Double-diffusive layering and mixing in Patagonian fjords

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A B S T R A C T

Double-diffusive layering was quantified for the first time in the Chilean Patagonian fjords region (41.5–56°S). Approximately 600 temperature and salinity profiles collected during 1995–2012 were used to study water masses, quantify diffusive layering and compute the vertical diffusivity of heat. Development of “diffusive-layering” or simply “layering” was favored by relatively fresh–cold waters overlying salty–warm waters. Fresh waters are frequently derived from glacial melting that influences the fjord either directly or through rivers. Salty waters are associated with Modified Subantarctic (MSAW) and Subantarctic Water (SAWW). Double-diffusive convection occurred as layering in 40% of the year-round data and as salt fingering in <1% of the time. The most vigorous layering, was found at depths between 20 and 70 m, as quantified by (a) Turner angles, (b) density ratios, and (c) heat diffusivity (with maximum values of 5 × 10^-5 m^2 s^-1). Diffusive-layering events presented a meridional gradient with less layering within the 41–47°S northern region, relative to the southern region between 47° and 56°S. Layering occupied, on average, 27% and 56% of the water column in the northern and southern regions, respectively. Thermohaline staircases were detected with microprofile measurements in Martinez and Baker channels (48°S), showing homogeneous layers (2–4 m thick) below the pycnocline (10–40 m). Also in this area, increased vertical mixing coincided with the increased layering events. High values of Thorpe scale (LT ~ 7 m), dissipation rate of TKE (ε = 10^-5 – 10^-3 W kg^-1) and diapycnal eddy diffusivity (Kz ~ 10^-6 – 10^-3 m^2 s^-1) were associated with diffusive layering. Implications of these results are that diffusive layering should be taken into account, together with other mixing processes such as shear instabilities and wind-driven flows, in biological and geochemical studies.

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Introduction

Double-diffusive convection (DDC) produces convection in a wide variety of fluids (Schmitt, 1994). It contributes to vertical mixing of the water column (You, 2002; Mack and Schoeberlein, 2003), as well as to thermohaline circulation and heat transport (You, 2002). When hydrodynamically stable water masses overlay each other, and the quantity of heat and salt within these layers vary, the process known as double-diffusive convection occurs as an attempt to restore equilibrium. In other words, double diffusive convection occurs in response to differences in molecular diffusion rates between heat and salt as heat diffuses approximately 100 times faster than salt (Schmitt, 2001). If heat or salt are unstably distributed within the water column, potential energy is released from the unstable component through molecular diffusion. Therefore, double diffusive convection develops in the water column when the slopes of vertical gradients of temperature and salinity have the same sign.

When both temperature and salinity decrease with depth, which is characteristic of latitudes where evaporation dominates over precipitation, double diffusive convection occurs as “salt-fingering” or “finger regime”. And when both temperature and salinity increase with depth, as is typical of high latitudes where precipitation dominates over evaporation, double diffusive convection is defined as “diffusive-layering” (Kelley et al., 2003) or simply “layering,” as will be called from hereon. When cold fresh water sits on top of warm salty water (conditions favorable for layering), heat from the lower layer is transferred to the colder upper layer by the release of potential energy from the temperature field. This is an overcompensating restoring force, which...
produces defined layers of well-mixed fluids separated by marked density gradients (Schmitt, 1994; You, 2002; Kelley et al., 2003). Typically, the top of the water column of Patagonian fjords is cold and fresh because of melting glaciers, and because of input from river discharge and precipitation.

Early evidence of double-diffusive mixing in the ocean was found through the development of simultaneous staircase structures in vertical profiles. Schmitt (1981) described double diffusive convection for the Tyrrhenian Sea (west of Italy), the Mediterranean outflow into the eastern Atlantic, and the subtropical underwater in the western subtropical Atlantic. In studies of the Tyrrhenian Sea, mixed layers exceeded 50 m, whereas in the other two cases they were approximately 10 m. Cascading isothermal layers, with thicknesses ranging from 2 to 10 m were reported for the Arctic Ocean by Neal et al. (1969). Staircase-like layers can be together or be separated by perturbations within the water column (e.g., internal wave). The absence of a well-defined vertical staircase does not indicate that DDC is not present (Farmer and Freeland, 1983).

The global atlas of DDC (You, 2002, based on the 1994 Levitus climatological atlas of the open oceans) shows that this process is favorable in 44% of the oceans, of which 30% is salt fingering and 14% is layering. The occurrence of salt fingering was associated with the Central Waters where evaporation exceeds precipitation, such as the case around Easter Island (Moraga and Valle-Levinson, 1999). But salt fingering also develops with relatively warmer and saltier deep waters overlying colder and fresher Antarctic Bottom Water. In high latitudes, where surface cooling and ice melting are ubiquitous, as in the polar and subpolar Arctic and Antarctic regions (e.g., the Gulf of Alaska, the Okhotsk, Labrador and Norwegian Seas) layering occurs (You, 2002). Evidence for the occurrence of layering in a Patagonian channel was presented by Pérez-Santos et al. (2013) in the form of a staircase structure, with layer thickness between 2 and 5 m, within a vertical profile of temperature and salinity.

The fjords region of Patagonia extends from 41°S to 56°S, occupying an approximate area of 240,000 km², which represents the region with the largest extension of fjords in the world (Fig. 1). Freshwater input in this region consists of river fluxes \( (27.8 \times 10^3 \text{ m}^3 \text{ s}^{-1}) \) and rain \( (33.5 \times 10^3 \text{ m}^3 \text{ s}^{-1}) \) (Dávila et al., 2002). The hydrography of this system is characterized by a vertical structure of two layers in temperature and salinity (Silva and Calvete, 2002). The cooler and fresher surface layer in the first few meters of the water column (5–10 m) is more variable than the bottom layer. This variability occurs at different scales and is caused mainly by fluctuations in solar radiation, freshwater inputs (rivers, rain, glacial ice melt and coastal runoff), advection of water to and from channels and vertical mixing by wind and tide (Silva and Calvete, 2002; Dávila et al., 2002; Sievers, 2008). However, the vertical distribution of warmer temperature and higher salinity in the deep layer tends to be much less variable than near the surface.

