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# Thermohaline structure and variability in the Terra Nova Bay polynya, Ross Sea

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**Abstract:** Hydrological measurements from three cruises during the summers 1994/95, 1995/96 and 1997/98 in the western sector of the Ross Sea allow summer and year to year changes in heat and salt content in the Terra Nova Bay polynya to be analysed. Changes in the surface layer (upper pycnocline) followed the expected seasonal pattern of warming and freshening from the beginning to the end of the summer. These near-surface changes, expressed as net heating and salting rate, were about 11 W $m^{-2}$ and -6 mg salt $m^{-2}$ $s^{-1}$. The heating changes were substantially lower than the estimated heat supplied by the atmosphere during the summer, which underlines the importance for this season of the advective component carried by the currents in the total heat budget. The year to year differences were about one or two orders of magnitude smaller than the seasonal changes in the surface layer. In the intermediate and deep layers, the summer heat and salt variability were of the same order as or one order higher than from one summer to the next. The differences in sign and magnitude for the heat change in the upper and in the lower pycnocline indicate a weak connection in the summer period between the surface heat fluxes and the deep waters. A local source of very cold water (with temperatures below the surface freezing point) of about 0.3 Sv has been detected close to the Terra Nova Bay coast. It arose out of the interaction of the shallow–intermediate layers of High Salinity Shelf Water with the coastal glaciers. The presence and the variability of this cold water point to the significant role of the thermohaline properties of Terra Nova Bay waters in controlling the floating glacier by governing the basal melting processes.

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**Key words:** air-sea fluxes, melting rate, polynya, Terra Nova Bay, thermohaline variability

## Introduction

The complex air–ice–sea interactions taking place in Antarctica affect the surface, intermediate and deep global ocean circulation, playing a major role in global change. The energy transfer (from and into the atmosphere) and energy balance develop over a variety of temporal scales. Strong heat loss by the sea and strong ice formation characterize coastal and continental shelf areas, where persistent polynya phenomena occur throughout the year. Moreover, in the Ross Sea, the water from the Antarctic Circumpolar Current is transformed into Antarctic bottom, deep, shelf and surface waters through complex processes of interaction with the atmosphere and with the glacial and sea ice. The intensity of the inflow and the outflow and the water mass conversion rate are clearly related to the seasonal and interannual fluctuations in surface wind stress and heat fluxes (Carmack 1977, Toole 1981, White & Peterson 1996, Jacobs & Giulivi 1998, 1999, Budillon *et al.* 2000a).

Interest in the processes occurring on the western continental shelf of the Ross Sea has increased in recent years largely due to an appreciation of the climatic relevance of polynyas, one of which appears regularly in Terra Nova Bay (TNB). The TNB polynya is a major source of High Salinity Shelf Water during the winter and can play a key role in the thermohaline characteristics of the Ross Sea (Bromwich & Kurtz 1984, Jacobs *et al.* 1985, Kurtz & Bromwich 1985, Jacobs & Cosimo 1989, Jacobs 1991, Van Woert 1999).

Last but not least, the behaviour of some of the floating glaciers in TNB has been hypothesized as also being the result of a change in melting at the ocean–ice interface due to a variability in the thermohaline properties of the water column in this region (Frezzotti & Mabin 1994, Frezzotti 1997).

The main aim of the present paper is to give a detailed description of the hydrological conditions of the Terra Nova Bay polynya as observed during the CLIMA (Climatic Longterm Interactions for the Mass-balance in Antarctica) cruises performed during three summers from 1994/95 to 1997/98 in the framework of the Italian National Research Project in Antarctica (PNRA - Programma Nazionale di Ricerche in Antartide). The good horizontal and vertical resolution of the CLIMA data-set permit a more complete census of the regional water masses, and contributes to our understanding of the TNB thermohaline variability and circulation.

In this paper we begin by describing the general circulation of the Ross Sea and its known thermohaline structure, before characterizing the typical summer "boundary conditions" (in terms of ice, currents and atmosphere) of the area studied. Using the hydrographic CLIMA data-set we describe the observed thermohaline field and its variability, before discussing the changes and the apparent correlation with some atmospheric forcings. Finally the evidence for the interaction of TNB waters with the ice shelves is analysed and quantified in terms of ice melting rate.

## The Ross Sea

One important dynamic feature of this sector of the Southern Ocean is the Antarctic Circumpolar Current (ACC) which moves from west to east around Antarctica and interacts with different water masses along its path. It carries the Circumpolar Deep Water (CDW), the most voluminous water mass of the Southern Ocean, extending down to a depth of about 4000 m. On reaching the Antarctic continental shelf the CDW moves upward in the water column and, at the continental shelf break, mixes with the shelf waters, forming bottom waters generically named Antarctic Bottom Waters (AABW) (Jacobs *et al.* 1985, Locarnini 1994, Gouretski 1999).

The mixing of the CDW with the surface and shelf waters forms a distinct water mass, the Modified Circumpolar Deep Water (MCDW), characterized by a subsurface potential temperature maximum and dissolved oxygen minimum. The MCDW is the only subsurface water mass reaching the Ross Sea shelf and plays a crucial role in the renewal of shelf waters and in the total heat budget (Jacobs *et al.* 1985, Trumbore *et al.* 1991, Locarnini 1994, Budillon *et al.* 2000a).

The shelf waters in the Ross Sea are generally formed during the winter when the upper layers cool and freeze, thus releasing part of their saline content, which increases the salinity of the subsurface waters (Jacobs *et al.* 1985). The high salinity of the western sector could be explained by taking into account the large extent of ice-free areas. Even during the winter period these are kept open by the strong winds, which blow the ice away as soon as it forms (e.g. Kurtz & Bromwich 1983, 1985, Jacobs *et al.* 1985). The rejected brine increases the salinity of the subsurface waters, thus forming the densest waters in the Southern Ocean. This water mass, with salinity increasing with depth from 34.75 to 35.00, is named High Salinity Shelf Waters (HSSW, as analysed by Jacobs *et al.* 1985). Part of the HSSW is known to move northward along the western sector of the Ross Sea as far as the continental shelf break and takes part in the formation of the AABW. Another branch flows southward under the Ross Ice Shelf, originating a different type of water named Deep Ice Shelf Water (DISW), which is characterized by a temperature lower than the freezing point at the sea surface. Cooling and melting at different depths under the Ross Ice Shelf basically forms these waters. DISW are located primarily on the central continental shelf; they also move northward, giving a further contribution to the formation of the Antarctic bottom waters (Jacobs *et al.* 1985, Trumbore *et al.* 1991, Bergamasco *et al.* unpublished data). Jacobs & Giulivi (1998, 1999) have recently presented a detailed and exhaustive general overview of the Ross Sea thermohaline characteristics, variability and circulation.

The Ross Sea is thus a well-known site of dense water formation associated with winter surface heat losses. The dense water formation process is likely to be extremely

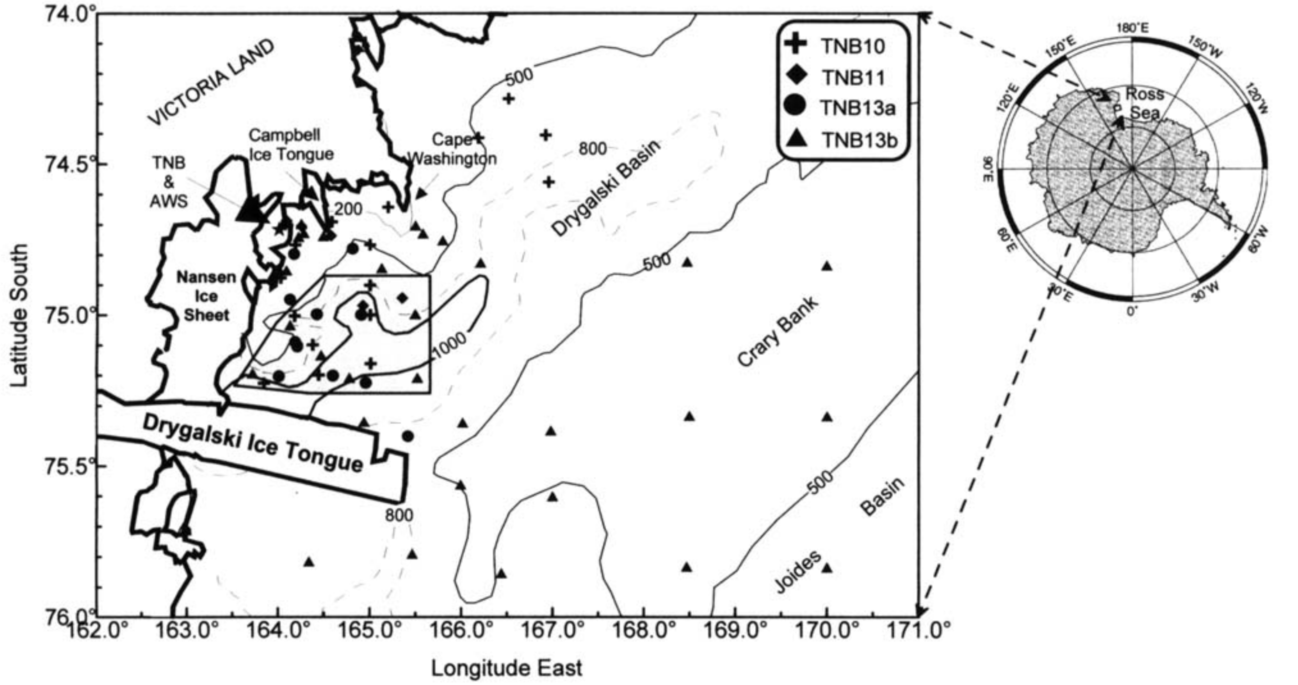


**Fig. 1.** Distribution of the hydrological stations, superimposed on the 200 m, 500 m, 800 m and 1000 m bathymetric contours. Separate symbols are used for different cruises. The locations of the Terra Nova Bay (TNB) Italian station and the Automatic Weather Station (AWS) are also shown. The coastline and grounding lines are from Frezzotti (personal communications 2000)).

sensitive to the inter-annual variability of the atmospheric forcing making different amounts of dense water available for the bottom water formation processes and for the ventilation of the deep layers outside the continental shelf break.

