Available online at www.sciencedirect.com
SCIENCE @ DIRECT®
Progress in Oceanography 56 (2003) 461–481
Progress in Oceanography
www.elsevier.com/locate/pocean

The diffusive regime of double-diffusive convection

D.E. Kelley [a,*], H.J.S. Fernando [b], A.E. Gargett [c], J. Tanny [d], E. Özsoy [e]

[a] Department of Oceanography, Dalhousie University, Halifax, NS, B3H 4J1, Canada
[b] Department of Mechanical & Aerospace Engineering, Arizona State University, Tempe, AZ 85287, USA
[c] Ocean, Earth & Atmospheric Sciences, Old Dominion University, Norfolk, VA 23529-0276, USA
[d] Holon Academic Institute of Technology, POB 305, Holon 58102, Israel
[e] Institute of Marine Sciences, Middle East Technical University, P.O. Box 28, Erdemli, Icel 33731, Turkey

Abstract

The diffusive regime of double-diffusive convection is discussed, with a particular focus on unresolved issues that are holding up the development of large-scale parameterizations. Some of these issues, such as interfacial transports and layer-interface interactions, may be studied in isolation. Laboratory work should help with these. However, we must also face more difficult matters that relate to oceanic phenomena that are not represented easily in the laboratory. These lie beneath some fundamental questions about how double-diffusive structures are formed in the ocean, and how they evolve in the competitive ocean environment.

Contents

* Corresponding author.
E-mail address: dan.kelley@dal.ca (D.E. Kelley).

0079-6611/03/$ - see front matter © 2003 Elsevier Science Ltd. All rights reserved.
doi:10.1016/S0079-6611(03)00026-0

1. Introduction

Of the two types of double-diffusive convection (DDC, henceforth) that may occur if large-scale gradients of salinity *S* and temperature *T* point in the same vertical direction, the salt-finger (SF, henceforth) mode has been studied more extensively than the diffusive-layer (DL, henceforth) mode. There appear to be three broad reasons for this focus, not all of which are compelling.

First, there are historical reasons. DDC research began with the SF mode, in the guise of the 'perpetual salt fountain' (Stommel, Arons, & Blanchard, 1956). This may explain why the seminal and defining paper of Stern (1960) relegated the DL mode to a footnote, and why followup studies dealt mainly with the SF mode.

Second, there are geographical reasons. SF is commonly found at locations that are easily accessible to research vessels, whereas DL is prevalent at high-latitude locations where the logistics of sampling are very demanding. Also, the geographical expanse of SF-susceptible waters exceeds that of DL-susceptible waters. This in itself says little, however, since the high-latitude DL zones may be especially important to the global climate system. Besides, DL may be important to interleaving across the globe.

Third, there are reasons that stem from the difficulty of extrapolating laboratory measurements to the ocean. In the laboratory, the SF mode is more vigorous than the DL mode, e.g. causing larger buoyancy fluxes across interfaces with analogous *T* and *S* steps. However, this implies nothing about the relative strength of SF and DL fluxes in the ocean, since the *S* and *T* steps need not be analogous, even if the background gradients are analogous. (Indeed, relating *S* and *T* steps to background gradients is a key goal discussed at some length below.) Another factor is the response of SF and DL to shear, which is common in the ocean but not in laboratory models. While SF transports are apparently inhibited by shear (Linden, 1974b; Kunze, 1994), it has been speculated that DL transports may not be (Padman, 1994). Given such things, it seems that the dominance of the SF mode in laboratory may not be relevant to the ocean.

Reasons such as these may explain the historical concentration on the SF mode, but they should not preclude future DL work. This case is underlined by the current interest in the role of high-latitude oceanography in the climate system (Walsh & Crane, 1992). A key climate component is watermass formation in northern seas, and DL fluxes may play an important role in this process (McDougall, 1983). Further evidence is provided by decades of observations of robust DL signatures in the Arctic, going back to Neal, Neshybya and Denner (1969). DL is also thought to be the main DDC driving agency of Arctic intrusions (May & Kelley, 2002). These intrusions, considered to be a key aspect of Arctic thermodynamics, have remarkably coherent intrusive signatures, traceable over basin scales (Perkin & Lewis, 1984) and perhaps over decades (cf. Carmack et al., 1997). They have also been implicated in the large changes currently taking place in the Arctic (Carmack et al., 1995a,1995b). Taken together, these things suggest that DL may be an important element of the global climate system.

With this potential importance in mind, we turn next to a discussion of DL signatures in the world ocean. After presenting an overview of susceptible ocean regions (Section 2.1) and some thoughts on methods of detecting staircases (Section 2.2), we offer an abbreviated list of some prominent DL examples that have been well sampled (Section 2.3). Then, in the second half of the paper, we turn to matters of modelling DL. We start with the issue of parameterizing fluxes between a pair of DL layers in the laboratory (Section 3.1) and then move on to the related, but more difficult, matter of parameterizing fluxes in oceanic

DL staircases (Section 3.2). This foundation is used for a cursory outline of some unresolved questions regarding oceanic staircase formation (Section 3.3) and evolution (Section 3.4).

Our focus is entirely on the DL case, and mostly on the staircase mode. Readers with wider interests may consult other contributions to this volume, along with various reviews (Turner, 1974; Huppert & Turner, 1981; Turner, 1985; Schmitt, 1994; Fernando & Brandt, 1994), and themed collections (Brandt & Fernando, 1995), as well as recent GCM sensitivity studies exploring the role of DDC under a range of surface boundary conditions (Zhang, Schmitt, & Huang, 1998; Merryfield, Holloway, & Gargett, 1999; Zhang & Schmitt, 2000). Although aspects of the DL mode appear to be simpler than those of the SF mode, it is not clear whether this is because the DL mode is actually simpler or because the DL has been so little studied that contradictory information has not been uncovered. We take up such issues at the end.

2. Observations

2.1. Susceptible regions

Laboratory studies indicate that DDC produces coherent structures, e.g. staircases and intrusions (together with mixed modes) that may be recognized from examination of the horizontal and vertical variation of S and T. Unfortunately, the signatures have length scales (meters to tens of meters) that are below the resolution of ocean atlases. Therefore, we cannot easily construct maps of global staircase occurrence, and are left to estimate the world-wide prevalence of DDC by mapping susceptibility.

Building upon theory and laboratory work, this may be done by comparing the patterns of S and T stratification at large scales. The DL mode of DDC is possible when S and T increase together with depth, and both theoretical and laboratory observations suggest that it is more intense when S and T nearly compensate in density terms. Given the thermal expansion rate α and the haline contraction rate β, then, DL is expected to be most intense when the density ratio

$$R_\rho = \frac{\beta \partial S / \partial z}{\alpha \partial T / \partial z} \tag{1}$$

is near the lower end of the DL-susceptible range $1 < R_\rho < \infty$. Buoyancy fluxes across laboratory DL interfaces drop by an order of magnitude as R_ρ is increased from 1.5 (a typical lower bound for the ocean) to 10 (see Fig. 4 below), suggesting that the range $1 < R_\rho < 10$ provides a good criterion for susceptibility to DL convection.

Fig. 1 illustrates world-wide DL susceptibility according to this criterion, calculated with vertical gradients of T and S from global atlases (Levitus & Boyer, 1994; Levitus, Burgett, & Boyer, 1994). To prevent difficulties with gradient computation, waters above 50 m and below 2500 m depth have been excluded from this analysis. Sparse sampling at high latitudes prevents great reliability in the map at high latitudes, especially in the Arctic. However, the expected pattern of enhanced susceptibility at high latitudes, clearly seen in numerous high-latitude case studies, is unmistakable even in this coarse presentation. In addition, there are regions of susceptibility in the western portions of the northern subpolar gyres, as well as in an Atlantic region near South America.