Water masses have been described in the fjords and channels of Patagonia by Sievers (2008) and Sievers and Silva (2008) on the basis of salinity values only. Estuarine water (EW, cold–fresh water) was identified in the first few meters of the water column, formed by the contributions of rivers and summer-time glacier ice melting. The EW included Estuarine Salty Water (ESW, salinity between 21 and 31 g/kg) and Estuarine Fresh Water (EFW, salinity between 11 and 21 g/kg), while the waters with salinities less than 11 g/kg were classified as Fresh Water (FW) (Sievers and Silva, 2008). Below ESW, Subantarctic Water (SAAW, salinity >33 g/kg) was described extending to approximately 150 m. Mixing between ESW and SAAW resulted in Modified SAAW (MSAAW) with salinities between 31 and 33 g/kg. Equatorial Subsurface Water (ESSW) extended from 240 m to 300 m, followed by Antarctic

**Fig. 1.** (a) Study area and distribution of hydrographic stations. Details regarding expeditions are listed in Table 1. (b) The main freshwater supply in the Patagonian fjord region, highlighting the Northern and Southern Patagonian Icefield and the most important river. Reference of river discharge was extracted from Meerhoff et al. (2013) to Baker and Pascua rivers, from Calvete and Sobarzo (2011) to Aysén and Cisnes rivers and from Dávila et al. (2002) to the Serrano river.
Intermediate Water (AAIW), whenever the local bathymetry allowed (Sievers and Silva, 2008).

According to the hydrographic features (temperature and salinity profiles increasing with depth) in Chilean Patagonia, layering processes are expected all year round. Using ~600 temperature and salinity (TS) profiles that covered the Chilean Patagonia, DDC events were quantified. Also, dynamic implications when layering is active were evaluated e.g., the vertical diffusivity of heat and other turbulent mixing parameters. The evidence of layering-driven turbulence was presented as a study case in Martínez channel (47.8°S) with the computation of Thorpe scales, diapycnal eddy diffusivities, and dissipation rate of turbulence kinetic energy derived from microstructure profiles.

Data sets

A total of 583 hydrographic stations from the Patagonian fjord region, collected during the years 1995–2011, were used in this study. These stations were either sampled with Sea-Bird CTDs (568 profiles) or with a Self Contained Autonomous MicroProfiler (SCAMP, 15 profiles (Table 1 and Fig. 1a).

Vertical CTD-profiles were analyzed to depths of 100 m due to irregular bathymetry and maximum depth sampled. The percentage of CTD-profiles that sampled from the surface to 100 m was 85.6%; gaps in the first 5 m were attributed to CTD measurements taken with a CTD/Rosette during the Crucero de Investigación Marina en Areas Remotas (CIMAR–fjord) expeditions. The vertical resolution of CTD profiles was ~12 cm working at 8 Hz with a descent rate of ~1 m s\(^{-1}\) (further description in Section 3.3) for COPAS-Tortel and Puyuhuapi expeditions. These data were averaged at 1 m. Therefore, this was the vertical resolution for CIMAR cruises.

The SCAMP instrument allowed for data recording at 100 Hz, or about 1 mm vertical resolution when descending at 10 cm s\(^{-1}\). This instrument has fast conductivity (accuracy of ±5% of the full conductivity scale) and fast temperature response sensors (accuracy of 0.01°C). The SCAMP profiling depth was up to 60 m. Measurements from SCAMP also allowed derivation of dissipation rate of turbulent kinetic energy (\(\varepsilon\)) by autonomously fitting observed vertical gradients in temperature to the Bachelor spectrum (Luketina and Imberger, 2001; Ruddick et al., 2000). This approach to determine \(\varepsilon\) was also compared with values derived from the Thorpe scale parameterization (Thorpe, 1977, 2005).

Methodology

Quantification of vertical double-diffusive convection processes

The density ratio or vertical stability ratio (\(R_p\)) and the vertical Turner angle (\(Tu\)) were calculated to identify and quantify DDC, applying the algorithms proposed in the Thermodynamic Equation of Seawater 2010 (TEOS-10). The Thermodynamic Equation of Seawater 2010 introduced new variables in oceanography, which include absolute salinity (\(S_A\)) and conservative temperature (\(\Theta\)), among others (IOC et al., 2010). Absolute salinity (g kg\(^{-1}\)) represents the spatial variation of composition of the seawater, taking into account the different thermodynamic properties and the gradient of horizontal density in the open ocean. The conservative temperature is similar to the potential temperature, but represents the heat content of seawater with more precision.

The stability ratio, also called the density ratio (\(R_p\)), measures the relative contribution of \(\Theta\) and \(S_A\) to water column stability (Tippins and Tomczak, 2003) and is defined in TEOS-10 as:

\[
R_p = \frac{x^0 \Theta_T}{\beta^H (S_A)_z}
\]

where \(x^0 = -\rho^{-1}(\frac{\partial \rho}{\partial T})\) is the thermal expansion coefficient, \(\beta^H = -\rho^{-1}(\frac{\partial \rho}{\partial S})\) is the haline contraction coefficient and \(\rho\) is the density of seawater. Subindices indicate differentiation with respect to that variable (depth \(z\) in this case).