## The Terra Nova Bay polynya

### *Physical setting*

The western sector of the Ross Sea is characterized by several banks and depressions, the latter deeper than the continental shelf edge. The bottom topography of the zone (Fig. 1) is rather irregular with a shelf of about 500 m separated from the shore by a depression, the Drygalski Basin, over 1000 m deep. This basin has a SWW axis, the Mawson Bank (not shown) and Crary Bank (the sill depth is approximately 250 m) separate it from the Joides Basin (which has the same orientation).

Terra Nova Bay is bordered to the south by the Drygalski Ice Tongue (DIT) that forms part of the coastline for approximately 50 km. In the northern part, the Campbell Ice Tongue extends in a south-east direction. The Nansen Ice Sheet (NIS) (even if it is technically an ice shelf, has been so called by the first explorers; Frezzotti *et al.* in press), is located along the western coast, it shows a mall ice thickness of the ice front between 34–63 m.

### *Meteorological forcing*

Kurtz & Bromwich (1985) reported on an automatic weather station (AWS) that had only operated, at that time, for a few months on the southern tip of Inexpressible Island along the western coast of TNB. They concluded that the combination of strong and persistent katabatic winds coupled with the blocking effect of the DIT is necessary to form and maintain the TNB polynya. More recently, Van Woert (1999) applied a one-dimensional coastal polynya model forced by data from an AWS in order to examine winter ice concentration fluctuations. He argued that not only is the sensible heat flux important in controlling fluctuations in polynya extent, but that the longwave heat flux is significant. In that study, monthly mean and standard deviations based on 3-hourly meteorological observations have underlined that the winter wind values (13 m $s^{-1}$) used by Kurtz & Bromwich (1985) in their flux estimations were substantially underestimated. Thus, the wintertime surface heat fluxes obtained by Van Woert are higher than those reported in Kurtz & Bromwich. Winter mean sensible heat flux ranges from 727 to 836 W $m^{-2}$, latent heat and longwave are almost regularly higher than 240 W $m^{-2}$ and 120 W $m^{-2}$, respectively. The winter mean net heat flux ranges from 1096 to 1233 W $m^{-2}$ in the years 1988–90, with an estimated ice production of 46–52 $km^3$ and a salt rejection of 1.0–1.1 $*10^{12}$ kg. To maintain a conversion of WMCO or LSSW to HSSW consistent with the observed ice production and salt rejection rates, a transport of 1 Sv of HSSW from TNB was estimated. The annual ice production is practically independent of the net heat flux, depending almost entirely on the surface wind speed (Van Woert 1999).

### *Sea ice*

Kurtz & Bromwich (1985) analysed satellite visible and infrared images from the National Snow and Ice Data Center, concluding that the mean polynya area is 1300 $km^2$ with a high variability ranging from 0 to 5000 $km^2$. The maximum polynya extension corresponds roughly to an offshore size equal to the length of the DIT (about 80 km, at that time).

During the sampling periods the area studied may be

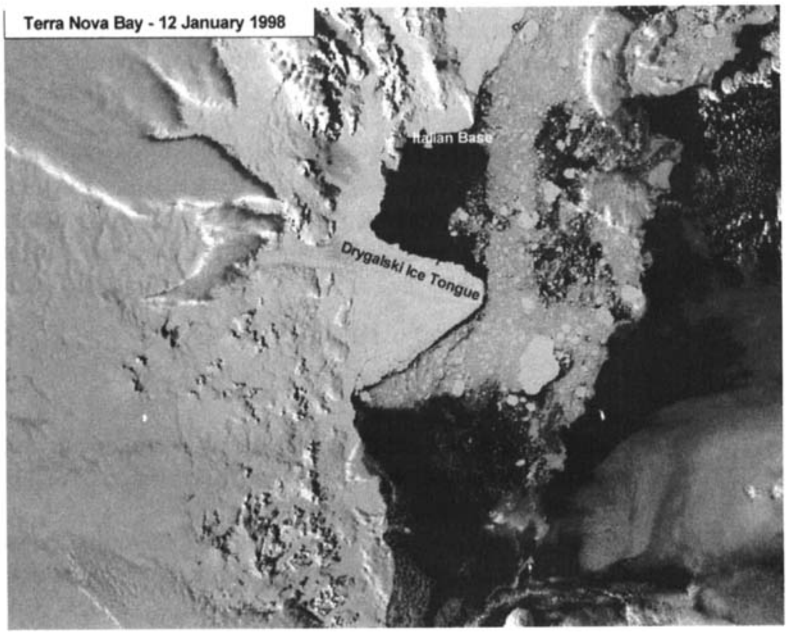


**Fig. 2.** Satellite visible image of Terra Nova Bay 12 January 1998.

considered free of ice, even if the presence of some thin or loose pack can be detected offshore. A typical situation is summarized by the satellite image of Fig. 2 taken on 12 January 1998. The interior of the TNB is completely ice-free and only a band of thin and fragmented pack is visible outside the bay. This situation has been observed several times and may be derived from the interaction of the northward drift of the ice and the local offshore wind forcing.

*Circulation*

The general circulation in TNB was recently analysed by means of current time series and modelling studies. Picco *et al.* (1998, 1999) reported on the data collected at a mooring located in the deepest TNB basin ($\varphi$ = 75°06.10'S, $\lambda$ = 164°13.04'E, depth = 912 m). During the measurement period (February 1995–January 1996), currents from the surface (55 m) to the intermediate layers (140 m and 402 m) are mainly directed north-east along the coast, while in the deeper layer (748 m and 882 m) the flow is essentially directed eastward. Currents are intense, especially at 55 m where more than 40% of daily values are higher than 15 cm $s^{-1}$ and display peaks higher than 50 cm $s^{-1}$. Near bottom currents, which are responsible for the HSSW transport, are also quite high, with an annual average of around 5 cm $s^{-1}$.

The vertically integrated transport of TNB has been investigated by Commodari & Pierini (1999) by means of a high-resolution model nested in a barotropic primitive equation model of the Ross Sea driven by the locally wind forcing and by the external action of the East Wind Drift. The stationary circulation field obtained in TNB was characterized by a cyclonic structure forming a strong coastal jet. The effect of the local wind was found to be negligible compared with the boundary basin scale forcing.

## Data sources and methods

*Sampling, calibration and processing*

The results presented in this paper originate from three quasi-synoptic cruises carried out by the RV *Italica* in the TNB area in the framework of the basin scale surveys conducted by the Italian CLIMA project in the Ross Sea. The first and second cruises (TNB10 and TNB11 respectively) were carried out during the summers 1994/95 and 1995/96. These expeditions were devoted to investigating the spreading of the DISW in the central part of the Ross Sea; consequently only a few stations were performed in the western sector.

In the summer 1996/97 no oceanographic survey was carried out in the Ross Sea by the CLIMA project. The last cruise, named TNB13, has been separated into two legs performed at the beginning (TNB13a) and at the end (TNB13b) of the summer 1997/98 (see Table I). During this cruise more emphasis was devoted to the western sector of the Ross Sea resulting in a better coverage in the TNB polynya (Fig. 1).

Conductivity-temperature-depth (CTD) profiles were performed with a SeaBird Electronics 9/11 plus instrument equipped with a standard sensor set which included a double temperature–conductivity duct for optimized spike reduction by defined time alignment between temperature and conductivity measurements. The CTD sensors were calibrated before each cruise at the SACLANT CENTRE (NATO) of La Spezia (Italy); a post cruise calibration was performed only after the TNB10 expedition. Additional checks were performed on board by means of two SIS RTM4200 digital reversing platinum thermometers; at every station several samples of water at different depths were collected and analysed on board by means of an Autosal Guildline 6400 Salinometer, but only the bottom samples were used for the conductivity calibration.

According to the pre- and post-cruise calibrations, an offset of -0.0028°C was applied to the raw data of the TNB10 cruise. For the same data-set, an offset of 0.00095 $S^{-1}$ m was applied to the conductivity measurements. It was not necessary to apply corrections to the temperature and conductivity data collected during the other cruises because the on board checks showed no significant drift in both sensors. The data were processed by standard methods using SeaBird software and averaged over 1 dbar in the final data-set.

The summer cruises TNB10, TNB11 and TNB13a covered essentially the same part of the region; the TNB13b cruise covered a larger area. The differences in the area covered and the sampling resolution for the four cruises were due to meteorological conditions, sea ice cover and available ship time.

A total of 51 CTD stations were sampled during the CLIMA expeditions in Terra Nova Bay from the summer 1994/95 to 1997/98 (Fig. 1 & Table I). Simultaneous standard meteorological parameters were recorded by an automatic weather station located at the Italian Terra Nova Bay station (Fig. 1).

**Table I.** Acronyms, date and number of the cast used in this study. Figure in parenthesis indicate the number of casts in the shaded area of Fig. 1 used for heat and salt flux computations.

| Survey | Summer | Acronym | Period | Total casts |
|---|---|---|---|---|
| X PNRA | 1994–95 | TNB10 | 18–20 Feb 1995 | 16 (7) |
| XI PNRA | 1995–96 | TNB11 | 02–05 Feb 1996 | 7 (4) |
| XIII PNRA | 1997–98 | TNB13a | 07–09 Dec 1997 | 11 (8) |
| | | TNB13b | 08–13 Feb 1998 | 12 (5) |
| | | | 22–23 Feb 1998 | 5 |

*Heat and salt changes*

The temperature and salinity changes between cruises can be obtained in terms of the heat and salt flux divergence required producing the observed change.

