1998, Budillon *et al*. 2002) which are not considered in the model.

Conclusions

In this paper we discussed aspects of the spreading and sinking of the DISW in the Ross Sea by analysing existing *in situ* data and by carrying out numerical simulations. The numerical model is a nonlinear, reduced gravity frontal numerical model that is able to simulate the dynamics of a bottom trapped current over realistic topography. It is based on the assumption that an infinitely thick, quiescent upper layer exists above the deep, active DISW layer and it computes the dilution of the bottom water density due to turbulent entrainment of water of the ambient ocean. Using this model, which is obviously not conceived for the description of the whole complexity of the near-bottom hydrodynamics in the Ross Sea, we were able to obtain results that are in good agreement with observations. In particular, the path followed by the DISW along its way northwards and the presence of regions of enhanced water entrainment and reduced layer thickness at the continental shelf break. The crucial role of the horizontal and vertical density field and of water entrainment in determining the DISW flow has been stressed. The density contrast between bottom layer and upper ocean determines the reduced gravity experienced by the vein of DISW, and hence, particularly in locations where no major topographic constrictions exist, it largely contributes to determine the path followed by the vein. The role played by water entrainment is subtle: it modifies the density of the bottom layer, increases its volume, and acts as a bottom friction. Our results indicate that, in regions of strong topographic variations, its role in diluting the DISW may be important. These regions can be thus considered as crucial ones in determining the fate of the DISW formed in the Ross Sea. Our investigation allowed us to suggest that the flow of DISW water does not exist at very high southern latitudes under the RIS, a region where *in situ* observations do not exist.

The approach of using a layered model to simulate bottom trapped currents in the Ross Sea seems, thus, to have been successful. The extension of this model to an (or more) upper active layer could however add valuable realism to the results.

Acknowledgement

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