The Turner angle (\(Tu\), expressed in degrees of rotation) was used by Ruddick (1983) to determine the contributions of DDC in hydrographic data. It can also detect the appearance of salt fingering or layering and quantify the influence of temperature and salinity in the stratification of the water column. From TEOS-10 the following equation is used:

\[
Tu = \frac{\Delta H}{\Delta S}
\]

Table 1

<table>
<thead>
<tr>
<th>No.</th>
<th>Expeditions</th>
<th>Date</th>
<th>Season</th>
<th>Stations</th>
<th>Sampling area</th>
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<td>14/11/2010</td>
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<td>9/09/2011</td>
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<td>23/12/2011</td>
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<td>Puyuhuapi V</td>
<td>27–29/9/2011</td>
<td>Winter</td>
<td>24</td>
<td>Northern Patagonia (41.5°/46.5°S)</td>
</tr>
</tbody>
</table>

Total 583
$Tu = \tan^{-1}\left[\left(x^0\theta_2 + \rho^0(S_{\lambda})_2\right)\left(x^0\theta_2 - \rho^0(S_{\lambda})_2\right)\right]$.  

Interpretation of the Turner angle and density ratio was illustrated in various schemes proposed by Bianchi et al. (2002), You (2002) and Tippins and Tomczak (2003). In this study, the scheme proposed by You (2002, his Fig. 3) was adopted, in which $Tu$ is expressed in the form of four-quadrant arc tangent, as in TEOS-10. Thus, as formulated by You (2002): when $Tu$ is between $-45^\circ$ and $90^\circ$, the density ratio ($R_p$) is between 0 and 1 layering is possible. $Tu$ is between $-45^\circ$ and $45^\circ$ the water column is stable with respect to both temperature and salinity; and when $Tu$ is between $-45^\circ$ and $90^\circ$ or $R_p$ between 1 and $\infty$, fingerling structures can be expected. According to this classification, layering can also be divided into strong ($Tu$ between $-90^\circ$ and $-75^\circ$, and $R_p$ between 0.5 and 1), medium ($Tu$ between $-75^\circ$ and $-60^\circ$ and $R_p$ between 0.25 and 0.5) and weak ($Tu$ between $-60^\circ$ and $-45^\circ$ and $R_p$ between 0 and 0.25) (You, 2002). Profiles of temperature and salinity were averaged every 10 cm for SCAMP measurements to compute $Tu$ and $R_p$ parameters.

**Vertical diffusivity of heat**

Values of vertical diffusivity of heat ($K_T$, m$^2$s$^{-1}$) were calculated with the CTD and SCAMP data sets, according to the empirical formulations of Kelley (1984, 1990):

$$K_T = C\left(0.25 \times 10^9 R_p^{11}\right)^{1/3} 1.4 \times 10^{-7},$$

and $C$ is the flux factor,

$$C = 0.0032 \exp\left(\frac{4.8}{R_p^{0.77}}\right).$$

In the convention of Eqs. (3) and (4) the density ratio is $R$-rho $= \rho(2,9), \rho(2,8)$, i.e. the reverse formulation as used in TEOS-10. Only $R$-rho between 1 and 10 were taken into account, because they correspond to the layering regime (Kelley et al., 2003).

**Estimation of the turbulent mixing in Martinez and Baker channels**

Martinez and Baker channels were chosen as a case study (Fig. 1, central Patagonia) for the estimation of turbulent mixing when layering occurred. This was allowed by the availability of very high vertical resolution from CTD and SCAMP instruments.

**Diapycnal eddy diffusivity and dissipation rate of TKE from Thorpe scale**

In order to estimate turbulent mixing when layering was active, the diapycnal eddy diffusivity, ($K_p$, referred to in text as $K_p$-thorpe) and the dissipation rate of turbulent kinetic energy (TKE), referred in text as $\epsilon$-thorpe, were expressed in terms of the Thorpe scale ($L_T$). As a first step, all CTD data were scrutinized. The CTD was lowered with a winch at an approximate descent rate of 1 m s$^{-1}$. With the CTD sampling rate of 8 Hz, 8 measurements per meter were recorded (one measurement every 12.5 cm). In order to accurately determine a density overturn, at least three measurements are needed. Eddies smaller than 12.5 cm $\times$ 3 (=37.5 cm) would not be seen with CTD data as an instability because the density overturn would occur and vanish before consecutive measurements were taken. In the stratified region, the measurement of turbulence requires a high resolution instrument to detect small eddies (<0.25 cm) (Rockwell et al., 2008). Taking into account this limitation, only the SCAMP density microprofiles (sampling at 100 Hz) from Martinez and Baker channels obtained during December 2011 were included in this analysis.

Estimates of $K_p$-thorpe and $\epsilon$-thorpe were anchored in measurements of overturning length scales, i.e., Thorpe displacements, ($L$). These displacements $L$, were found from sorted density profiles. Once the displacements were established, the Thorpe length scales ($L_T$) were calculated as the root mean square of binned $L$, within a 95% confidence interval (Thorpe, 1977, 2005). Bin sizes were determined with the Ozmidov Scale, $L_o = \sqrt{e/\nu}$, where $N$ is the buoyancy frequency and $\epsilon$ is the dissipation rate of TKE derived from SCAMP by fitting a theoretical Bachelor spectrum (more details in Section 3.3.2). The Ozmidov scale indicates the size of the largest turbulence eddies expected in the region of interest and it was also calculated with the Thorpe scale as: $L_T = 0.8L_o$ (Ozmidov, 1965). This comparison provided appropriate bin sizes for $L_T$ estimates by matching $L_o$ from the relationship above. A bin size of 0.5 m was used in this study.

Diapycnal eddy diffusivity using the Thorpe scale was calculated according to Thorpe (2005), as:

$$K_p$-thorpe $= 0.1NL_T^2,$

and the Thorpe (2005) formula was applied to determine the dissipation rate of TKE:

$$\epsilon$-thorpe $= c_1L_T^2N^3$,  

where $c_1 = (L_o/L_T)^2$. The value of $c_1$ was found to be ~0.64, similar to that reported by Dillon (1982).