At each depth, values of heating rate (Q in units of W $m^{-3}$) and salting rate (R in units of g $m^{-3}$ $s^{-1}$) are computed in this work as:

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$$R = \frac{\rho \Delta S}{\Delta t} \quad (1)$$

$$Q = \frac{\rho C_p \Delta T}{\Delta t} \quad (2)$$

where ρ is the reference density (1028 kg m$^{-3}$), $C_p$ is the specific heat (3980 J kg$^{-1}$ K$^{-1}$), ΔT and ΔS are respectively the temperature (°C) and salinity change, and Δt is the time interval (s) between measurements (cf. Klink 1998).

These calculations were made for some selected stations chosen in the deepest region of the area studied (within the shaded region of Fig. 1) in order to examine a homogeneous data-set in terms of number and location of casts. For the first cruise analysed (TNB10) seven casts have been considered, four casts for the TNB11, eight and five for the TNB13a and TNB13b, respectively (Table I). For each survey, these casts were averaged in order to obtain a standard thermohaline profile representative of the investigated period (Fig. 3). The remaining casts have been analysed to describe the three-dimensional thermohaline field and some observed peculiarities of the TNB oceanography.

The lower limit of meaningful heat and salt fluxes is established by the accuracy of CTD sensors. An accuracy of 0.005°C, which is appropriate for the SeaBird thermistor, is equivalent to a heat change of 6.5 10$^{-4}$ W m$^{-3}$ over one year (365 days). Similarly, a year to year salinity change of 0.005 is equivalent to 1.6 10$^{-7}$ g m$^{-3}$ s$^{-1}$.

It must be stressed that the vertical integral of the flux divergence is equivalent to the net heat or salt flux over a water column with an area of 1 m$^2$ and extending over the depth range considered. These integrated fluxes can be thought of as the net surface flux required to make the observed changes. This conversion is used merely for the sake of interpretation and we do not consider that all measured changes are only due to fluxes at the ocean–atmosphere interface. As a convenient conversion we note that a salt flux of ±1 mg salt m$^{-2}$ s$^{-1}$ is equivalent to a daily sea ice production of 3 mm of ice (assuming water and ice salinities of 34 and 5, respectively) or a precipitation of 2.5 mm day$^{-1}$ (Klink 1998).

In the western Ross Sea the greatest variability in the water column occurs in the upper 100 m, primarily due to the influence of the atmospheric forcing (Jacobs *et al.* 1985, Locarnini 1994, Budillon *et al.* 1999). Moreover, we must also consider the high frequency variability (e.g. internal waves) that may be interpreted as a seasonal change of heat and salt. To some extent, integrating the changes over depth can eliminate this bias. Therefore, to study the interannual variability the heat and salt changes have been computed and averaged with a step of 100 m. The importance of the high frequency changes has been investigated with stations that have been re-sampled within a few hours during the same cruise. For these repeated casts, the temperature and salinity

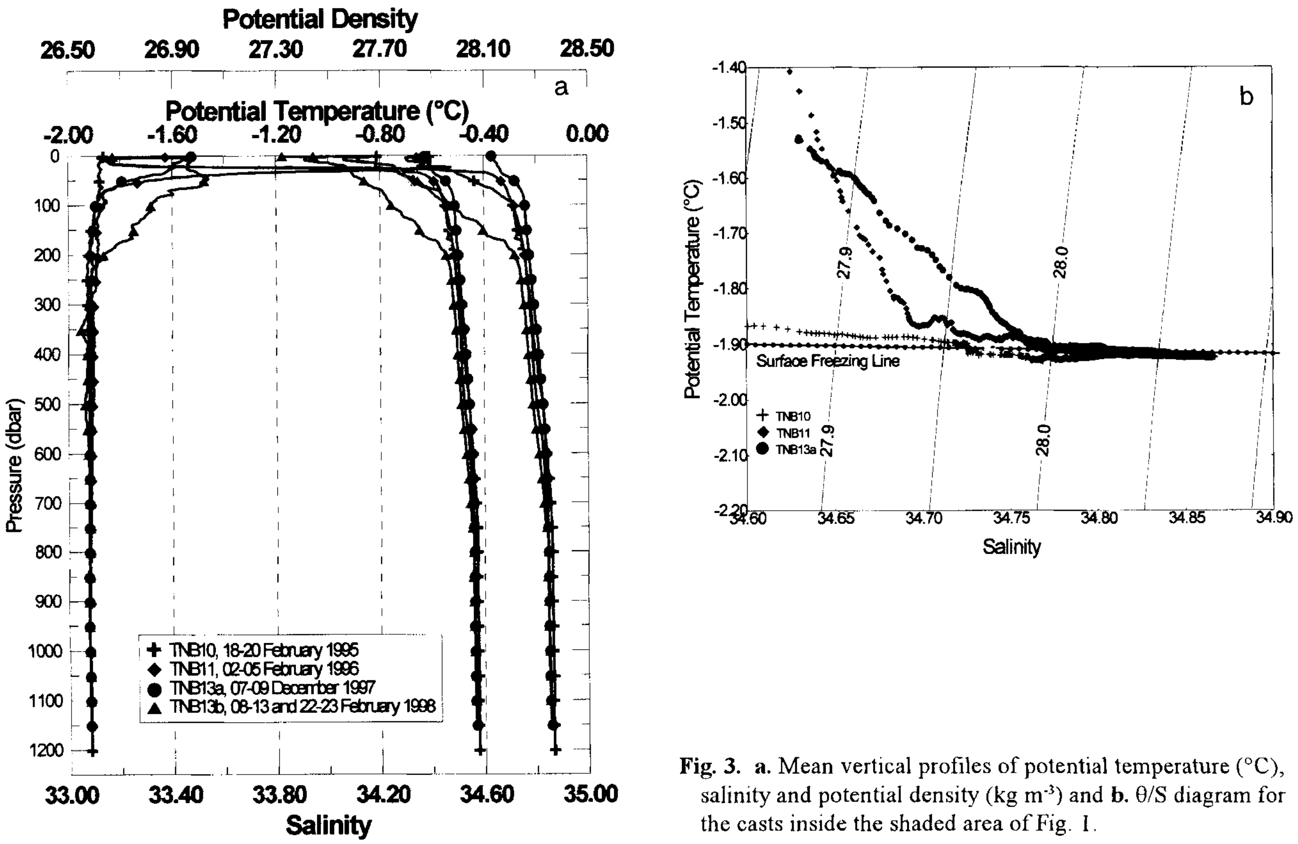


**Fig. 3.** **a.** Mean vertical profiles of potential temperature (°C), salinity and potential density (kg m$^{-3}$) and **b.** θ/S diagram for the casts inside the shaded area of Fig. 1.

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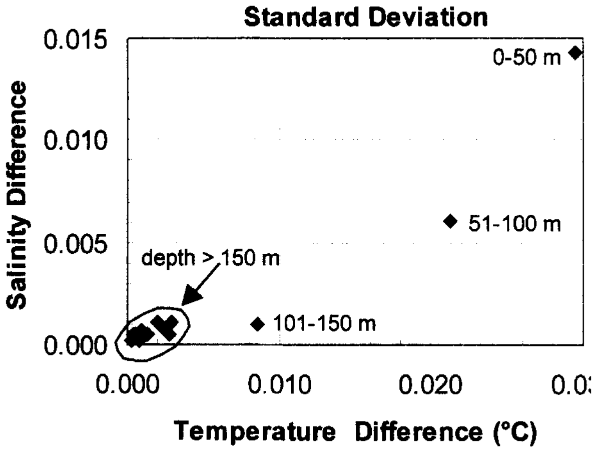


**Fig. 4.** Plot of the potential temperature and salinity standard deviations for the repeated casts in December 1997. They have been calculated for depth intervals of 50 m. The averaged depth is only shown for the upper layers .

differences have been calculated from the surface to the bottom and the standard deviation (s d) averaged for each layer of 50 m is shown in Fig. 4. The most important variability is observed in the upper layer (up to 150 m) while below it (and down to the bottom) both the salinity and potential temperature s d are the same order of the lower limit of meaningful readings of the CTD sensors. This analysis shows that high-frequency variations are not a crucial source of aliasing in the study of the heat and salt variability in the subsurface layers (below the thermocline) but could be relevant in the surface layer.

On the other hand, the effect of mesoscale variability, with time scales of few days and weeks, could also introduce a sort of aliasing in our data that is hard to investigate and quantify with our hydrographic data-set. Thermohaline water properties in TNB were measured on a number of occasions and were always found to have similar properties except for seasonal changes in the upper pycnocline. A time series recorded recently in TNB at a different depth by a mooring located in the deepest sector (see previous section on Terra Nova Bay circulation) have recently been graphically reported by Manzella *et al.* (1999) and showed a relevant variability in the upper layer (140 m) and less remarkable changes in the intermediate and deep layer. Over the entire water column the variability in less enhanced in the summer period (December–February) in comparison to the other months.

In this work we have re-processed this moored time series to calculate the temperature and salinity changes in the summer season (data courtesy of Meloni 1999). The Aanderaa temperature sensors (arctic range) were located at four different depths. Temperature and conductivity measured by a SeaCat16 instrument were recorded at approximately 750 m.