The diagram indicates regions that are susceptible to DL at any depth in the water column. This is useful for a summary diagram such as this, since a map at any particular level misses some key regions. For example, the northern polar regions have $R_\rho < 3$ (i.e. highly unstable) through the depth range of 200–300 m. The subpolar regions (e.g. the Labrador Sea and the Sea of Okhotsk) have similarly low R_ρ values in a depth range that extends 100 m closer to the surface. The depth range is extended somewhat in the southern polar regions, covering 100–400 m. In addition to these high-latitude cases, there is also a DL-

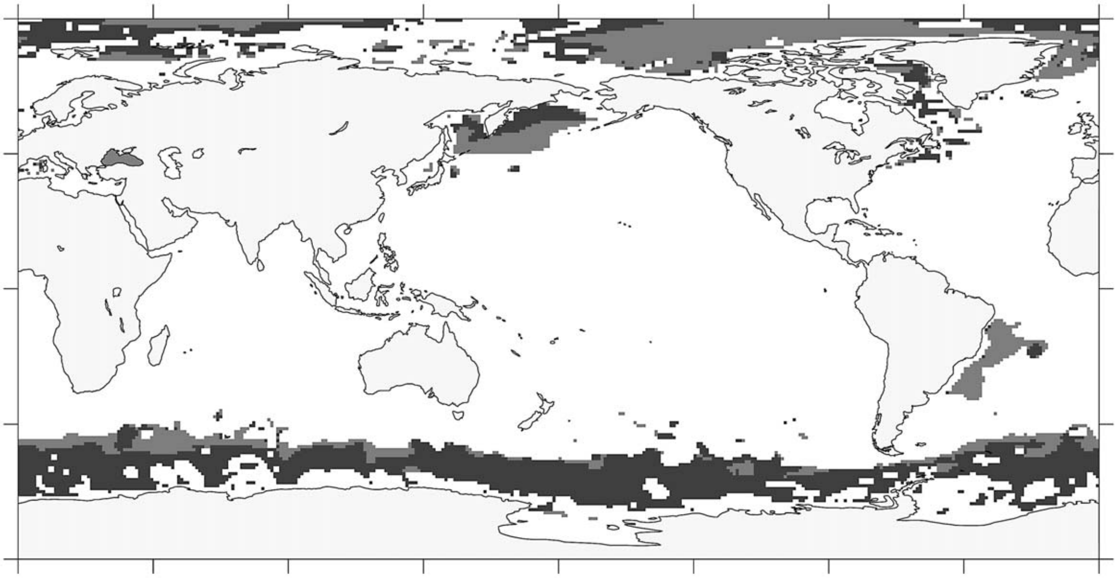

Fig. 1. Ocean regions that are susceptible to DL convection. The light gray indicates areas with density ratios in the range 3–10, somewhere in the water column, and the dark grey indicates $1 \leq R_\rho < 3$.

susceptible region to the east of Brazil. This is much deeper (1000–1400 m) than in polar waters, presumably resulting more from juxtaposition of water masses (Bianchi, Piola, & Collino, 2002) than from surface forcing. It also has somewhat higher density ratios, with $R_\rho \approx 5$. All other things being equal, these higher values might indicate that DL is somewhat less likely in the Brazil region than in the other regions of susceptibility. However, there is no reason to expect all other things to be equal. For example, disruptive effects such as turbulence could be weaker in deeper water.

Thus, diagrams such as Fig. 1, indicating susceptibility, cannot be taken as a trustworthy indication of incidence. One might hope that comparing maps of susceptibility with maps of incidence would shed light on the mechanism of formation and evolution of DL (and DDC in general) in a competitive ocean environment. Unfortunately, the mapping of incidence is challenging. We take this up in detail in the next section.

2.2. Detection methods

We lack an agreed-upon quantitative method of recognizing signatures of DDC structures in the ocean, in the sense that there is no parameter that is routinely calculated and compared from experiment to experiment. This stands in the way of global mapping exercises and the comparison of case studies.

For the construction of atlases as well as first-order descriptions of individual study regions, it would be useful to have a single measure of meter-scale 'steppiness' for a whole pycnocline. Several possibilities for this come to mind. One is to examine the difference between observed profiles and artificial profiles constructed by smoothing over large vertical scales (Galbraith & Kelley, 1996). Mixing an initially linearly stratified profile over a thickness D increases the potential energy per unit mass by $D^2N_0^2/12$, where N_0 is the initial buoyancy frequency. Therefore, computing potential energy anomalies gives information not just on the energy itself, but also on an integral estimate of layer thickness, perhaps estimating N_0^2 by smoothing the observed density profile. Distinguishing the regular steppiness of staircases from the irregular steppiness of turbulent mixing might require further steps, e.g. examination of the depth variation of the integral lengthscale.

For the identification of individual layers, e.g. for flux computation and for tracing staircase extents, other methods might be more useful. These could be based on qualitative signatures that have been used informally for decades, based on finescale signatures on *S* and *T* profiles and in TS diagrams (Tait & Howe, 1968; Neal, Neshyba & Denner, 1969; Federov, 1970). Steppy profiles are often taken as evidence of thermohaline staircases of DDC origin, but some provisos are required. First, the vertical *S* and *T* gradients should vary in intensity, but not in sign, over depth. Otherwise, interleaving could have caused the finestructure. Second, the *S* and *T* gradients should be of the DDC variety, since otherwise internal waves (Lazier, 1973; Lazier & Sandstrom, 1978) or mixing striations (Phillips, 1972) could have caused the finestructure. Third, the patterns should be traceable laterally and temporal; *TS* diagrams may be especially useful in this regard since tracing features in profiles can be difficult unless sampling is very rapid (Padman & Dillon, 1988). Furthermore, *TS* diagrams are effective in distinguishing staircases from interleaving. This is important since sometimes the difference between signatures of staircases and interleaving is slight, if only profiles are examined: weak reversed gradients are difficult to distinguish from weak unreversed gradients. Since interleaving scales tend to be much larger than layer scales in staircases, misinterpreting intrusions for layers has the potential to confound the development of scaling laws for layer thickness and thus for large-scale diffusivities (see Section 3.2 below).

Fig. 2 provides a good example for discussion. It results from CTD profiling in the Arctic Eurasian Basin. The main pattern is that *T* increases with depth (as does *S*, not shown here), owing to the presence of a relatively warm and salty Atlantic watermass that enters the Arctic at a depth of approximately 400 m. Thus, the column is DL-susceptible. Steppiness is evident through much of the displayed depth range, especially in the lower half. Resolvable gradient reversals are rare in this portion of the water column, so the presence of a staircase seems likely. The inset shows a 20 m tall portion of the profile, along with two possible methods for determining the geometry of this presumed staircase.

The first method, displayed using vertical gray bars in the diagram inset, is a layer-detection scheme patterned on that used by Padman and Dillon (1988). It is based on the histogram of temperature. The basic parameters are the sampling interval δz and the temperature gradient $\partial\overline{T}/\partial z$ smoothed over the depth range of interest. The product of these quantities provides a natural histogram bin width δT, in the sense that we would expect a linear profile to have one sample in each such bin. However, a layer within a

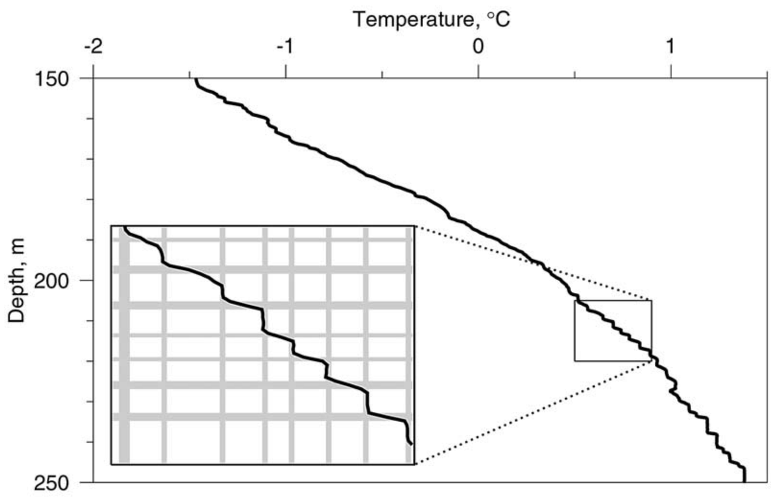

Fig. 2. Temperature profile at station 508 of the Eurasian Basin Experiment (Perkin & Lewis, 1984). Magnified in the inset is a region from 205 to 220 m depth, which has several easily identified layers. The vertical and horizontal lines under the inset profile indicate the results of two staircase-detection methods discussed in the text.

steppy profile should have several samples in its histogram bin, the number depending on the ratio of the layer thickness to δz. In the present case $\delta z = 0.25$ m and $\partial \bar{T}/\partial z = -0.027$ °C, so $\delta T = 0.007$ °C was used for the T histogram. The idea is to flag temperature bins with high occurrence rates. Therefore the main adjustable parameter in the method is the minimum number n of elements that we demand in a bin before declaring a layer to have been detected. A natural scale for n is $\hat{H}/\delta z$, where $\hat{H}$ is an initial estimate of the layer thickness (perhaps inferred visually or from the integral method). For the region of the inset $\hat{H}\approx 2$m so $n\approx 10$. Writing $n = a_1\hat{H}/\delta z$, we tested sensitivity with trials using $a_1 = 0.3$, 0.4, and 0.5, and found the results to be robust across this range. Fig. 2, with $a_1 = 0.3$, shows the method agrees well with layers detected visually. Clearly, the task would have been impossible with CTD data sampled at $\delta z = 2$ m, since the layers are only 2 m thick. Herein lies a major difference between the SF and DL cases, since SF layers are often tens of meters thick, and are easily detected with archival CTD records having $\delta z = 2$ m, whereas DL layers are often so thin as to be difficult to resolve even with raw CTD records. This explains why the best examples of DL in the ocean employ microstructure instrumentation, often from stable (ice) platforms that are immune to wave heaving.