**Diapycnal eddy diffusivity using the dissipation rate of TKE derived from SCAMP**

Diapycnal eddy diffusivity ($K_p$ hereafter $K_p$-SCAMP) could be obtained also from the SCAMP hydrographic profiles using the dissipation rate of turbulent kinetic energy ($\epsilon$, reference in text as $\epsilon$-SCAMP), which was in turn derived from the vertical gradient of temperature. Then, the dissipation rate of TKE derived from SCAMP was:

$$\epsilon$-SCAMP $= (2\pi)^4\nu D_2^2K_T^4,$$

where $\nu$ is the molecular viscosity of water, $D_2$ is the molecular diffusivity of heat and $K_T$ is the Bachelor wavenumber. More details on this approach are given in Ruddick et al. (2000), Luketina and Imberger (2001) and Steinbuck et al. (2009).

The dissipation rate of TKE could be obtained from the temperature gradient by fitting a theoretical Bachelor spectrum as described above and by also using shear probes. Kocsis et al. (1999) conducted an experiment of dissipation measurements using both methods in the same location but under different oceanographic conditions e.g., during low and high winds regime and thermal induced convection and stratification ($\epsilon$, ranged from $10^{-11}$ to $10^{-9}$ W kg$^{-1}$). The comparison of the results of both experiments carried out by Kocsis et al. (1999) showed good agreement within a factor of 2 magnitudes.

The most used formulation for diapycnal eddy diffusivity estimation using $\epsilon$ from microprofilers was proposed by Osborn (1980):

$$K_p$-SCAMP $= \frac{I}{N^2} \epsilon$-SCAMP,$$

where $I$ is the mixing efficiency, generally set to 0.2 (Thorpe, 2005), and $N$ is the buoyancy frequency. Shih et al. (2005) noted that when the ratio $\epsilon/N^2$, with viscosity $\nu = 1.9 \times 10^{-6}$ m$^2$s$^{-1}$, is larger than 100, the Osborn equation results in an overestimation. They propose a new parameterization for this case given by:

$$K_p$-SCAMP $= 2\nu \left( \frac{\epsilon$-SCAMP}{\nu N^2} \right)^{1/2}.$$

More recently, Cuypers et al. (2011) used Eq. (9) when $\epsilon$-SCAMP/$\nu N^2 > 100$, Eq. (8) when $7 < \epsilon$-SCAMP/$\nu N^2 < 100$, and considered null eddy diffusivity when $\epsilon$-SCAMP/$\nu N^2 < 7$. This approach was followed here.
Results

Water masses and hydrographic features

Five water masses were identified in the first 100 m of the water columns from the data set based on the findings by Sievers and Silva (2008) (Fig. 2a). Fresh Water (FW) occupied the first 10 m of the water column when present, with an average $\Theta$ and $S_A$ of 10.9 °C and 4.5 g kg$^{-1}$, respectively (Table 2). Estuarine Freshwater (EFW) was found at the surface or underneath the FW, with average $S_A$ of 16.9 g kg$^{-1}$ and $\Theta$ of ~10.3 °C, representing a very similar percentage (1.6%) to the FW.

Estuarine Salty Water (ESW) was present either close to the surface or in sub-surface waters below its fresher counterparts occupying about one third (30.7%) of the water column, with an average $S_A$ and $\Theta$ of 28.8 g kg$^{-1}$ and 9.3 °C, respectively. Modified Subantarctic Water (MSAAW) constituted most (45.9%) of the top 100 m of the water column with an average depth of 49.2 m. Absolute salinity
(32.2 g kg\(^{-1}\)) and temperature (9.8 °C) of this water mass were both, on average, greater than those of the overlying ESW. The increase of temperature and salinity with depth give rise to the potential for layering at the interface of these water masses. Saltier (33.3 g kg\(^{-1}\), on average) and slightly colder (9.7 °C) SAAW was located below MSAAW. Absolute salinity, thus, was hydrodynamically stable within the water column in these Patagonian fjords, while conservative temperature was unstable (e.g. Fig. 2c and d). Therefore, stabilization of the water column occurred when warmer water overlaid colder water and destabilization of the water column occurred when conservative temperature increased with depth.

The CIMAR-9I campaign (Fig. 2b) shows an example of the advection of heat by the oceanic waters (SAAW and MSAAW) to the interior of fjords and channels (Fig. 2c and d). In the open ocean and Guapo mouth, the higher temperature and salinity data recorded in the first 100 m depth of the water column coincided with the position of the SAAW. When the oceanic water mass came in contact with the estuarine water, the differences in density located the SAAW and MSAAW below the ESW (Fig. 2e), which created favorable conditions for the occurrence of the layering.

Fig. 3 shows two profiles where conservative temperature and absolute salinity increased with depth in some segments and layering events occurred. The strongest stratification of the water column was identified in the top 10 m, produced by the absolute salinity gradient. A typical profile of Θ and S\(_a\) in the Steffen-Baker region showed that layering (\(\text{Tu} = -59.8^\circ\), \(R_p = 0.26\)) occurred exactly where the conservative temperature and absolute salinity increased at 5 m depth, two meters above the maximum stratification of the water column (Fig. 3, upper panel) and at around 26 m (\(\text{Tu} = -68.6^\circ\), \(R_p = 0.44\)). Two salt fingering events occurred below 70 m because of small changes of both Θ and S\(_a\). At Messier Channel, 90 km from the Baker River mouth, the influence of the ESW above the MSAAW caused layering in the first meters of the water column (\(\text{Tu} = -51.5^\circ\) and \(R_p = 0.11\)) and also, at 12 m with \(\text{Tu} = -50.9\) and \(R_p = 0.1\) (Fig. 3, lower panel). Below 20 m depth, layering was observed within the MSAAW from values of \(\text{Tu} = -75.9^\circ\) and \(R_p = 0.6\). In contrast, fingering was not observed at this station.