During the summer 1995/96 the mooring was recovered and deployed with the instruments at the same depths. Table II reports the weekly averaged temperature value and the standard deviation at different depths for different periods during the summer 1995/96. The same analysis has been performed for the salinity at the depth of 747 m. Most of the variability is restricted to the upper layer (140 m); below it the characteristic magnitude of the variability is one order less. The results are below the lower limit of heat and salt fluxes that can be measured within the accuracy of CTD sensors. This analysis also shows that mesoscale variability during this period is not an important source of aliasing to a study of the heat and salt variability in the subsurface layers (below the thermocline) but is a not negligible component in the surface layer.

## Results

### *Water masses*

The potential temperature (θ) and salinity (S) measurements obtained from CTD casts inside the shaded area of Fig. 1 were used to construct the vertical profiles and the θ/S diagram

**Table II.** Weekly average and standard deviation (s d) for salinity (S) and temperature (T) time series recorded at different depths in the central region of Terra Nova Bay.

| | 140 m (Aanderaa) | | 400 m (Aanderaa) | | 747 m (Seacat) | | | | 882 m (Aanderaa) | |
|---|---|---|---|---|---|---|---|---|---|---|
| Week | T | s d | T | s d | T | s d | S | s d | T | s d |
| 03–09 Dec 95 | -1.86 | 0.260 | -1.89 | 0.006 | -1.906 | 0.001 | 34.83 | 0.001 | -1.88 | 0.002 |
| 10–16 Dec 95 | -1.84 | 0.036 | -1.88 | 0.007 | -1.906 | 0.001 | 34.83 | 0.001 | -1.88 | 0.002 |
| 17–23 Dec 95 | -1.86 | 0.025 | -1.89 | 0.007 | -1.907 | 0.001 | 34.82 | 0.002 | -1.88 | 0.003 |
| 24–30 Dec 95 | -1.89 | 0.028 | -1.88 | 0.006 | -1.906 | 0.002 | 34.82 | 0.004 | -1.88 | 0.002 |
| 31 Dec–06 Jan 96 | -1.90 | 0.031 | -1.88 | 0.003 | -1.904 | 0.001 | 34.81 | 0.002 | -1.88 | 0.003 |
| 07–13 Jan 96 | -1.8 | 0.0318 | -1.88 | 0.006 | -1.907 | 0.001 | 34.820 | 0.003 | -1.88 | 0.002 |
| | | | 586 m | | | | | | 835 m | |
| 2–8 Feb 96 | -1.89 | 0.016 | -1.89 | 0.004 | | | | | -1.89 | 0.003 |
| 9–15 Feb 96 | -1.90 | 0.014 | -1.90 | 0.004 | | | | | -1.89 | 0.003 |
| 16–22 Feb 96 | -1.91 | 0.024 | -1.90 | 0.004 | | | | | -1.89 | 0.003 |
| 23–29 Feb 96 | -1.88 | 0.020 | -1.89 | 0.006 | | | | | -1.89 | 0.003 |

Accuracy was 0.05°C for Aanderaa, 0.01°C and 0.001 S $m^{-1}$ for Seacat.

shown in Fig. 3. The vertical structure of the water column in TNB polynya is relatively simple. The greatest variability is observed in the surface layer that extends to a depth of 50–150 m. Below it the water column is nearly isothermal, with its vertical stability preserved by the increasing salinity. The vertical structure of the summer water column in TNB may therefore be considered as being two-layered. In the 1997/98 season these layers were separated by a pycnocline that occupied different depths at the beginning and at the end of the summer (deepening from December to February from 50–100 m to 150 m, Fig. 3a). The Summer Surface Water (SSW) occupies the upper layer which, during summer, becomes fresher and warmer because it is influenced by sea ice melting and by the heat gain from solar radiation which is the main constituent of the surface heat balance in summer in this region (Budillon *et al.* 2000a). Below this layer the water column is occupied by the HSSW with a potential temperature close to the surface freezing point and S > 34.7 (Fig. 3, Table III).

In this quasi-isothermal layer of HSSW, a water mass showing temperatures of less than the surface freezing point ($T_f$ = -1.914°C for $P$ = 0 dbar and S = 34.85, Fofonoff & Millard 1983) has been detected at intermediate depths; this is here called TISW (Terra Nova Bay Ice Shelf Water) and it has been observed over the whole area of TNB even though characterized by different depths and thickness. A similar water mass (DISW) is well known in the southern sector of the Ross Sea, emerging beneath the Ross Ice Shelf and spreading to the shelf break and playing an important role in deep water formation; in the same paper Jacobs *et al.* (1985) mentioned the presence of a similar DISW in the TNB area referring to unpublished data.

The CLIMA data-set allows us, for the first time, to define the extent and properties of this ISW in TNB. Following the definition of the ISW, a water mass characterized by temperature below the surface freezing point, we calculated for each station the maximum of the difference between the *in situ* potential temperature and the surface freezing point $T_f$ (with S = 34.85) for the most extensive cruises in TNB (TNB10, TNB13a and TNB13b). This non-horizontal distribution (referred to a mean depth of 330 m ± 170 s d, 490 m ± 320 s d and 450 m ± 200 s d for TNB10, TNB13a and TNB13b respectively) is reported in Fig. 5. The extent of the TISW is clear, and negative anomaly values indicate that most

**Table III.** The major water masses that occur during the summer season in the Terra Nova Bay area and their definitions in terms of potential temperature and salinity.

| Name | Acronym | potential temperature (°C) | salinity |
|---|---|---|---|
| Summer surface water | SSW | -1.7 to 1.5°C | 33.00 to 34.50 |
| Terra Nova Bay ice shelf water | TISW | < -1.92°C | 34.74 to 34.82 |
| High salinity shelf water | HSSW | -1.91°C | > 34.75 |

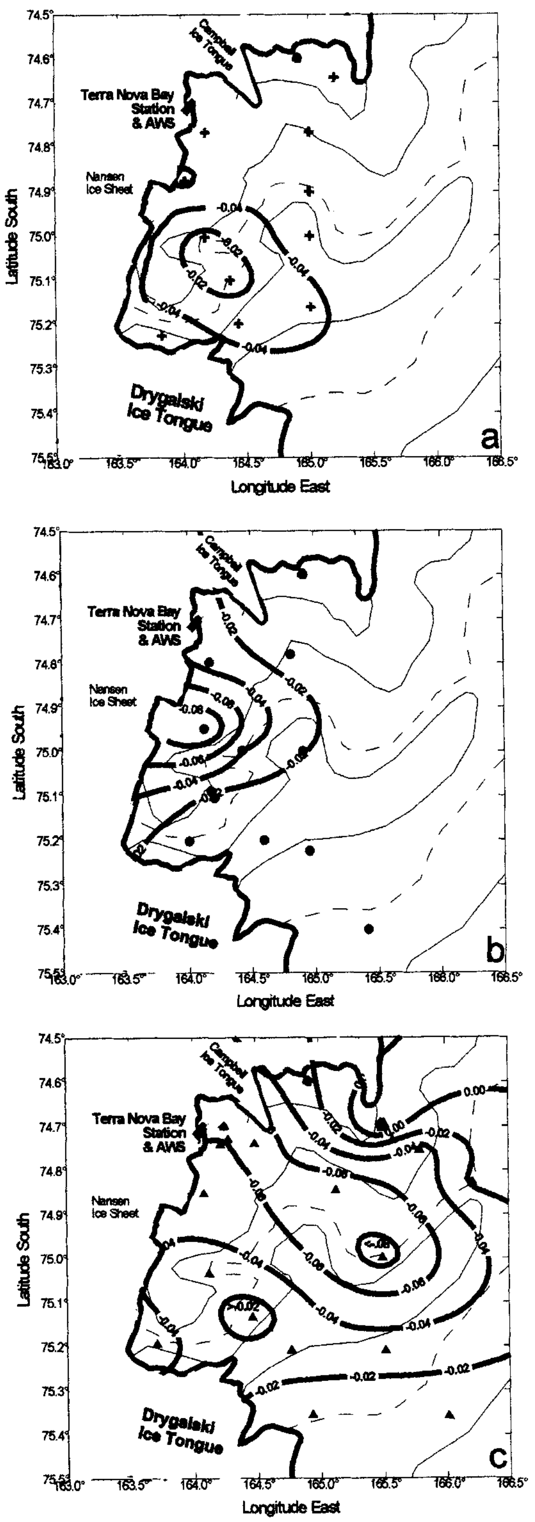


**Fig. 5.** Distribution of the maximum anomaly (°C) between the *in situ* potential temperature and the surface freezing point (S = 34.85) for the three most extensive CLIMA cruises in Terra Nova Bay. **a.** TNB10, **b.** TNB13a, and **c.** TNB13b. These non horizontal surfaces are referred to a depth ranging from: **a.** 330 m ± 170, **b.** 490 m ± 320 and **c.** 450 m ± 200.

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of the concentration occupies a significant part of the TNB region. The most likely mechanism by which this water mass is formed is by the contact with the ice shelves or tongues of TNB. Due to the high salinity values detected in this TISW layer (S > 34.7) this water must be primarily formed by the interaction of the salty HSSW with the base of the glacial ice.

During the CLIMA surveys the TISW was present in massive quantities in the western sector of TNB, which suggests that the TISW flows north-eastward after exiting from the NIS. It is possible that the northern sector of TNB could also play a small but not negligible role in the cooling of the shelf waters in TNB. Cold waters originated in the northern sector are then eventually advected in the central region of TNB by the clockwise coastal circulation.