The second method is based on gradients. One may focus on either high-gradient zones (interfaces) or low-gradient zones (layers). For Fig. 2 we searched for interfaces, flagging them as regions in which centered first-difference estimates of $\partial T/\partial z$ exceed the large-scale value $\partial \bar{T}/\partial z$ by some factor of a_2. The horizontal gray bars in the inset of the figure show the results with $a_2 = 1.5$. The visual agreement is reasonably good in terms of interface spacing but not in terms of interface thickness. In this case, the results are quite sensitive to the choice of a_2, and a glance at the figure shows why (e.g. consider the second shallowest interface shown in the inset). In our tests, the gradient technique proved less robust than the histogram technique, in terms of parameter sensitivity. This may not be a general result, but some general explanations come to mind. For one thing, gradient calculations are sensitive to 'noise'. For another, the profile (and others like it) seems to indicate that homogeneity of layers is more reproducible from layer to layer than is interface thickness. This may be because of poor resolution in the present example. For example, the second interface in the inset could be an example of a 'split layer' (see Section 3.2 below).

Tests with varied profiles should expose the relative power of these and other proposed techniques, and indicate choices for the parameters (e.g. a_1 and a_2 here) that are optimal for various profiling instruments and various physical regimes. The hope is that common techniques used in different studies will facilitate DDC incidence mapping, and that comparing these maps with maps of susceptibility and of forcing will shed light on the dynamics of DDC in the ocean. In the meantime, we are left to examine case studies. This task we take up in the next section.

2.3. Prominent DL examples

2.3.1. Lakes and trenches

The geophysical systems that seem most similar to the early laboratory experiments with bottom-heated salt gradients (see section 3.3.2 below) are geothermally-heated salty lakes and semi-isolated deep trenches.

In some lakes, the salt stratification is sufficient to stabilize partially the water column against the action of bottom heating. Several such cases have been studied. The first was Lake Vanda in Antarctica (Hoare, 1966; Shirtcliffe & Calhaem, 1968; Hoare, 1968; Huppert & Turner, 1972). Soon after, DL signatures were found in Powell Lake, a remnant fjord off British Columbia (Osborn, 1973), and in Lake Kivu in Africa (Newman, 1976). The signature in such cases can be striking; for example, Newman (1976) counted 150 regular layers in a single profile.

A similar situation is that of small pockets of saline water trapped near the ocean bottom in the trenches or 'brine lakes', where there is a supply of geothermal heat either from sediments or hot vents. These systems often display evidence of DL in the form of sharp interfaces between homogeneous deep water masses. Some of the earliest studies of this were done in the Red Sea (Swallow & Crease. 1965; Voorhis &

Dorson, 1975). These interfaces are very stable, with $R_\rho \approx$ from 9 to 25 according to Turner (1969). Similar brine lakes have also been found in the deep eastern Mediterranean basin by Boldrin and Rabitti (1990), wherein the density ratio was even higher, $R_\rho \approx 600$, and the interfacial heat flux was estimated to be of the same order as the geothermal flux. These facts together with the isotopic composition of lake waters (Stenni & Longinelli, 1990) suggest that the water in the bottom layer is of fossilized origin, developed approximately 4500 years ago. Similar features were found in other Eastern Mediterranean brine lakes (MEDRIFF Consortium, 1995), which also showed the recent invasion of a new dense water mass of Aegean Sea origin. After the renewal event, the more saline deep waters become warmer but less saline than the fossil water below the interface, making the interface stable. It appears that brine lakes can retain a memory of deep water history of a basin, which can be inferred from the water mass structure. McDougall (1984) investigated brine layers capped by a DL interface and supplied by hot saline plumes in a depression, as for the Red Sea brine layers. He found the system to converge to a stable fixed point in *TS* space, perhaps explaining their observed stationary character.

2.3.2. High-latitude zones

Foster and Carmack (1976) observed distinct DL staircases in the Weddell Sea, and further details of the spatial extent of layering in the central and western Weddell Sea have been provided by Muench, Fernando, & Stegen (1990) and Robertson, Padman and Levine (1995) respectively. The latter authors pointed out that the vertical fluxes are so low as to be near the detection limit of dissipation-rate profilers. This leads to questions as to whether DL occurs only when disruptive agencies are weak.

Indeed, some of the most compelling evidence of DL has come from the Arctic, which is another weakly-mixed region. The earliest evidence was provided by microstructure profiling carried out under Ice Island T-3 (Neal, Neshyba & Denner, 1969; Neshyba, Neal, & Denner, 1971). Similar patterns were found in later Arctic sampling (Padman & Dillon, 1987, 1989; Muench, Rudels, Bjork, & Schauer, 1997). As is the case for the Antarctic, the staircase fluxes are weak compared with mixing rates expected in the open ocean. Padman and Dillon (1987) estimate thermal fluxes to be just ten times molecular fluxes. One might compare this with a factor of 100 typically found for turbulent mixing in the open sea (Ledwell, Watson, & Law, 1993; Kelley & Van Scoy, 1999). Perhaps more relevant is a comparison with local heat balances. Padman and Dillon (1989) estimate that Arctic DL fluxes are roughly two times larger than turbulent fluxes. This argues that DL is important, but since the sum of these vertical fluxes is still much too small to account for historical estimates of Arctic heat losses, Padman and Dillon (1989) suggest that the balance may be determined by benthic mixing along the continental slope followed by isopycnal diffusion into the interior. Such a scenario is familiar in the wider context of studies of ocean mixing, in which vertical turbulent fluxes in the ocean interior are too small to explain inferred large-scale balances (Munk, 1966; Munk & Wunsch, 1998). As for turbulent mixing in the ocean interior, then, it is not clear whether DL mixing is a rate-limiting step, directly or indirectly.

The indirect mode is possible through watermass transformation, an example of which may be found in the Greenland Sea. McDougall (1983) has argued that DL fluxes may govern the rate of formation of Greenland Sea Bottom Water (GSBW), a major factor influencing the strength of the global ‘conveyor belt’ . His analysis suggests that GSBW is formed by sub-surface modification of warm salty Atlantic water, flowing into the center of the gyre beneath colder fresher Greenland Sea surface water. A simple model of this process, incorporating differential DL diffusion of *T* and *S*, predicts a rate of formation of GSBW close to that observed. This interpretation, which should be tested in high-resolution numerical models, suggests that DL could be of major importance to the properties of the global ocean.

2.3.3. Black Sea

Sheltered regions without major forcing are conducive to DL and have been the major showcases of oceanic diffusive layers. A compelling example is provided by the Black Sea, the largest land-locked basin of the world with a positive surface water budget (Özsoy & Ünlüata, 1997, 1998). Its near isolation, despite the some two-way exchange through the Turkish Straits, has led to stagnant conditions and anoxia in its deep waters, the average age of water increasing to more than 2000 years near the bottom (Östlund, 1974; Grasshoff, 1975). The warm and saline Mediterranean Water is the historical source of the deep waters, evolving since the end of the last ice age when the Black Sea was opened to the Mediterranean (Stanley & Blanpied, 1980). Diffusive convection in the Black sea is facilitated by the warm, saline Mediterranean water intruding from the Bosphorus, a diffusive interface capping the bottom convection layer, and vertical fluxes in the deep waters. The stratification of the entire water column indicates DL susceptibility, with $R_\rho \approx 2$ at the top (below the mixed layer) and bottom (above the bottom convective layer), although intermediate depths have much higher values (Özsoy, Ünlüata, & Top, 1993).