The impact of vertical temperature distribution on destabilizing the water column was illustrated with a hydrographic section from Martínez Channel to the Baker River mouth (Fig. 4). The section was located between the northern and southern Patagonian ice fields (labeled “COPAS-Tortel 3” in Table 1 and shown in Fig. 4a), and sampled during fall 2009. Conservative temperature and absolute salinity transsects along this channel revealed two distinct thermohaline layers (Fig. 4b and c). The upper layer (0–10 m) was coldest with conservative temperatures between 6 and 8 °C and freshest with absolute salinities close to 0 g kg\(^{-1}\) at the surface and 20 g kg\(^{-1}\) at 10 m depth. This distribution highlighted the presence of fresh and estuarine water masses (FW, EFW and ESW) combined with a steep vertical absolute salinity gradient associated with a stable water column (Fig. 4d). Absolute minima of conservative temperature were observed between stations 2 and 3 (6.1 °C) and stations 8 and 9 (6.0 °C). The first of these minima was related to contributions from Manio creek, which originated from ice melting in the Northern Patagonia icefield. The second minimum corresponded to ice melting from Steffen glacier. The highest conservative temperature in the surface layer (~8 °C) was observed at the mouth of the Baker River (between stations 10

**Fig. 3.** Typical profiles of the \(\Theta\), \(S_a\), Buoyancy frequency (\(N^2\)), Turner angle (\(\text{Tu}\)) and density ratio (\(R_p\)) in Steffen-Baker region (upper panel) and in Messier channel (lower panel). Gray dashed line in absolute salinity profiles limits the water masses. The gray-shaded areas in \(\text{Tu}\) and \(R_p\) indicate DL (diffusive-layering) and SF (salt-fingering) zones.
and 11). Although Baker River is fed by cold water of glacial origin from General Carrera Lake (second largest lake in South America), the lake is ~370 km from the Baker River mouth and consequently the river water was heated by the sun along its trajectory. Absolute salinity was close to zero (0.1–0.6 g kg\(^{-1}\)) in the top 4 m of the water column at the mouth of the Baker River, related to the annual average discharge of 1360 m\(^3\) s\(^{-1}\) during the sampling period. In general, the cold freshwater layer, fed by river discharge and
glacier ice melt, was found in the depth range of 0–10 m, typical for the Patagonian fjords and channels (Calvete and Sobarzo, 2011).

Below this cold and fresh surface water, a lower layer (from 10 to 100 m depth) of warmer and saltier water was observed. This layer is a result of the advection of oceanic water from the Gulf of Penas into the Baker channel (Aiken, 2012). Highest conservative temperatures were detected between 15 and 60 m with a core of 12 °C at 25 m, corresponding to MSAAW. The layering appeared between 3 and 20 m at the interface between EW and MSAAW, with Tu between –45° and –55° (Fig. 4e). Diffusive-layer appeared again at ~40 m depth with its highest intensity around station 3 (Tu = –74.2°). This hydrographic transect demonstrates the importance of the interaction between colder freshwater inputs and advection of ocean water in the Patagonian fjords to produce layering in the first 60 m of the water column.

Quantification of vertical double-diffusive convection

To quantify the possibility for double-diffusive convection processes (layering and fingering) in the study area, Turner angle (Tu) values were investigated in the same oceanic depth windows where the density ratio (R_q) exhibited values of R_q = 0–1 to the layering and R_q = 1–3 to the fingering (You, 2002). Double-diffusive convection, as layering or fingering, was identified in approximately 40% of these data (570 of the 583 stations at some depth interval) confirming its presence throughout the year (Fig. 5a and b).

Favorable conditions for layering, i.e. Tu between –90° and –45°, were detected in 39.3% of the measurements and occurred from the surface down to 100 m depth (Fig. 5c). Highest occurrence was observed between 20 and 50 m depth where EW and MSAAW merged and conservative temperature and absolute salinity both increased with depth. When Fresh Water overlaid Estuarine Water, which in turn overlaid Estuarine Salty Water, layering was detected in the upper 20 m of the water column. Modified Subantarctic Water and somewhat saltier SAAW was found below 50 m; mean temperatures of both water masses were quite similar, 9.8 ± 1.1 °C and 9.7 ± 0.4 °C, respectively. At some locations conservative temperature slightly increased from 50 to 100 m, therein giving rise to layering.

When hydrographic conditions for layering were favorable, this process could be characterized as ‘weak’ in most of these records (87.3%, Fig. 5c). The intensity was ‘medium’ in 10.4% of all layering cases and only 2.3% were identified as ‘strong’. In both the medium and strong cases, layering was located in the ESW and the MSAAW. Double-diffusive convection in the form of fingering was only observed in 0.8% of all data analyzed (Fig. 5d).

Horizontal distribution of layering showed that this process is present throughout the Chilean Patagonia area (Fig. 6a–c). Most stations presented layering events (weak, medium and strong), with the southern Patagonia, from 50° to 55°S, showing major occurrences of layering events (Fig. 6a and b). Diffusive-layering defined as ‘weak’ occurred in the water column, with an average of 53 events and maximum of 90 layering events in the south region. In the central area (45–50°S) and the north Patagonia the averages were lower, 33 and 27, respectively.