The northward flow along the coast and to the central part of TNB agrees with the $\theta$–$T_f$ contours of TNB13b cruise which show the tongue of TISW with a south-east direction (Fig. 5c). During the TNB13a the area covered by the TISW was bounded close to the NIS (Fig. 5b). During the first cruise (TNB 10) the TISW spreading was not detectable. The whole TNB area was homogeneously occupied by this cold water except for the sector between the NIS and the DIT. Actually the summer water column of this sector in the southern part of TNB is often characterized by the higher temperature and salinities of the whole bay (Figs 5 & 6). It is suggested that the polynya is still functioning in this area during the summer season and therefore contributes to maintaining the intermediate water column saltier and warmer (close to the surface freezing point) than the neighbouring (see Discussion).

The typical depth showed by the TISW ranges from 30 m to 600 m with the shallower values in the north-west sector close to the TNB Italian station and getting deeper by the coast before its characteristics become undetectable after mixing with the surrounding waters. Also the depth thickness changes considerably from approximately 400 m (close to the coast) to 150 m in the central region of TNB. The presence of the TISW modifies the vertical salinity profile creating the apparent break in slope in the salinity profile at about 750 m (Fig. 3a).

*Horizontal distribution*

The horizontal distribution of the thermohaline properties observed in TNB during the three cruises exhibit some similarities in the intermediate layers, shown in Fig. 6 for the 500 dbar layer. At this depth the influences of the cold TISW is still detectable. The coldest potential temperatures observed ($\theta$ < freezing surface temperature) are associated with relatively fresher waters (S < 34.77), thus supporting the idea that this water derives from the interaction of HSSW (saltier) with the underwater base of glacial ice. Actually, the salinity field observed at an intermediate depth typically shows a decrease ($\Delta S = 0.02 - 0.05$) from the deeper central stations toward the coast and south-eastward.

The bottom salinity distribution (not shown), as deducible from Fig. 3 (below the intermediate layer the salinity increases with depth and the potential temperature is warmer), exhibits maximum values (S > 34.84) corresponding to the deepest stations inside the Drygalski Basin with a massive spreading along its axis. The greatest salinity was observed during the TNB13a cruise (S > 34.88) in the southern region close to the DIT and spreading northward along the basin axis but we have no evidence of a similar southern path. Minimum values of potential temperature ($\theta < -1.93°C$) were often observed close to the northern sector of TNB, restricted to a narrow tongue protruding from the coastline into the open sea.

Moreover, it must be stressed that in TNB the maximum salinity values (S > 34.86) are observed close to the bottom layer (more than 1000 dbar); this water (HSSW) is supposed to be produced by the surface interaction with the cold atmosphere. At the surface, the maximum cooling is limited by the surface freezing point but the potential temperatures measured in the bottom layer are often slightly colder than the $T_f$ ($\theta < -1.93°C$ for $P > 1000$ dbar). A direct (glacial melting) and indirect (water mixing with coldest water derived from the glacial melting) interaction of these layers can be assumed to explain these observed temperatures. Obviously, the possible contact with the ice shelf produces greater effects on the temperature in comparison to those on the salinity because of the different values of the eddy diffusivity for the heat and salt exchanges.

The cold temperatures observed in the surface layer (Fig. 7) in the central region of TNB may be caused locally by heat loss at the air–sea interface, forced by the strong winds blowing from the coast also in the summer period. Moreover, the lower temperatures are sometimes correlated with maximum salinity values, which supports the idea of a strong interaction with the atmosphere producing ice formation with salinity release also during these summer investigations.

*Water property changes*

In order to study thermohaline variability in the water column in TNB we analysed the heat and salt changes from the surface to the bottom as described in Equations 1 & 2.

Figure 8 shows the mean heat and salt changes obtained for the cluster of stations inside TNB (shaded area in Fig. 1) performed at the beginning (TNB13a) and at the end (TNB13b) of the summer 1997/98 (Table IV).

These changes show the warming of the surface layer (occupied by the SSW) driven by seasonal change in the interaction with the warmer atmosphere and by solar heating as observed during the TNB13 cruise. At intermediate depth a decreasing temperature characterizes the water column, unlike the upper pycnocline layer; at greater depths the heat variability is not appreciable because it is less than the accuracy of the CTD sensors. The surface layer displays a freshening probably caused by the ice melting (Fig. 8b). Like the surface layer, the intermediate one (300–600 dbar) becomes fresher with a similar magnitude down to a depth of 600 dbar.

These surface salinity changes are in agreement with the

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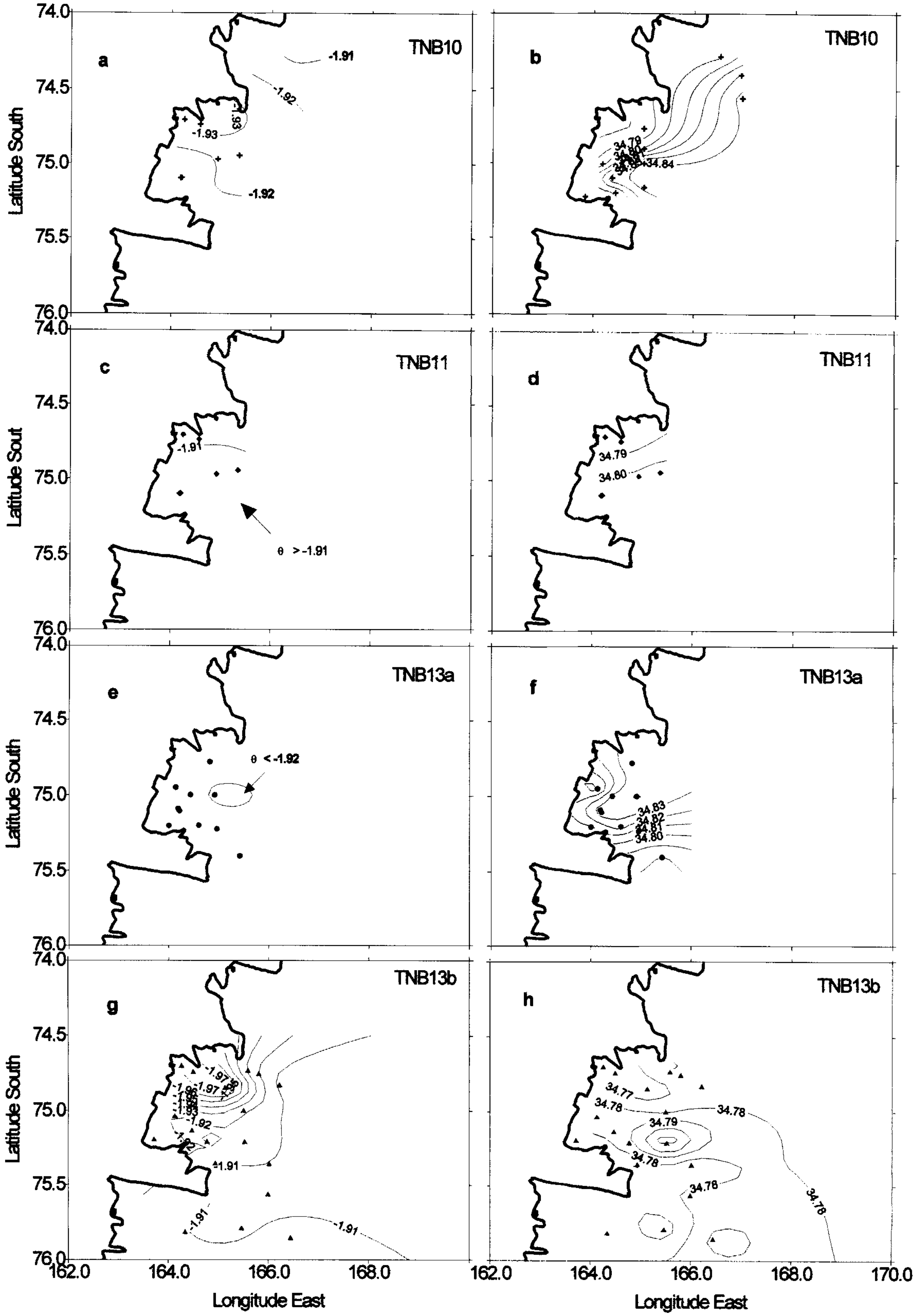


**Fig. 6. a, c, e & g.** 500 dbar potential temperature (°C) and **b, d, f & h.** salinity distribution for the four cruises.