The dense waters from the Mediterranean, cascading down the continental slope of the Black Sea, generate intrusions of double-diffusive character that penetrate into the interior at depths of 100–500 m (Özsoy, Ünlüata & Top, 1993; Özsoy & Besiktepe, 1995). The DL-susceptible interior may favor DL convection over SF convection (Turner, 1978), in the context of a side-wall buoyancy source (heat and salt) introduced into the stratified environment (Tsinober, Yahalom, & Shlien, 1983; Tanny & Tsinober, 1988). Boundary mixing (Garrett, 1990; Woods, 1991), in this case driven by the double-diffusive convection (Özsoy, Ünlüata & Top, 1993; Özsoy & Besiktepe, 1995), may be the dominant ventilation mechanism across the permanent halocline of the Black Sea, given the inferred transport of shelf-derived materials, including dissolved nutrients, Chernobyl tracers and inorganic particulates between surface and deep layers (Buesseler, Livingston, & Casso, 1991; Buesseler & Livingston, 1997; Codispoti, Friederich, Murray, & Sakamoto, 1991; Özsoy, Top, White, & Murray, 1991; Özsoy, Ünlüata & Top, 1993). The relatively rapid penetration of tracers across the halocline has been confirmed by measurements of Tritium (Top, Östlund, Pope, & Grall, 1991) and Carbon-14 (Östlund, 1974; Östlund & Dyrssen, 1986). The inclusion of plume parameterizations (which represent vertical mixing) in general circulation models has significantly improved the simulation of ventilation and tracer distributions in the Black Sea (Staneva & Stanev, 1997).

The bottom convection layer of the Black Sea, with a thickness of approximately 450 m, is the largest known of its kind in the world's oceans. A destabilizing geothermal heat flux of approximately 0.040 W/m^2 at the bottom (Zolotarev, Sochel′Nikov & Malovitskiy, 1979) acts against an otherwise stable salinity stratification. The transport between the bottom convective layer and the overlying waters occurs through a single DL interface. The observed homogenization of water properties vertically and across the basin inside the convective layer is expected to occur in about 40 years, which may also have a bearing on the homogeneity in deposition/diagenesis of bottom sediments (Özsoy, Top, White & Murray, 1991; Özsoy & Besiktepe, 1995). Based on the comparison of flux ratios above the DL interface and in the water column, DDC is most likely to be the main vertical transport mode for heat and salt in the deep waters extending from the lower part of the pycnocline to the diffusive interface above the bottom convective layer (Özsoy, Top, White & Murray, 1991; Murray, Top, & Özsoy, 1991). The observed features and environmental conditions suggest that the convective layer is far from the initial phase of convective layer growth identified in laboratory studies (Turner, 1968a; Fernando, 1987; Huppert & Linden, 1979) and corresponds instead to the long time limit of the 'low stability regime' (Fernando, 1987, 1989a; Fernando & Ching, 1991), in which interfacial entrainment, and therefore the layer growth, become negligibly small (Özsoy, Top, White & Murray, 1991; Özsoy & Besiktepe, 1995). The heat flux at the stable Black Sea interface estimated from the Huppert and Linden (1979) and Fernando (1989a) models was 5–8 times larger than the geothermal heat flux from the bottom, but these differences can be reconciled considering the fact that the above models typically overestimate fluxes when $R_{i*} > 240$ (Fernando, 1989a). Here $R_{i*} = \Delta b h_s / w_3$ is a Richardson number based on the buoyancy step Δb, the layer thickness h_s and a vertical velocity scale w_s for convection.

3. Modelling

3.1. Fluxes between two layers

Early DDC laboratory work with thin interfaces between well-mixed layers suggested that vertical heat fluxes are determined not by the details of the interfaces, but rather by the temperature and salinity differences, δT and δS, between the two layers (Turner, 1965). (For brevity, we focus here on the heat flux. Salt fluxes are typically assumed to be given by the product of the heat flux and a dimensionless function, the 'flux ratio', that depends on layer-based density ratio and the background turbulence (Turner, 1968a; Huppert, 1972; Linden, 1974a; Crapper, 1975, 1976; Linden & Shirtcliffe, 1978; Takao & Narusawa, 1980; Narusawa, 1986; Taylor, 1988; Fernando, 1989a).)

For the SF case, such 'layer-based' flux laws have been contradicted by oceanographic microstructure observations made during the C-SALT field program (Gregg & Sanford, 1987; Lueck, 1987) and by new theoretical treatments that account for background shear (Kunze, 1987, 1994). The error in using layer-based laws as calibrated in the laboratory is large, e.g. a factor of 30 in C-SALT. In this respect the DL case is in complete contrast with the SF case, since oceanographic microstructure measurements of DL interfaces confirm laboratory layer-based flux laws (Melling, Lake, Topham, & Fissel, 1984; Padman & Dillon, 1987). This fundamental variance between the DL and SF cases may result from differences in how the interfaces react to shear (Padman, 1994), or just from the fact that DL interfacial shears are lower, e.g. Arctic DL interfaces experience velocity steps of 0.003 m/s (Padman, 1994) while Caribbean SF interfaces experience 0.014 m/s (Kunze, 1994). Either way, we are left with a working hypothesis that ocean DL fluxes may be predicted in terms of layer properties, ignoring interface properties.

However, there are still significant questions about the form of layer-based flux laws. In the ground-breaking analysis of Turner (1965) it was suggested that DDC fluxes should be analogous to a simple model of single-component convective fluxes. In this model, it is assumed that the heat flux through a convecting layer is independent of the layer thickness H, so that dimensional analysis yields a flux law of the form

$$\mathrm{Nu} \propto \mathrm{Ra}^{1/3}, \tag{2}$$

in which the Nusselt number Nu is the ratio of the convective flux Q to the conductive flux $\kappa\Delta T/H$ and the Rayleigh number $Ra = g\alpha\Delta TH^3/(\nu\kappa)$ measures the boundary forcing. (Here g is the gravitational acceleration, ν is the viscosity, and κ is the diffusivity.) This relationship yields $Q\propto\Delta T^{4/3}$ for thermal convection. The Turner (1965) extension for DDC adds a factor $C = C(R_\rho')$ related to layer-based density ratio, yielding $\mathrm{R}'_\rho = (\beta\Delta S)/(\alpha\Delta T)$, yielding

$$Q \propto C\cdot(\alpha\Delta T)^{4/3}. \tag{3}$$

Equations of this form have come to be called '4/3 flux laws'. Although commonly used by oceanographers for DDC calculations, they are based on an uncertain foundation, since (2) is demonstrably false for single-component convection. For example, in the case of thermal convection in water, a suite of laboratory studies contradict the exponent 1/3 in (2) over a wide range, although a 1/3 law may be tenable for a certain Ra range. Consider Fig. 3, which summarizes several studies in the range $10^5 < Ra < 10^9$. Fitted to the form $Nu\propto Ra^p$, these data suggest $p = 0.284$, and certainly not $p = 1/3$, given the statistical error bars. This matches the prediction, $p = 0.27 \pm 0.02$, of the simple convection-cell model of Kelley (1990), as well as the analysis of Castaing et al. (1989), and the results of direct numerical simulations by Kerr (1996). Sommeria (1999) suggests that this exponent holds up to $\mathrm{Ra}\sim10^{13}$ in thermal convection, whereupon the exponent rises to 1/2. More laboratory work will be required to settle the matter for arbitrary Rayleigh numbers.

Turning to DDC, the question is whether the generalized flux law

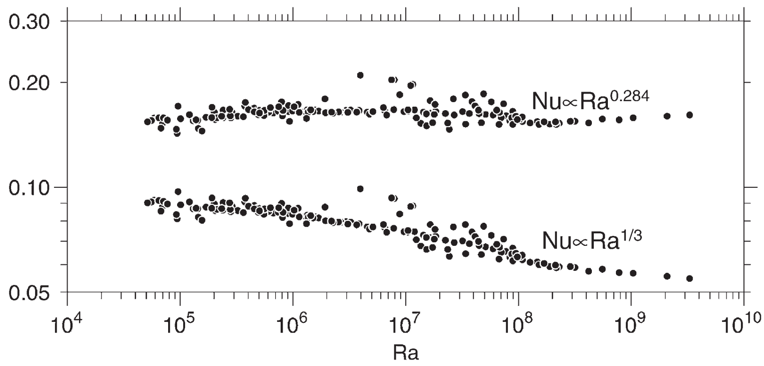

Fig. 3. Empirical refutation of the conventional $Nu \propto Ra^{1/3}$ flux law for thermal convection in water. The dots indicate measurements from five laboratory studies as compiled by Kelley (1990), limited to the *Ra* region of time-dependent convection (Krishnamurti, 1970). The two clusters illustrated arise from using two values for the exponent in a more general flux law $Nu \propto Ra^{P}$. Using $p = 1/3$ leaves a trend, whereas $p = 0.284$ collapses the data (Kelley, 1990).