Fig. 5. (a) Turner angle and (b) density ratio, for all CTD stations presented in Table 1. (c) Histogram of frequency of weak layering (Tu between –60° and –45°, black bars), medium layering (between –75° and –60° Tu, red bars) and strong layering (Tu between –90° and –75°, blue bars) according to You (2002). (d) Histogram of occurrence of SF. The gray-shaded areas in (a and b) indicate DL (diffusive-layering) and SF (salt-fingering) zones. Percentages of each class with respect to the number of total measurements are given in (c and d). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
Density ratios (R-rho) between 1 and 10 were used to compute $K_T$ (Eq. (3)) according to the quantification of layering (Fig. 7a). As fingering occurrence was under 1% of the total CTD and SCAMP data, salt diffusivity fluxes were not considered here. Vertical diffusivity of heat increases from $10^{-7}$ m$^2$ s$^{-1}$ for R-rho = 10 (very weak) to $5 \times 10^{-5}$ m$^2$ s$^{-1}$ for a density ratio of R-rho = 1 (very strong) (Fig. 7b). The largest values of $K_T$ illustrate the vertical heat transfer efficiency of layering, which can be $\sim 100$ times more efficient than molecular heat diffusion (Kelley, 1984, 1990). For the Patagonian fjord region, the occurrence of layering was mostly “weak”, corresponding to a $K_T$ range between $10^{-7}$ and $10^{-6}$ m$^2$ s$^{-1}$. “Medium” layering occurred in 10.4% of all occurrences, with $K_T$ varying from $10^{-6}$ to $5 \times 10^{-6}$ m$^2$ s$^{-1}$. “Strong” layering developed in only 2.3% of the cases, with $K_T$ ranging from $5 \times 10^{-6}$ to $5 \times 10^{-5}$ m$^2$ s$^{-1}$ (Fig. 7c and d).

**Turbulent mixing in Martinez channel, central Patagonia**

**Hydrographic features**

The transect from Martinez Channel to the Baker River mouth (Fig. 4a), carried out during fall 2009 (hydrographic description in Section 4.1), was repeated in December 2011 (late austral spring) using a SCAMP microprofiler. Hydrographic sections in December 2011 exhibited different features than in May 2009 (austral fall) (Fig. 8). Whereas the vertical absolute salinity structure was similar, featuring a 10 m thick upper fresh layer consisting of EW and a lower saltier layer occupied with MSAAW and SAAW, the vertical conservative temperature showed a three layer structure in December (Fig. 8a and b).

The highest conservative temperatures, due to solar heating, were measured in the surface layer (0–10 m) reaching 14.4 °C at station 7; of particular interest was a slight increase in conservative temperature within this layer. Much lower conservative temperatures were observed underneath the stratified layer (Fig. 8c) ranging between 7 and 9 °C. These low conservative temperatures were caused by the input of melt water from the surrounding northern and southern Patagonian Icefields (Fig. 1b). Below the cold second layer, warmer (~10 °C) MSAAW and SAAW followed. The different layering scheme influenced the occurrence of diffusive-layer processes as compared to fall. Surface layering was not observed in fall but was in spring (Fig. 8d). Weak and medium sub-surface layering was apparent in spring, like in fall, at the interface of cold, fresh EW and warmer, saltier MSAAW and SAAW. Isolated strong layering was found in MSAAW and SAAW, with greatest $T_u$ between −80° and −84°, indicating strong layering. In general most layering events occurred in the transition layer between the SAAW and MSAAW water masses (Fig. 8d).

The high vertical resolution of SCAMP microprofiles allows the detection of thermohaline staircases in Martinez and Baker channels (Fig. 9). These staircases (steps of 2–4 m thickness) were observed beneath the pycnocline, between 10 and 40 m (Fig. 9a and b). Conservative temperature and absolute salinity increased up to 0.5–1.0 °C and 0.5–1.5 g kg$^{-1}$, respectively, from one layer to the next and showed solid evidence of layering occurrence in Patagonian fjords.

**Diapycnal eddy diffusivity and dissipation rate of TKE**

In order to understand the mechanisms that produce turbulent mixing in Central Patagonia, 15 SCAMP microprofiles (Fig. 9a) were used to quantify the diapycnal eddy diffusivity using the Thorpe scale (Eq. (5)) and the derived dissipation rate of TKE from SCAMP (Eqs. (8) and (9)) parameterizations (Thorpe, 2005; Osborn, 1980; Cuypers et al., 2011) (Fig. 10).

The Thorpe displacement records showed two areas where the overturns concentrated; one between the surface and ~7 m depth, and the second from 12 m down to 45 m. Large overturns were observed between 15 and 27 m, and between 33 and 43 m (Fig. 10a). A total of 1594 overturns were detected, 280 in the first layer with an average $L_T$ of 0.47 m and 798 overturns between 15 and 43 m, with an average $L_T$ of 0.78 m, and an absolute maximum of 6.52 m (Fig. 10b), detected in station 17. In the sub-surface...
layers, the higher Thorpe scale values coincided with the layering region, where the density ratio was 1–10 (Fig. 10e).

Dissipation rate of TKE from Thorpe (2005) parameterization ranged from $4 \times 10^{-10}$ to $3 \times 10^{-3}$ W kg$^{-1}$ (Fig. 10c, black dots). Maximum values of $\epsilon_{\text{thorpe}}$ were in the range $10^{-5}$ to $10^{-3}$ W kg$^{-1}$ and they coincided with the same region where the Thorpe scale was high, because of the direct relation between both variables. The red dots in Fig. 10c illustrate the results from dissipation rate of TKE derived from SCAMP by fitting the Bachelor spectrum. The $\epsilon_{\text{SCAMP}}$ varied from $1 \times 10^{-10}$ to $8 \times 10^{-4}$ W kg$^{-1}$, similar to $\epsilon_{\text{thorpe}}$. Diapycnal eddy diffusivity ranged from $10^{-6}$ to $10^{-2}$ m$^2$ s$^{-1}$ in both parameterizations (Fig. 10d) and the higher values were also estimated in the surface ($K_{\text{p-thorpe}} = 0.73$ m$^2$ s$^{-1}$) and sub-surface layer ($K_{\text{p-thorpe}} = 0.9$ m$^2$ s$^{-1}$). The minimum values coincided with the position of the pycnocline where high buoyancy fluxes occurred (Fig. 10f) and damped turbulence.