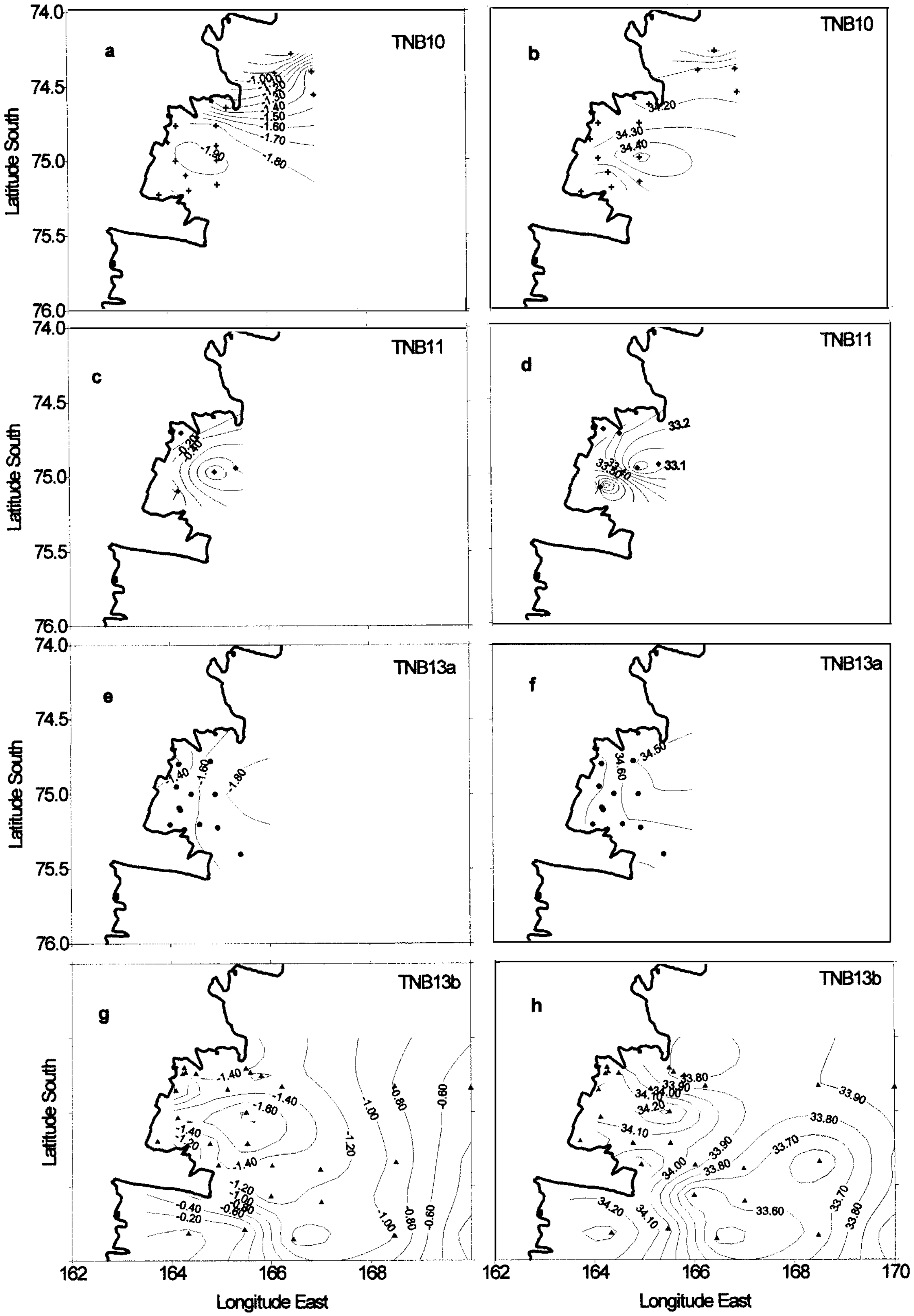


**Fig. 7. a, c, e & g**. surface potential temperature (°C) and **b, d, f & h**. salinity distribution for the four cruises.

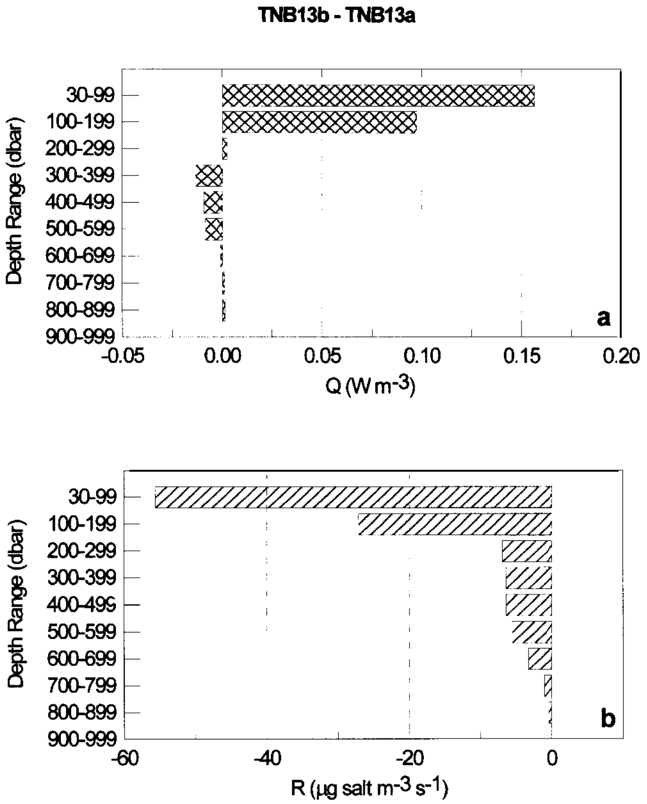


**Fig. 8.** Mean **a.** heat (W $m^{-3}$) and **b.** salt (µg salt $m^{-3}$ $s^{-1}$) estimated changes for the cluster of stations inside the shaded area of Fig. 1 from the beginning (TNB13a) to the end (TNB13b) of summer 1997/98.

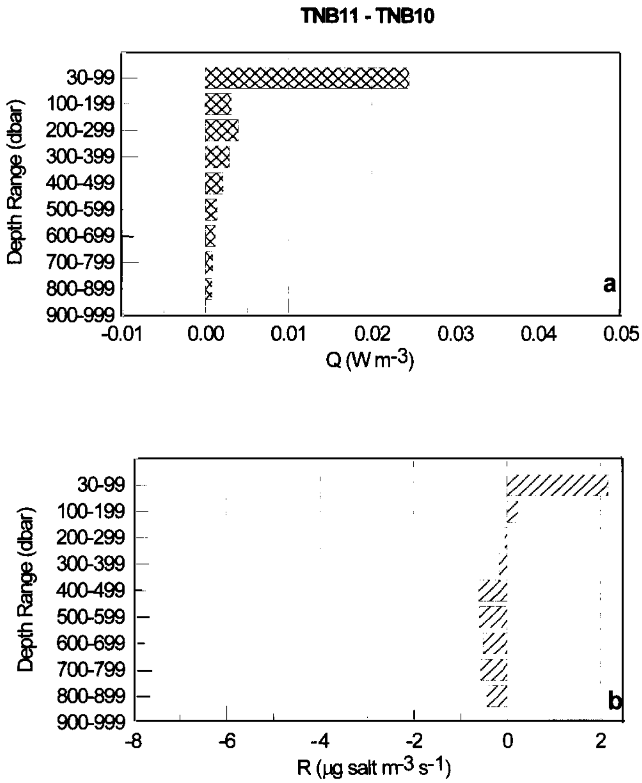


**Fig. 9.** Mean **a.** heat (W $m^{-3}$) and **b.** salt (µg salt $m^{-3}$ $s^{-1}$) estimated changes for the cluster of stations inside the shaded area of Fig. 1 considering TNB10 (summer 1994/95) and TNB11 cruises (summer 1995/96).

time series recorded in TNB by moored SeaBird instruments showing a progressive freshening in the summer season (Picco *et al.* 1998, 1999). On the other hand, the thermal variability from the beginning to the end of the summer in the intermediate and deep layers, as detected by our hydrological measurements, is less than 0.02°C; thus it is not comparable with the time series collected by the Aanderaa thermistors mounted at this depth in the same mooring.

Year to year changes are analysed by considering cruises TNB10 vs TNB11 (Fig. 9) and TNB10 vs TNB13b (Fig. 10). In both cases we observed a warming of the upper pycnocline layer, although with different magnitudes (Table IV). The intermediate and deep layers show a temperature increase between TNB10 and TNB11 sampling (Fig. 9a), while a cooling is detected during TNB13b compared to TNB10 cruises (Fig. 10a).

A different situation is obtained for the salinity changes from TNB10 to TNB11 between surface and deep layers. In the surface layers the salinity increases, while a quasi-constant freshening of the water column is detected from 400 to 900 dbar, corresponding to an average $\Delta S = -0.016$ (Fig. 9b). A similar freshening, relative to an average $\Delta S = -0.014$, is

**Table IV.** Mean heat and salt changes obtained for the cluster of stations inside TNB (shaded area in Fig. 1) performed at the beginning (TNB13a) and at the end (TNB13b) of the summer 1997/98. Years to year changes are analysed considering cruises TNB10 vs TNB11 and TNB10 vs TNB13b.

| | TNB13b-TNB13a | | TNB11-TNB10 | | TNB13b-TNB10 | |
|---|---|---|---|---|---|---|
| Depth range m | Integrated heat flux W $m^{-2}$ | Integrated salt flux mg salt $m^{-2}$ $s^{-1}$ | Integrated heat flux W $m^{-2}$ | Integrated salt flux mg salt $m^{-2}$ $s^{-1}$ | Integrated heat flux W $m^{-2}$ | Integrated salt flux mg salt $m^{-2}$ $s^{-1}$ |
| 0–10 | -0.2 | -0.6 | 1.3 | -0.2 | .3 | 0.0 |
| 0–30 | 0.1 | -2.1 | 4.9 | -0.8 | 1.2 | -0.2 |
| 0–50 | 3.1 | -3.4 | 6.1 | -0.8 | 2.2 | -0.3 |
| 0–100 | 11.5 | -6.0 | 6.5 | -0.7 | 4.1 | -0.7 |
| 100–200 | 9.6 | -2.7 | 0.3 | 0.0 | 2.0 | -0.4 |
| 200–300 | 0.2 | -0.8 | 0.4 | 0.0 | 0.2 | -0.1 |

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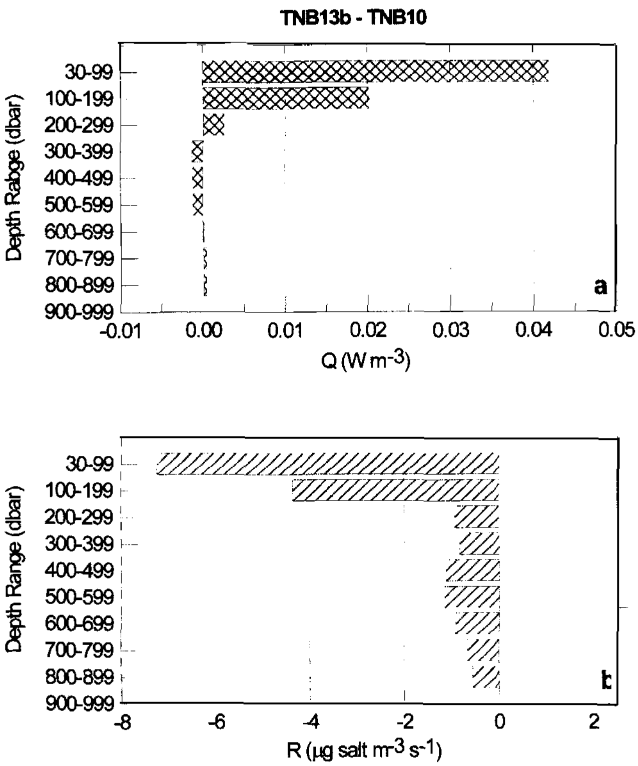


**Fig. 10.** Mean **a.** heat (W $m^{-3}$) and **b.** salt (μg salt $m^{-3}$ $s^{-1}$) estimated changes for the cluster of stations inside the shaded area of Fig. 1 using data from TNB10 (summer 1994/95) and TNB13b cruises (summer 1997/98).

detected between TNB10 and TNB13b at the intermediate and deep layers (Fig. 10b).