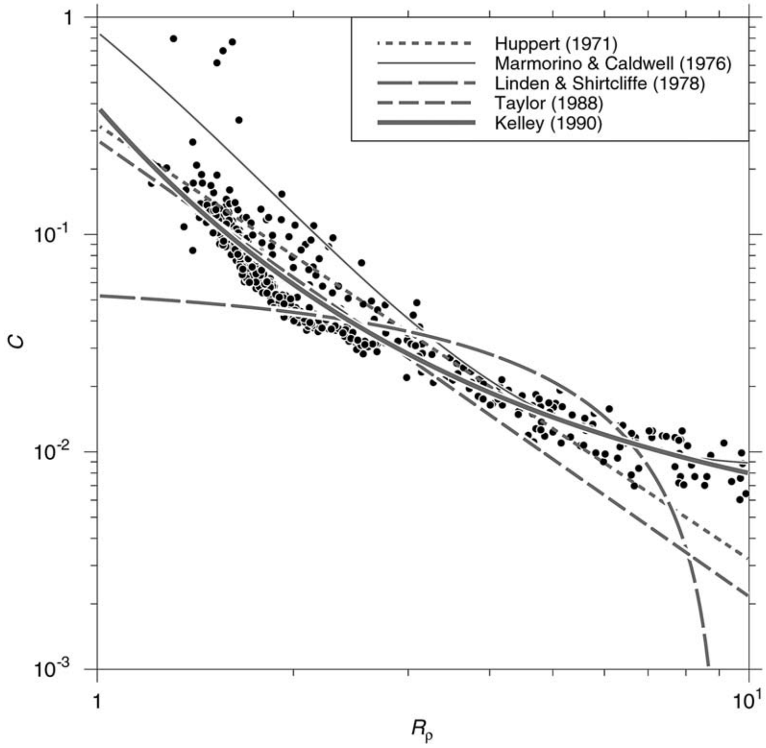

Fig. 4. Coefficient $C(R_\rho{}')$ in DL flux law, as extracted from reports of various laboratory experiments (Turner, 1965; Shirtcliffe, 1973; Crapper, 1975; Marmorino & Caldwell, 1976; Newell, 1984; Taylor, 1988). The dashed curve is the empirical relationship proposed by Marmorino and Caldwell (1976) based on their own measurements (shown in isolation in the inset), while the solid curve is that proposed by Kelley (1990), based on all measurements shown here.

$$Q \propto C\cdot(\alpha\Delta T)^{(1+p)}, \tag{4}$$

has $p = 1/3$ as conventionally assumed, or perhaps a lower value to match thermal convection. For the SF case, Schmitt (1979, Table 3) reported equivalent p values from 0.24 to 0.37, i.e. spanning both exponents under discussion. In another laboratory SF study, McDougall and Taylor (1984) found p = 0.23, although experimental scatter prevented them from rejecting an exponent of 1/3. In the DL case, there seem to be no laboratory studies with enough statistical power to measure p to the resolution required. Although Marmorino and Caldwell (1976) reported $p = 0.33 \pm 0.01$, which would appear to confirm the Turner (1965) hypothesis, it seems the p range was determined by a graphical technique. Our re-analysis of the original measurements (Marmorino, 1974, Table V) yields a 95% C.I. of $0.27 < p < 0.47$, based on boot-strapped linear regression of logged values (Venables & Ripley, 1999). This large range thwarts the desired test on p.

In summary, the 1/3 exponent is untenable for single-component convection, and questionable for the less-studied DDC case. This uncertainty in exponent leads to two practical questions, one relating to the scatter of $C(R_\rho')$ in laboratory experiments, the other to extrapolation from laboratory to the ocean.

The first question is whether the scatter in $C = C(R'_\rho)$ could result from intercomparing different experimental results with erroneous flux laws. In reports of laboratory work this quantity is calculated from the ratio of Nu to $Ra^{1/3}$. Therefore, if the true exponent has a different value, errors will result in comparing studies with different values of Ra. Fig. 4 illustrates that C has a scatter of a factor of ≈40%. Could this result from the use of an incorrect exponent in comparing experimental runs with different Ra values? The answer lies in computing, $\xi^{p-1/3}$, where ξ is the ratio of Rayleigh numbers in two cases. Within a given experiment it is common for Ra to range greatly, e.g. from 1×10^{10} to 2×10^{11} in the Marmorino and Caldwell (1976) setup. Thus, taking $\xi = 20$ and $p = 0.28$, we could expect a scatter of 17% in C values. This is about half the observed scatter, and the value matches the observed scatter if $|p-1/3|$ is made as large as 0.12. This deviation is comparable to the C.I. for the Marmorino and Caldwell (1976) measurements, suggesting the possibility that a large part of intercomparison scatter might indeed result from the use of erroneous flux laws.

The second question is whether extrapolating erroneous flux laws from the laboratory to the ocean will incur significant errors. The wide range of ocean layer thickness leads to an even wider range of Ra, but for a representative value, we may pick $\Delta T = 0.32$ °C and $H = 2$ m for the Arctic (Padman, 1994, Table 1). This yields $Ra = 3\times10^8$. Compared with the high end of the Ra range in the experiments of Marmorino and Caldwell (1976) and using $p = 0.28$, this implies that the 4/3 flux laws underestimate oceanic fluxes by 40%. This is small compared with the factor of 2 variation between the Marmorino and Caldwell (1976) and Kelley (1990) formulations of $C = C(R'_\rho)$, illustrated in Fig. 4. This suggests that using the 4/3 flux laws, while perhaps incorrect in principle, may not cause major problems in practical oceanographic application. The larger issue, and the more important one from the point of view of parameterization, is what controls layer thickness in oceanic staircases.

3.2. Fluxes within staircases

If vertical fluxes are governed by the layer thickness H and the interfacial steps ΔS and ΔT, then since the steps are given by the product of the large-scale gradients and H, it follows that fluxes could be parameterized in terms of large scales if only H can be so parameterized.

This issue was first taken up by Kelley (1984), who proposed on dimensional grounds that layer thickness should scale as $H_K = (\kappa/\bar{N})^{1/2}$, where $\bar{N}$ is the large-scale buoyancy frequency. The ratio $G = H/H_K$ should thus depend only on molecular properties of the fluid and the large-scale gradient density ratio $\bar{R}_\rho$. (Implicit in such a formulation in terms of properties of the water column is that no length-scale is imposed by other agencies, whether during formation or during evolution; see sections 3.3 and 3.4 below.) If the molecular properties of seawater are similar from location to location, then G should depend only on $\bar{R}_\rho$, and a test of the scaling would be to see whether this scaling collapses oceanographic observations. The Kelley (1984) study suggested that this is indeed the case for the DL mode of DDC. His later study suggested that the resulting quasi-empirical relationship could be explained in terms of a model of conditional layer splitting (Kelley, 1988). Fig. 5 shows the data used by Kelley (1984), together with the results of other studies. The observations of Fedorov (1988) are in rough agreement with those of Kelley (1984). However, those of Padman and Dillon (1987) disagree significantly, by a factor of ≈ 3 in layer thickness and thus by a factor of ≈ 4 in fluxes. This disagreement exceeds the uncertainty in fluxes (Fig. 4) and is thus sufficient to motivate further study. Adding more observations to a figure of this form would be helpful, as would further theoretical and laboratory work. For example, the Kelley (1988) model suggested that H is controlled by a process of layer splitting, in which new layers are formed from existing interfaces. Signatures of such a process are common in oceanographic profiles (see Fig. 7 below), but since there have been no direct observations of time evolution in the ocean, laboratory work may be the preferred next step. One advantage of such studies of staircase evolution is that they may also shed light on the mechanisms by which staircases are formed. As the next section illustrates in some detail, the topic of formation is problematic not because of a lack of theories, but rather because of a surfeit.