The Thorpe scale values along the Martinez channel better clarified the origin of the surface turbulence. The higher Thorpe scale (2–4 m) was located in the position of the principal fresh water supply in this area, e.g. the Baker River mouth, the Steffen fjord mouth and the Manio creek (Fig. 11a). The second maximum region of Thorpe scale in Martinez channel was observed between 20 and 50 m depth, with a core region around the 30 m at station 8 ($L_T = 3.17$ m). The large turbulent mixing in the sub-surface layer was mainly produced by layering events.

The dissipation rate of TKE ($\epsilon_{\text{thorpe}}$) and diapycnal eddy diffusivity ($K_{\text{p-thorpe}}$) calculated from Thorpe scale (Fig. 11b and c) presented the high value at the same areas as $L_T$. However, station 14 recorded a more homogenous turbulence, with high values of $\epsilon_{\text{thorpe}}$ and $K_{\text{p-thorpe}}$ caused by the vertical structure of $L_T$, which was on average 1 m along the water column. At the same time, the dissipation rate of TKE ($\epsilon_{\text{SCAMP}}$) and diapycnal eddy diffusivity ($K_{\text{p-SCAMP}}$) derived from the SCAMP (Fig. 11d and e) showed a region of similar values (as Fig. 11b and c), highlighting the maximum at the surface and sub-surface layer. However, the intensity of $\epsilon_{\text{SCAMP}}$ was 1–2 orders of magnitude less than $\epsilon_{\text{thorpe}}$. 

Fig. 7. (a) Density ratio for CTD and SCAMP stations in which layering occurred and (b) vertical diffusivity of heat versus density ratio for layering ($1 < R_{\rho} < 10$); (c) Related heat diffusivities, and (d) histogram of frequency of strong (blue) and medium (red) layering; depth range of water masses are added (see Table 2). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)
**Discussion**

**Evidence of DDC in Patagonia**

Analysis of 568 CTD profiles and 15 SCAMP stations taken at different times of the year in the fjords and channels of Patagonia yielded the first double-diffusive-convection (DDC) quantification in this region. The main condition for the formation of DDC in the first 100 m of the water column was: (1) fresh water influx from glacier ice melting and river runoff at the surface combined with low water temperatures, and (2) the presence of warmer, saltier oceanic water underneath. The presence of these water bodies (cold-fresh water overlying warm-salty water) favored the upward transfer of heat from oceanic water masses (SAAW and MSAAW) to estuarine waters, as shown in the heat diffusivity calculations (Fig. 7). Double-diffusive convection occurred as “diffusive-layering” in about 40.0% of the data and as “salt fingering” in <1% (Fig. 5). This was in contrast to the Global Atlas quantifications of DDC, where fingering records represented 30% and layering 14% (You, 2002).

The horizontal distribution of layering events in Patagonia revealed a meridional gradient with more occurrences in the southern region (50–56°S) (Fig. 6). A larger proportion of layering events occurred in the south. Studies of the latitudinal distribution of fresh water input in Patagonia detected two regions that received the majority of the fresh water input, (~4000 m² s⁻¹ from precipitation and ~3500 m² s⁻¹ from river discharge). The first area was located at 46°S and the second at 52°S (Dávila et al., 2012).

![Fig. 8. Hydrographic section along Martinez Channel–Baker River mouth during the SCAMP expedition, on December 19, 2011. (a) Conservative temperature, (b) absolute salinity and (c) buoyancy frequency. (d) Turner angle showing only the occurrence of layering (Tu between –45° and –90°). Isohaline contours in (b) correspond to the boundaries of the water masses identified.](image-url)
The varying rate of fresh water contribution along Patagonia (Fig. 1b) created differences in formation and mixing between estuarine water (cold–fresh water) and oceanic water (warm–salty water), increasing the occurrence of layering events in southern Patagonia. This report of layering processes in Chilean Patagonia fjords complements the world-wide map of DDC created by You (2002) and Kelley et al. (2003).

Thermohaline staircases were detected with high-resolution vertical microstructure profiles in Martinez and Baker Channels (Fig. 9), strengthening the evidence of the occurrence of layering in Patagonia. This thermal staircase had vertical differences that were one order of magnitude greater than those reported in the Arctic Ocean (Schmitt, 2001, and references therein). However, the Patagonian thermohaline staircase was weaker than that detected in the Brasil–Malvinas Confluence (1–4 °C and 0.1–2 g kg \(^{-1}\)) (Bianchi et al., 2002, their Fig. 1). Although a thermohaline staircase was present in most stations, the form of the steps was not always as sharply pronounced as in the Tyrrhenian Sea (Schmitt, 1981) or the Arctic Ocean (Schmitt, 2001, his Fig. 6). Nevertheless, more research is required to determine how layering interacts with the tidal and wind regimes reported in some of the fjords in the region (Valle-Levinson and Blanco, 2004; Valle-Levinson et al., 2006, 2007).

**Dynamic implications of layering**

The strongest vertical diffusivity of heat (\(K_T\), \(10^{-6}–10^{-5} \text{ m}^2 \text{ s}^{-1}\)) in Patagonia developed in the layering layer between 20 and 70 m (\(Tu \sim 90\)° and \(K_T \sim 1\) due to the advection of heat by the warmer–salty oceanic water (MSAAW and SAAW) (some examples are shown in Figs. 2c and d, 4b and c, 8a and b). Similar vertical diffusivities of heat, averaged value of \(K_T = 2.8 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}\) with an average of \(1.9 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}\) below the 400 m depth, were reported in the Arctic Ocean by Sirevaag and Fer (2012). Thus, the maximum value of vertical diffusivity of heat found in the Patagonian fjord region was in the same order of magnitude as previous studies, at some locations and depths. Pertinent to climate variability, the upward heat transfer produced by the occurrence of layering, could contribute to the natural cycle of glacier melting, as reported in many glaciers of the Northern and Southern Patagonia Icefield, e.g. Jorge Montt glacier (Rignot et al., 2003).