In the surface layer, as expected, the seasonal changes in the heat and salt contents are larger than the year to year changes. Given the high latitude, it seems reasonable that the extreme seasonal changes in the atmosphere produce a large seasonal

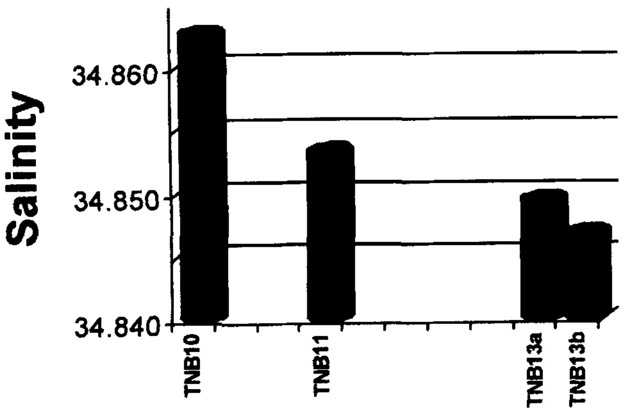


**Fig. 11.** Average salinity in a 10 m thick layer at 900 dbar (that is the maximum common depth for the analysed cruises). This layer is characterized by a decrease in salinity from the summer 1994/95 to the summer 1997/98.

variability in water. It is known that surface heat fluxes in this region can vary from year to year (Van Woert 1999, Budillon *et al.* 2000a, 2000b), thus altering forcing of the surface layer.

## Discussion

The thermohaline data analysed are the first to document the water column structure and variability through successive measurements made both within year and between years in a crucial area of the Ross Sea such as the Terra Nova Bay polynya.

It must be stressed that the latitudinal extent of the area used in this investigation is about 50 km; assuming a mean speed of 5 cm $s^{-1}$ for both surface and deep layers, the residence time can be estimated at about 25 days. Thus hydrological measurements performed with a time lag of one month or more, as in our data-set, can describe primarily the variability produced by the horizontal advection. Therefore the subsurface thermohaline variability can be attributed primarily to the large-scale atmospheric forcing acting on the Ross Sea, since the low variability of the other source of heat for the water column (the MCDW) can not modify substantially the thermohaline properties of the shelf waters (Jacobs & Giulivi 1998).

### *Heat and salt variability.*

We computed the magnitude of the heat flux in the upper pycnocline layer (100 m) and the deeper layers between the beginning and the end of the summer 1997/98 and successive cruises. Considering a 100 m deep water column, an integrated heat flux of about 11 W $m^{-2}$ would be necessary to produce the thermal changes observed during the summer 1997/98 in TNB. Using the ECMWF (European Center for Medium-range Weather Forecasting) operational analyses, Budillon *et al.* (2000b) estimated the surface heat fluxes in TNB for the period 1994–97; the average summer (December–January) heat budget has been estimated of about 170 W $m^{-2}$. Kurtz & Bromwich (1985) estimated an average heat flux of about 140 W $m^{-2}$ for the summer period. Both results are substantially higher than the summer heat changes observed in the water column, thus confirming that the TNB area supplies heat to the water column that is used to melt ice (at the beginning of the summer) and then exported by the upper pycnocline flow.

The thermohaline properties of the water above the pycnocline undergo a strong seasonal pattern of change (warming and freshening) driven by the different forcing of the atmosphere from the beginning to the end of the summer and by the ice melting (Fig. 8). Also the intermediate layer below the pycnocline exhibits a clear variability from the beginning to the end of the summer becoming less saltier and colder, suggesting a key role for the advective component. The differences in sign and magnitude for the heat change in the upper and in the lower pycnocline indicate a weak connection in the summer period between the surface heat

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fluxes and the deep waters.

Figure 11 shows the average salinity in the near bottom layer (900 dbar) for all the cruises analysed (that is the maximum common depth for all the cruises analysed). This 10 m thick layer is characterized by a decrease in salinity from the TNB10 (summer 1994/95) to the TNB13b cruise (summer 1997/98). Jacobs & Giulivi (1998) reported that the 1994 HSSW deep salinity measured in the western sector of the Ross Sea had been the lowest since 1967. Thus the TNB deep measurements ($\Delta S$ = -0.015 during four years) are certainly consistent with the freshening trend of the HSSW in the Ross Sea ($\Delta S$ = -0.03 per decade) which could be related to the decrease of ice formation or to a shorter shelf water residence time coincident with a decrease of the winter surface forcing (Jacobs & Giulivi 1998).

*Summer polynya.*

As already observed, the surface of the central-southern region in TNB is colder and saltier than the surrounding area, and this pattern is also confirmed in the intermediate layer (Figs 6 & 7). This region is under the direct influence of the katabatic winds blowing from the interior of the coast. Actually, summer ice formation (at the end of January and at the beginning of February) has been observed and documented during some mini-surveys performed along the southern coast of TNB.

During the last cruise (TNB13b), in the region offshore the NIS (Fig. 1), three casts have been performed before and four after a strong katabatic regime recorded by the AWS in TNB Italian station. In order to quantify the thermohaline changes resulting from such atmospheric forcing we computed the salting rate (as defined by Eq. 2) from the surface to the depth of 400 m (Fig. 12). From the surface to the depth of 25 m, the salting rate is positive with a maximum value at about 15 m, while it is negative down to 400 m. The vertically integrated salt rate from the surface to 20 m depth indicates a positive salt flux of about 0.44 mg salt m$^{-2}$ s$^{-1}$ which is equivalent (for the corresponding period - 12 days) to a mean daily sea ice production of about 1.3 mm (assuming the water and the ice salinity of 34.0 and 5.0 respectively); that is a small but realistic value in the summer period. Likewise, in the same vertical range, the water column exhibits stronger cooling over all the depth, of about 30 W m$^{-2}$. It is therefore possible to hypothesize a possible functioning of the polynya, even in the summer period, as characterized by a small rate of ice production. Owing to this modest rate and the absence of the preconditioning phase due to the summer vertical stratification, it is not reasonable to suppose a HSSW formation in the strict sense for this period, which therefore remains a typical winter process. Actually the rejected brine is added to the near surface layers that are warmer and fresher than the deeper layers (Fig. 3).

The low surface salinity values observed northward of NIS close to the coast (Fig. 7b, d & h) are well associated with the warmer surface temperature and could be related to a local upwelling induced by the offshore wind drag. Due to the summer season the water upwelled from the near surface layer is warmer and fresher respect to the water under the direct influence of the katabatic forcing (Povero *et al.* in press).

*Modified Circumpolar Deep Water (MCDW).*

Simultaneous warming and freshening (cooling and increase of salinity) of the upper intermediate layers may be associated with a greater (lesser) influence of the MCDW in this region. By identifying the thermal anomaly at an intermediate depth, the intrusion of MCDW into the western Ross Sea was observed (Budillon *et al.* 1999) to reach the area south-east of Coulman Island at 73.5°S. This relatively warm tongue was identified over a 200 km path along the western flank of the Mawson Bank (supporting the idea that its position is controlled by the bathymetry of the sea floor) from the slope region to 74°S. The CLIMA data-set does not show any clear evidence of intrusion of such warmer and fresher water into TNB area even if a warm core with $\theta > -0.4$°C at a depth of 100–150 dbar was detected further east at a longitude of 170°W on the Pennel Bank (not shown).

Obviously, the most important MCDW intrusions close to the TNB area are expected when the atmospheric forcing and brine release are greater and when massive quantities of HSSW are formed, which is typically during the strong katabatic surge events in winter. It is not surprising then that we do not have any evidence of such intrusions from the

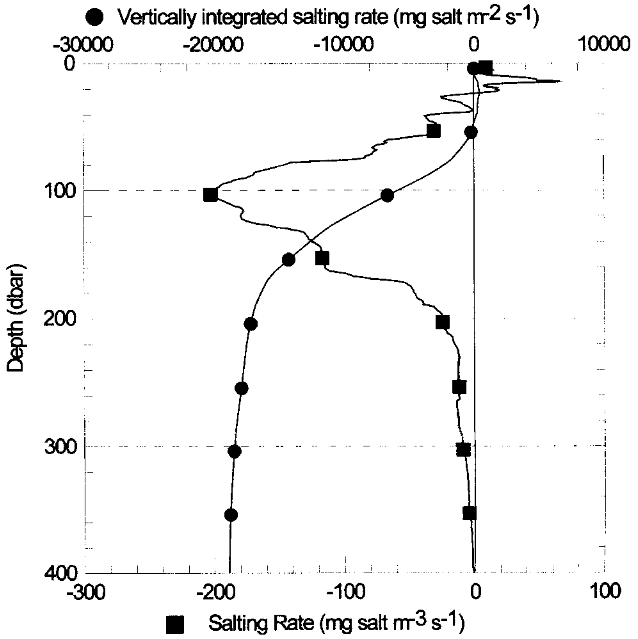


**Fig. 12.** Estimated salting rate (as defined by Eq. 1) from the surface to the depth of 750 dbars during strong katabatic wind episode occurred during TNB13b cruise (February 1998)

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summer only data-set.