3.3. Staircase formation

3.3.1. Negative-diffusivity mechanism

In contrast to other mixing agencies, DDC transports density up gradients. This may be expressed as a negative density diffusivity, i.e. $K_\rho < 0$ in an evolution equation of the form

$$\frac{D\rho}{Dt} = \frac{\partial}{\partial z}\left(K_\rho \frac{\partial \rho}{\partial z}\right) + \ldots \tag{5}$$

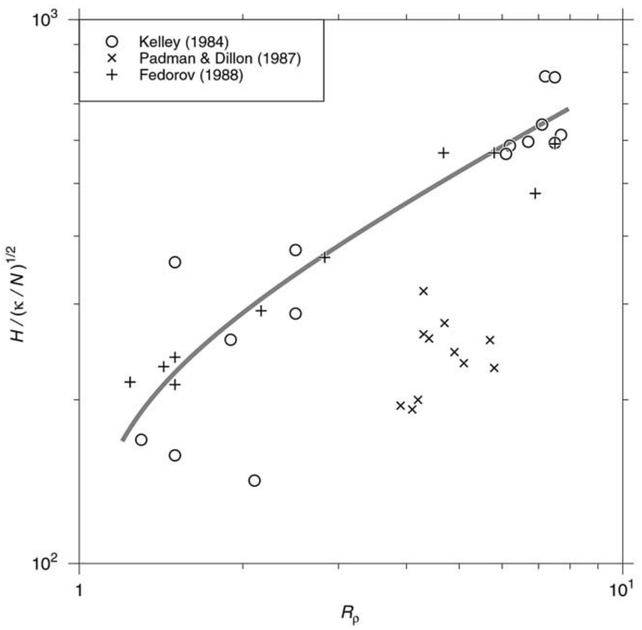

Fig. 5. Density–ratio dependence of nondimensional layer thickness $H/(\kappa/\bar{N})^{1/2}$ in ocean/lake DL staircases. The curve is the prediction of a model of convectively-sheared DL interfaces (Kelley, 1984, 1988). Data sources: ○ for Kelley (1984); + for Fedorov (1988), and × for Padman and Dillon, 1987.

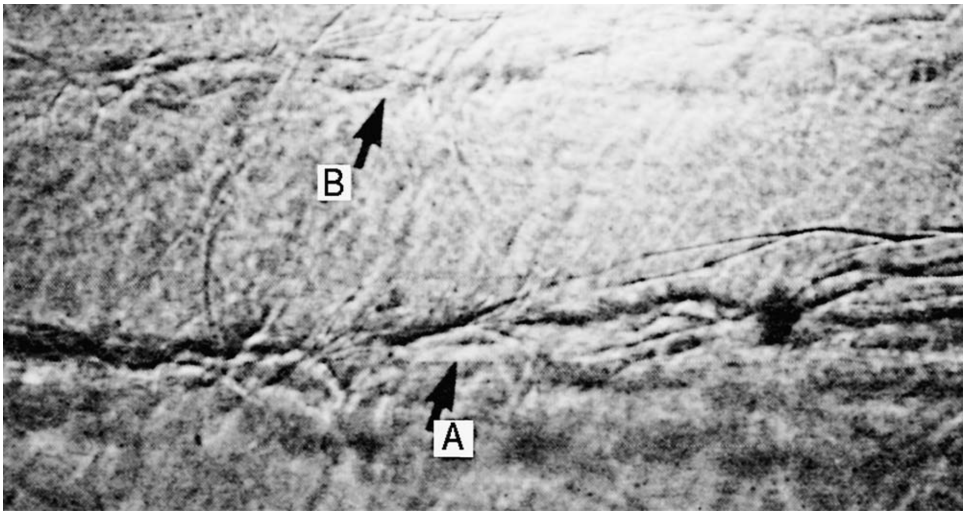

Fig. 6. Side view of DL created by heating a salt gradient from below, after Fig. 3 of Fernando (1987). Two interfaces are revealed by the striations: 'A' is the first quasi-stationary density interface, while 'B' is a secondary interface formed above it.

which illustrates that perturbations of the form $\exp(\lambda t + imz)$ added to a linear profile will tend to grow, rather than decay as they would with down-gradient density transport. If K_ρ were a constant, this would yield an ultraviolet catastrophe, since the timescale $(m^2K_\rho)^{-1}$ for perturbation growth would increase without limit as wavelength approached zero. However, DDC motions cannot extend to arbitrarily small lengthscales, so K_ρ must decrease in small scales, and this could provide a method of scale selection. So too could variations of K_ρ with stratification. However, given our present lack of understanding of how to express DDC fluxes in terms of diffusivities, it does not seem useful to pursue these ideas much further in a diffusivity framework. More useful would be laboratory work and, eventually, direct numerical simulations (Kelley, 2001, Appendix). These could be analogous to the related problem of layers formed by flux divergences associated with stratification-dependent turbulent fluxes (Phillips, 1972; Posmentier, 1977; Ruddick, McDougall, & Turner., 1989). Indeed, interactions between DDC and turbulent mechanisms of finestructure generation seem particularly worthy of study.

3.3.2. Applied-flux mechanism

Turner and Stommel (1964) were the first to create a DL staircase in the laboratory, by heating a stable salinity gradient from below. They found that the initially smooth salinity gradient broke down into a series of convecting layers separated by relatively sharp interfaces. Fig. 6, from the followup study of Fernando

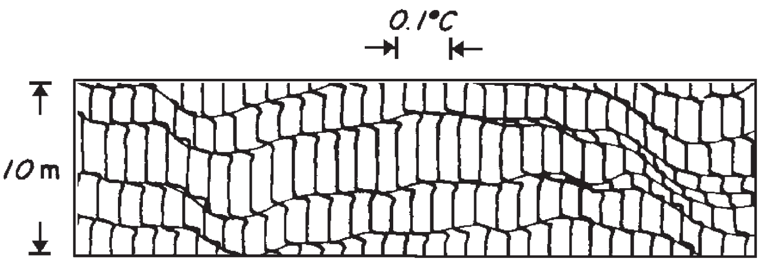

Fig. 7. Temperature profiles near 210 m depth, on 5 December 1970 at Ice Island T3 in the Arctic, after Neshyba et al. (1971). The total time interval is about 2 hours, with about 4 minutes between the offset profiles. Note the appearance of 'split' layers near the end of this interval.

(1987), reveals the complicated nature of such systems. The focus in many studies has been on simple aspects of the response, and particularly on the thickness of layers that are created. For example, Turner, 1968a) proposed that the bottom convective layer grows while maintaining marginal stability at the interface between the mixed layer and the stratified layer aloft, i.e. that the convection was non-penetrative. As this bottom-most mixed layer grows, a thermal boundary layer above it also grows, and Turner, 1968a) argued that the time of creation of a second convecting layer is controlled by the evolution of the thermal boundary layer to a convectively critical state. Denoting by Ra_c the Rayleigh number of the thermal boundary layer at this time, Turner, 1968a) suggested that the thickness of the bottom layer at the onset of the second layer is

$$H_T = \left(\frac{1}{4}Ra_c\right)^{\frac{1}{4}}\left(\frac{\nu J_B^3}{\kappa^2 N_{S0}^8}\right)^{\frac{1}{4}} \tag{6}$$

where N_{S0} is the buoyancy frequency of the initial salt stratification and J_B is the buoyancy flux applied at the bottom. Huppert and Linden (1979) have extended the analysis to describe the formation of a series of layers.

One disparity in the argument put forward by Turner, 1968a) is that the critical Ra_c was an order of magnitude larger than the theoretical estimate. Fernando (1987) provided an explanation, arguing that the convection is penetrative, rather than non-penetrative as had been assumed by Turner, 1968a). Using scaling arguments, Fernando (1987) suggested that the depth of convecting layers in a thermohaline staircase is determined by a balance between the vertical inertia and buoyancy forces of the mixed layer eddies.

From the point of view of application to the ocean, a weakness of the Turner, 1968a) analysis is that the "initial" state may involve thermal as well as haline stratification, and is unknown in any case. Furthermore, there are very few situations in which an applied buoyancy flux has been measured (geothermally-heated regions being the exception). To avoid such problems, Fernando, 1989b) extended the analysis, proposing that the thickness of convecting layers is given by

$$H_F = c_1\left(\frac{J}{N_S^3}\right)^{\frac{1}{2}} \cdot \frac{(1-c_2R_\rho)^{\frac{1}{2}}}{(1-R_\rho^{-1})^{\frac{3}{4}}}, \tag{7}$$

where c_1 and c_2 are constants to be determined empirically, and N_S is the haline component of stratification at the time of observation (not at some unobserved initial time), R_ρ is the instantaneous density ratio, and J is the instantaneous buoyancy flux. Molemaker and Dijkstra (1997) carried out a comparative direct numerical simulation study of the mechanisms proposed by Turner and Fernando and found support for the latter. However, their simulation was two dimensional, so further studies are needed to confirm their inferences.