Layering events (Fig. 8d) generated turbulent mixing in the sub-surface layer, as demonstrated by the Thorpe scale computation in Martinez and Baker channels (Figs. 10b and 11a). The average \(L_T\) between 15 and 45 m (\(L_T = 0.78\) m) and its absolute maximum (\(L_T \sim 7\) m) were higher than the average \(L_T\) at the surface layer (\(L_T = 0.47\) m). This was caused by the variety freshwater sources, e.g. Baker River, waterfalls of Manio creek and ice melting from Steffen glacier. However, station 14 of SCAMP, located close to the Baker River mouth, presented an average \(L_T\) of 1 m, highlighting the importance of river discharge to produce turbulence in the water column. Stansfield et al. (2001) reported the major number of overturns in the summer sampling of Juan de Fuca strait (\(\sim 48\)°N) between 20 and 120 m deep with \(L_T\) ranging from 0.14 to 5.1 m and an average \(L_T\) of 0.5 m, due to the strong shear generated during the high discharge of the Fraser River. Recent ADCP measurements in the Steffen-Baker region (first 40 m of water column and close to station 10 of SCAMP, Fig. 9) revealed the presence of vertical shears and semi-diurnal internal tides (5–15 m depth) around the pycnocline and not in the position where the “strong” sub-surface layering occurred (Ross et al., 2014, manuscript submitted to Progress in Oceanography, SI: Chilean Fjords [IGeo05274]). Other studies showed the importance of internal tides, generated by shear instability, to produce vertical mixing with \(L_T = 0.48\) and \(K_T = 10^{-4} \text{ m}^2 \text{ s}^{-1}\) (Kitade et al., 2003).

The highest dissipation rates of TKE (\(\varepsilon_{\text{Thorpe}} = 10^{-6}–10^{-3} \text{ W} \text{ kg}^{-1}\)) and diapycnal eddy diffusivity (\(K_{\text{D}}_{\text{Thorpe}} = 10^{-3}–10^{-2} \text{ m}^2 \text{ s}^{-1}\)) calculated using \(L_T\) values were recorded also in the sub-surface layering layer. These were higher than the \(K_T\) values produced by internal tides (Kitade et al., 2003). Using both parameterizations,
Thorpe (2005) and Cuypers et al. (2011), $\epsilon_{\text{Thorpe}}$ and $\epsilon_{\text{SCAMP}}$ were in general higher than values reported in central Arctic Ocean ($\epsilon = 10^{-10}$–$10^{-7}$ W kg$^{-1}$) by Sirevaag and Fer (2012) and by the recent measurements of dissipation rate in southern Ocean ($\epsilon = 10^{-11}$–$10^{-8}$ W kg$^{-1}$) by Sheen et al. (2013), where also diapycnal diffusivity was reported in the order of $K_p = 10^{-5}$–$10^{-3}$ m$^2$ s$^{-1}$.
over the first 500 m depth. A difference in magnitude was observed between the estimations of the dissipation rate of TKE and the diapycnal eddy diffusivity in Martinez channel, both results evidenced the occurrence of high turbulent mixing in the “strong” diffusive-layering (Fig. 11).

Conclusion

For the first time, double-diffusive convection in the form of “diffusive-layering” and “salt fingering” were successfully quantified in the fjords region of Patagonia. Diffusive-layering occurred in 40% of the first 100 m of the water column with most of the layering events located between 20 and 70 m depth. This depth range could be attributed to the presence of fresh-cooler estuarine water masses overlying salty-warm oceanic water. The layering events occurred in all Patagonia, but showed a meridional gradient with less layering events in the northern fjords. The southern fjords were influenced by layering down to 100 m in 50–90% of the observations. This prevalence of layering in the southern fjords was caused by the high fresh water input in this region (~8000 m$^3$ s$^{-1}$ from precipitation, river discharge and ice melting), which increased the influence of estuarine waters to ~90 m depth. In layering events, the vertical diffusivity of heat was on the order of 10$^{5}$ m$^2$ s$^{-1}$. This result was consistent with the lower values of density ratio (1 < R-$\rho$ < 3) reported when strong layering was present.

Profiles obtained with a microstructure profiler (SCAMP) in Martinez and Baker channels (~48$^\circ$S) demonstrated that increased turbulence in a sub-surface layer (around 20–70 m depth) coincided with elevated layering. High values of Thorpe scales ($L_T$ ~ 7 m), dissipation rates ($\varepsilon = 10^{-5}$–$10^{-3}$ W kg$^{-1}$), and diapycnal eddy diffusivity ($K_q = 10^{-6}$–$10^{-3}$ m$^2$ s$^{-1}$) were found at this sub-surface layer. These values provided evidence of the importance of layering as a relevant mixing mechanism. Moreover, results on $\varepsilon$ and $K_q$ profiles should motivate further studies in Chilean fjords at other spatial and temporal scales. Resolution of those scales will improve the understanding of $\varepsilon$ versus layering relationships and of vertical mixing in this region. The landward advection of ocean waters provides heat to Patagonian fjords. This heat may be transferred to near-surface waters by layering. Such process needs to be investigated in the context of climate change and atmospheric variability. The heat supply from the ocean could contribute to the natural cycle of glacier melting, as reported in many glaciers of the Northern and Southern Patagonia Icefield.

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