*Terra Nova Bay Ice Shelf Water (TISW) and ice interaction.*

The David Glacier (not shown) originates the Drygalski Ice Tongue; its grounding line is about 50 km inland from the coast (Frezzotti 1997). Jacobs *et al.* (1985) outlined the process by which the shelf waters evolve into ISW beneath ice shelves (in particular for the Ross Ice Shelf); a similar process may happen in TNB.

Since in TNB the HSSW temperature is warmer than the *in situ* freezing points at these depths, this shelf water provides the latent heat for the basal melting, thus originating the fresher and colder TISW. The thermohaline circulation associated with this process provides the mechanism to transports the less dense TISW out of the sub-ice shelf cavities. As a result of this process, the intermediate layer (200–500 m) in TNB is fresher ($\Delta S \cong -0.08$) and colder ($\Delta T \cong -0.02$°C) than the deep-bottom layer as showed by the Fig. 3a.

The heat utilized for the melting of the ice and the subsequent decrease of salinity determines the cold temperature and the slope in the salinity profiles in the intermediate layers.

The cooling and freshening rates of these intermediate layers of HSSW interacting with this glacial ice depends on the residence time beneath the ice shelves and could be modified by the coastal circulation of the TNB. Indeed the presence of the TISW in TNB has shown a strong variability, having been observed with different distributions in the thermohaline properties during the CLIMA surveys. It has been detected in all surveys but as warmer during the TNB11 (not shown) and TNB13a and colder in TNB10 and TNB13b (Figs 3, 5 & 6). It is possible that these observations could be somehow wrong or very sensitive to the different sampling of each survey, thus we do not put forward any hypothesis to explain such variability.

Using the CTD casts collected during the most intensive surveys (TNB10, TNB13a and TNB13b) it is possible to roughly estimate the mass, heat and freshwater transport associated with the TISW outflow and consequently to calculate the equivalent melting rate. Obviously these estimates are biased by the small number of casts, by their different position and by the non-synoptic sampling.

**Table V.** Estimates of the melt rate and fresh water transports of TISW in TNB using a reference temperature of -1.91°C (correspondent to the surface freezing temperature for S = 34.8), a reference salinity of 34.8 and an average velocity of 2.5 cm s$^{-1}$ (1 Gt = 10$^{12}$ Kg). Both salinity and temperature values have been chosen taking into account that the TISW is originated by the HSSW. Results obtained for the TNB11 survey are not representative because of the small number of casts and, principally, the limited area sampled.

| Acronym | No of casts | Melt rate Gt a$^{-1}$ (m$^3$ s$^{-1}$) | Freshwater transport Gt a$^{-1}$ (m$^3$ s$^{-1}$) |
|---|---|---|---|
| TNB10 | 7 | 2.6 (82) | 1.7 (311) |
| TNB11 | 6 | 0.6 (18) | 1.3 (300) |
| TNB13a | 10 | 1.7 (55) | 5.7 (76) |
| TNB13b | 11 | 3.4 (107) | 3.7 (376) |

The horizontal transport is calculated using the mean values measured by a mooring located in TNB (see section 3.5 in Picco *et al.* 1999). In the intermediate layer currents are primarily barotropic with an eastward direction and a mean speed of 2.5 cm s$^{-1}$ (year average). Assuming a mean TISW layer of 150 m (Fig. 3a) and an horizontal section of 80 km (approximately the distance between the DIT and Cape Washington, Figs 1 & 5) the inferred TISW mass transport is 0.3 Sv (1 Sv = 10$^6$ m$^3$ s$^{-1}$). Thus it is possible to calculate the heat transport $Q_h$:

$$Q_h = C_p \cdot \rho \iint \left(\vartheta - \vartheta_0\right) \cdot V \cdot dz \cdot ds \qquad (3)$$

and the fresh water transport $Q_f$:

$$Q_f = \frac{1}{S_0} \cdot \iint \left(S - S_0\right) \cdot V \cdot dz \cdot ds \qquad (4)$$

where V is the speed (m s$^{-1}$), $\theta$ and S are respectively the measured potential temperature (°C) and salinity, and $\theta_0$ and $S_0$ are the potential temperature (°C) and the salinity spatially averaged in the area studied (i.e. the potential temperature and the salinity of the surrounding HSSW: $\theta_0$ = -1.91°C and $S_0$ = 34.80); $C_p$ and $\rho$ are the specific heat capacity (4218 J kg$^{-1}$ K$^{-1}$) and the water density (1028 kg m$^{-3}$) respectively.

The estimates of heat and fresh water transport are summarized in Table V. Computations performed for TNB11 are not very representative because of the small number of stations and the limited area covered during this survey. However, these results must be representative of the basal melting rate of the NIS and DIT if the hypothesis that all the water involved in this process flows into TNB and then the TISW is adequately described by our measurements. This assumption is verified by the absence of any kind of ISW south of the DIT in the CLIMA data-set and supported also by the known general circulation in TNB area that shows a prevailing NNE direction.

Frezzotti (1997), and Frezzotti *et al.* (in press) estimate the basal melting in TNB using airborne radar survey and satellite image analysis and obtain a value of about 4 Gt a$^{-1}$. This value is in surprising agreement with our results considering the different source of data and analysis used.

The estimates obtained by Eqs 3 &4 are obviously very sensitive to the reference values used. For the temperature the reference value of the surface freezing point is a robust assumption, but it is not very easy to define an exact reference salinity for all the cruises. Actually, because of the change (increase) of the salinity with the depth in TNB, to calculate the fresh water transport from Eq. 4 it is strictly necessary to know the depth of the layer that interacts with the basal ice.

The freshwater results reported in Table V, obtained with a reference salinity of 34.8, show a higher variability and also a not negligible difference to those obtained in terms of heat budget. This suggests that the estimation from the heat transport is the most accurate method of obtaining the ice shelf melt rate in TNB considering its peculiar hydrography. On the other hand some ocean heat could be lost between the measurements and the ice shelf front by the inadequate sampling; consequently it must be emphasized that these estimations could be only representative of a reference minimum value.

## Conclusions

The hydrological measurements from three cruises in the coastal area of the Terra Nova Bay polynya from summer 1994/95 to summer 1997/98 offer a unique opportunity to analyse salinity and temperature structure and their changes in a high-latitude system subjected to strong atmospheric forcing throughout the whole year. The high density of observations allows analysis of the year to year thermohaline variability as well as the short term changes in water properties between the beginning and the end of the summer.

The summer vertical structure of TNB is formed by a surface layer (SSW) subject to the atmospheric forcing. This layer is superimposed onto the layer of TISW originated by the HSSW modified by the interaction with the ice shelves and ice tongues, and it is then characterized by a temperature colder then the freezing point at the surface and by lower salinity than those of the HSSW. Below it, the HSSW occupies the water column down to the bottom with a potential temperature close to the surface freezing point (and sometimes colder).

Also it is suggested that in the summer season one role of the polynya activity is to increase salinity in the surface layer during katabatic events and to provide a saltier and warmer (with respect to the TISW temperatures) area in the southern sector of TNB. Due to this modest rate and the absence of the preconditioning phase of water column, it is not reasonable to hypothesize an HSSW formation in this period, which therefore remains a typical winter process.

Summer water property changes above the permanent pycnocline (about 50–100 m and 150 m at the beginning and the end of the summer, respectively) follow the expected seasonal pattern of warming and freshening from the beginning to the end of the summer. Conversely in the deep layers (> 600 m) the thermal variability is negligible.

The year to year changes in heat and salt content in the surface and intermediate layer are greater than those expected from sensor precision and internal wave activity in the surface and intermediate layers. The observed decrease in salinity of the deep layers is in good agreement with the salinity trend of -0.03 per decade recently reported by Jacobs & Giulivi (1998).

Compared to the TNB10 survey (summer 1994–95) the two successive cruises, TNB11 (summer 1995–96) and TNB13a (summer 1997–98), measured a freshening of the subsurface layers even with different magnitudes associated with a warming and a cooling respectively. The differences in sign and magnitude for the heat change in the upper and in the lower pycnocline indicate a weak connection in the summer period between the surface heat fluxes and the deep waters.

The year to year differences observed below the pycnocline cannot be related to the surface fluxes but are produced by the advective component of the general circulation in the western Ross Sea. The subpycnocline waters of this portion of the western Ross Sea may therefore be affected more by variations in the large-scale oceanic current than by local oceanic or atmospheric processes providing a different preconditioning of the water column for the winter vertical convection during the HSSW formation phases.

The different thermohaline properties observed during the three CLIMA cruises also emphasise the possible variability in the HSSW winter formation processes due to a different preconditioning of the surface and subsurface layers of the water column.

Finally, the presence of a very cold water mass (TISW, $\theta < -1.92°C$) originating from the interaction with the glacial ice (by a similar process to that forming the Ice Shelf Waters beneath the Ross Ice Shelf) was detected during all CLIMA cruises. The calculation of the basal melting rate (from 1.7 to 3.4 Gt $a^{-1}$) associated with the TISW cold temperatures and salinities is consistent with those obtained from radar and satellite image by other research within TNB. The presence and the variability of this water mass indicates the key role played by thermohaline properties in TNB waters in controlling the floating glacier by regulating possible basal melting processes.

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