While the Fernando (1989b) analysis of layer thickness circumvents several difficulties with the earlier theory, it is still difficult to apply to the ocean since the buoyancy flux J is not easy to measure (Gregg, 1987; Gargett, 1997). Furthermore, testing (7) using J calculated from layer- based flux laws is problematic since ΔS and ΔT depend on the observed H and the observed stratification parameters, so the two sides of the equation are linked. The best possibility of testing the two theories cited above may be in geothermally heated domains. These should afford measurements of J that can be trusted to within a factor of 5 (e.g. see Table 1 of Zolotarev, Sochel'Nikov and Malovitskiy (1979), for the Black Sea), so that if H can be measured to within $\sqrt{5}$, hypotheses based on the theories could be tested.

More remains to be done in the laboratory as well. Kerpel, Tanny and Tsinober (1991) addressed the case in which the heat flux arises from a boundary of constant temperature, as opposed to constant flux. Several other cases come to mind, ranging widely in difficulty of execution in the laboratory, e.g. supplying both salt and heat at a boundary, seeking an equilibrium solution by removing from one boundary the heat/salt supplied at another, etc. Issues of the timescale of staircase creation and evolution are also particularly relevant for application to unsteady problems in the ocean, especially in rapidly developing interleaving systems.

3.3.3. Modified-intrusion mechanism

Recently a new idea has been presented for the formation of staircases, namely that they are modified intrusions (Merryfield, 2000). This model, to date applied only to the SF case, produces reasonable predictions of measurable quantities such as layer thickness and the critical range of R_ρ for staircase formation. As with many intrusion models, it is based on a diffusivity formulation of DDC fluxes, and as such it shares a dependence on parameters that are essentially unknown since we lack a theory for the DDC fluxes arising out of a given smooth stratification. Application to the DL case, e.g. for application to the Arctic, and including special Arctic features such as along-ridge flows of warm/salty watermasses, would be most useful. So would more detailed studies of the formation of initial intrusions, e.g. accounting for baroclinicity in ocean fronts (Kuzmina & Rodionov, 1992; May & Kelley, 1997, 2001, 2002), and the possibility of interleaving being set up by sloping boundaries even in the absence of initial lateral variation in water properties (Linden & Weber, 1977). Field work is needed to provide constraints on theories, and an ideal location for this may be the Black Sea, which is subject to both interleaving and staircase modes, perhaps linked (recall Section 2.3.3 above).

3.4. Staircase evolution

The staircase-formation analysis of Turner (1968a) was followed up by Huppert and Linden (1979), who added a set of ordinary differential equations representing the evolution of (H,S,T) for a suite of layers coupled by layer-based flux formulas. Migration of interior interfaces was ignored, as was layer splitting, but layer merging was modeled crudely. They found reasonable agreement with laboratory observations, suggesting that such a model (perhaps with flux laws adjusted as discussed in Section 3.1 above) could be useful in oceanographic applications. Still, this leaves out several oceanographic aspects that we might expect to be important.

Firstly, consider the lateral environment. Analyses of bottom-heated experiments typically regard the situation as one-dimensional, in the sense that lateral variations may be ignored in a statistical sense, even if they are present in instantaneous snapshots (Fig. 6). Of course, this viewpoint is entirely at odds with the Merryfield (2000) model of staircase creation from interleaving fields. There have been no field tests of this model. As for lateral variation in general, only one observational program, the C-SALT study of SF staircases (Schmitt, Perkins, Boyd, & Stalcup, 1987), has sampled a thermohaline staircase at a wide range of lateral scales. This revealed the expected intrusions near the edge of the staircase zone, but also isolated intrusions within the staircase interior. High-resolution sampling with towed thermistor chains (Marmorino, 1989, 1991) reveals that the lateral features have lengthscales of 1 km. Similarly, Padman and Dillon (1988) found that station spacings of less than 1 km were required to track individual layers in a DL staircase in the Canadian Basin of the Arctic. This is in line with the observations of Neshyba, Neal and Denner (1971), an excerpt of whose observations is presented as Fig. 7 here. Note in particular the occurrence of split layers in this diagram, whose variation in this graphical representation is probably the result of advection of horizontal variability, as opposed to temporal development (Kelley, 1987b). Perhaps the ideas of Stamp, Hughes, Griffiths and Nokes (1998), regarding solitary varicose waves within interfaces, are relevant to the dynamics of split layers as well. For now, we must admit that the dynamics of these split layers, and of lateral variability in general, are poorly understood. If these split layers are important to maintaining a critical layer thickness, as a counter to layer merging by convergent buoyancy fluxes (Kelley, 1988), then further study is definitely warranted. This is because they may hold the key to a usable parameterization of layer thickness, and thus vertical fluxes, in terms of large-scale properties (recall section 3.2 above).

Along with such external effects, there are possibilities for staircase evolution that arise from the properties of seawater. One is interface migration, which may occur due to the nonlinearity of the equation of state of seawater (McDougall, 1981). Preferential interfacial migration may cause substantial flux contributions (Kelley, 1987b; Rudels, 1991), which can be estimated to the first order by considering the entrainment velocities at the interface caused by each layer when the opposite layer is not entraining (Fernando, 1990). The entrainment is effective mainly at low interfacial Richardson numbers; if $R_i > 10$ then entrainment essentially ceases and the interface becomes either stationary or detraining (Hull, Bushnell, Sempsrote, & Pena, 1989; Zangrando & Fernando, 1991).

Another general category of great interest is the effect on double-diffusion of classical oceanographic effects, such as planetary rotation, turbulent mixing, and background shear. For technical reasons, these issues are difficult to work with in the laboratory. Still, each merits examination.

The effect of planetary rotation is to quench convection by preventing vortex stretching. However, Kelley (1987a) argues that planetary rotation is unlikely to inhibit DL convection given the small scales of typical oceanic staircases. This raises the question as to whether the scales of observed convecting layers are small because of the inhibition of larger-scale convection. The answer, which may affect ideas on layer scaling and thus on diffusivity formulation, may come from constructing better models of rotating double-diffusive convection.

The effect of turbulence on SF and DL interfaces has been explored in the laboratory by Linden (1971) and Crapper (1976) respectively, but it is not clear how directly these relate to the ocean, in which mixing may have a different character (e.g. different length and time scales) than in these laboratory situations. A related topic, and one which needs more study, is the relationship between differential mixing and DDC. Differential mixing (DM, henceforth) results from the differing diffusivities for heat and salt acting on water parcels momentarily put into contact, e.g. by incompletely mixed breaking waves (Turner, 1968b; Altman & Gargett, 1987; Ruddick, 1997). Two-dimensional direct numerical simulations suggest that DM may be significant at low turbulent mixing rates (Merryfield, Holloway, & Gargett, 1998), i.e. in the same realm in which DDC seems most prevalent. This raises questions about how DM and DDC might interact in the staircase mode. There is already good reason to study DM since it has recently been proposed as a cause of intrusions (Merryfield, 2002), something that has long been thought to have a double-diffusive cause.

The effect of shear has been alluded to in Section 3.1 above. There are indications that the SF mode is inhibited greatly by ocean shears, but that the DL mode is not. But is this just because the DL observations are from areas of low shear? Perhaps not, since it has been proposed that background shear inhibits SF fluxes (Linden, 1974b; Kunze, 1994) but that shear may enhance DL momentum transport (Padman, 1994). Whether this proposed difference in SF and DL interface dynamics is verified or not, it illustrates of the danger in relying on a direct analogy between the two modes of DDC.

4. Summary

In many ways, the DL case seems simpler than the SF case. For example, consider fluxes. For decades, oceanographers estimated SF fluxes in the ocean using layer-based flux laws based on laboratory experiments, but the C-SALT program (Gregg & Sanford, 1987; Lueck, 1987) revealed this to be in error by well over an order of magnitude. This may be the result of SF disruption by internal-wave shear (Linden, 1974b; Kunze, 1994). In the DL case, whether by virtue of weaker shears or differences in the dynamical nature of DL interfaces, empirical evidence suggests that it is entirely reasonable to extrapolate from the laboratory using layer-based flux laws. It remains to be seen whether this holds in general, but at least it provides a hypothesis worth testing, and a tentative basis for further computations, while awaiting further work on how DL fluxes are affected by turbulence, shear and other oceanic effects.

If we may express vertical DL fluxes of salt and heat in terms of layer height H and the steps in S and T from layer to layer (without regard to internal-wave shear or turbulence levels, say) then the goal of parameterizing DL fluxes in terms of the large-scale stratification can be met if only we can parameterize H in such terms. Such a parameterization has been proposed for the DL case (Kelley, 1984) but the evidence for it is divided enough to demand more empirical tests. The proposed mechanism behind the parameterization involves the splitting of layers when a critical thickness is achieved (Kelley, 1988). Because of this, and because layer splitting may provide a large component of lateral variability, this general topic seems greatly deserving of further study.

Any parameterization scheme based on an equilibrium assumption will be irrelevant if the ocean changes too quickly for staircases to be formed or to evolve to equilibrium. For this reason some of the most important questions about DDC in the ocean center on the mechanism of staircase formation. There may be several modes. The bottom-heating mode, easily set up in the laboratory, may also occur in geophysical domains such as geophysically-heat lakes, and oceanic trenches and brine lakes. However, several other possibilities exist. These include instabilities associated with up-gradient DDC density transports and a newly-proposed mode of formation from interleaving systems (Merryfield, 2002). Field programs may be needed to select and differentiate between such alternatives. The Black Sea and the Arctic are prime candidates for field sites. Each displays prominent signatures of DL staircases as well as intrusions. If the SF case provides any guide, an intensive field program using a variety of sampling protocols may be needed. The C-SALT field program re-invigorated research on the SF mode, motivating studies of the effects of shear on interfaces, of intrusions within staircases, of the linkage between vertical and lateral effects, etc. An intensive field program for the DL case might be as productive as C-SALT proved to be for the SF case, and go a long way to establishing whether DL is important in the ocean, globally or locally.

References

Altman, D. B., & Gargett, A. E. (1987). Differential property transport due to incomplete mixing in a stratified fluid. In Proceedings of the third international symposium on stratified flows, Pasadena, CA. Cal. Inst. of Technology.

Bianchi, A. A., Piola, A. R., & Collino, G. J. (2002). Evidence of double diffusion in the Brazil–Malvinas confluence. *Deep-Sea Research I*, *49*, 45–56.

Boldrin, A., & Rabitti, S. (1990). Hydrography of the brines in the Bannock and Tyro anoxic basins (eastern Mediterranean). *Marine Chemistry*, *31*, 21–33.

Brandt, A., & Fernando, H. J. S. (1995). *Double-diffusive convection*. American Geophysical Union Monograph No. *94*.

Buesseler, K. O., & Livingston, H. D. (1997). Time-series profiles of 134Cs, 137Cs and 90Sr in the Black Sea. In E. Özsoy, & A. Mikaelyan (Eds.), *Sensitivity to change: Black Sea, Baltic Sea and North Sea*. *NATO ASI Series*, Kluwer Academic Publishers.

Buesseler, K. O., Livingston, H. D., & Casso, S. (1991). Mixing between oxic and anoxic waters of the Black Sea as traced by Chernobyl cesium isotopes. *Deep-Sea Research*, *38*, S725–S745.

Carmack, E. C., Macdonald, R. W., Perkin, R. G., McLaughlin, F. A., & Pearson, R. J. (1995a). Evidence of warming of Atlantic water in the southern Canadian Basin of the Arctic Ocean: results from the Larsen-93 expedition. *Geophysical Research Letters*, *22*, 1061–1064.

Carmack, E. C., Aagaard, K., Swift, J. H., Perkin, R. G., McLauglin, F. A., Macdonald, R. W., Jones, E. P. (1995b). Thermohaline transitions. In J. Imberger (Ed.), *IUTAM symposium on Physical Limnology*. International Union of Theoretical Applied Mechanics.

Carmack, E. C., Aagaard, K., Swift, J. H., MacDonald, R. W., McLaughlin, F. A., Jones, E. P., Perkin, R. G., Smith, J. N., Ellis, K. M., & Killius, L. R. (1997). Changes in temperature and tracer distributions within the Arctic Ocean: results from the 1994 Arctic Ocean section. *Deep-Sea Research II*, *44*, 1487–1502.

Castaing, B., Gunaratne, G., Heslot, F., Kadanoff, L., Libchaber, A., Thomae, S., Wu, X.-Z., Zaleski, S., & Zanetti, G. (1989). Scaling of hard thermal turbulence in Rayleigh-Benard convection. *Journal of Fluid Mechanics*, *204*, 1–30.

Codispoti, L. A., Friederich, G. E., Murray, J. W., & Sakamoto, C. (1991). Chemical variability in the Black Sea: Implications of data obtained with a continuous vertical profiling system that penetrated the oxic/anoxic interface. *Deep-Sea Research*, *38*, S691–710.

Crapper, P. F. (1975). Measurements across a diffusive interface. *Deep-Sea Research*, *22*, 537–545.

Crapper, P. F. (1976). Flux of heat and salt across a diffusive interface in the presence of grid-generated turbulence. *International Journal of Heat Mass Transfer*, *19*, 1371–1378.

Federov, K. N. (1970). On the step-like structure of temperature inversions in the ocean. *Atmospheric and Oceanic Physics*, *6*, 1178–1188.

Fedorov, K. N. (1988). Layer thickness and effective diffusivities in 'diffusive' thermohaline convection in the ocean. In J. C. J. Nihoul, & B. M. Jamart (Eds.), *Small-scale turbulence and mixing in the ocean* (pp. 471–479). Elsevier.

Fernando, H. J. S. (1987). The formation of layered structure when a stable salinity gradient is heated from below. *Journal of Fluid Mechanics*, *182*(425–442), 525–541.

Fernando, H. J. S. (1989a). Buoyancy transfer across a diffusive interface. *Journal of Fluid Mechanics*, *209*, 1–34.

Fernando, H. J. S. (1989b). Oceanographic implications of laboratory experiments on diffusive interfaces. *Journal of Physical Oceanography*, *19*, 1707–1715.

Fernando, H. J. S. (1990). Comments on 'interfacial migration in thermohaline staircases'. *Journal of Physical Oceanography*, *20*, 1994–1996.

Fernando, H. J. S., & Brandt, A. (1994). Recent advances in double-diffusive convection. *Applied Mechanics Review*, *47*, c1–c7.

Fernando, H. J. S., & Ching, C. Y. (1991). An experimental study on thermohaline staircases. In R. W. Schmitt (Ed.), *Double diffusion in oceanography, proceedings of a meeting, 1989*, pp. 141–149. Technical Report, Woods Hole Oceanographic Institution.

Foster, T. D., & Carmack, E. C. (1976). Temperature and salinity structure in the Weddell Sea. *Journal of Physical Oceanography*, *6*, 36–44.

Galbraith, P. S., & Kelley, D. E. (1996). Identifying overturns in CTD profiles. *Journal of Atmospheric and Ocean Technology*, *13*, 688–702.

Gargett, A. E. (1997). 'Theories' and techniques for observing turbulence in the ocean euphotic zone. In C. Marrasé, E. Saiz, & J. Redondo (Eds.), *Lecture notes on plankton and turbulence*, 61(Suppl), 25–45. Scientia Marina.

Garrett, C. (1990). The role of secondary circulation in boundary mixing. *Journal of Geophysical Research*, *95*, 3181–3188.

Grasshoff, K. (1975). The hydrochemistry of landlocked basins and fjords. In J. P. Riley, & G. Skirrow (Eds.), *Chemical Oceanograph*, 647. Academic Press.

Gregg, M. C. (1987). Diapycnal mixing in the thermocline: a review. *Journal of Geophysical Research*, *92*, 5249–5286.

Gregg, M. C., & Sanford, T. B. (1987). Shear and turbulence in thermohaline staircases. *Deep-Sea Research*, *34*, 1689–1696.

Hoare, R. A. (1966). Problem of heat transfer in Lake Vanda, a density stratified Antarctic lake. *Nature, London*, *10*, 787–789.

Hoare, R. A. (1968). Thermohaline convection in Lake Vanda, Antarctica. *Journal of Geophysical Research*, *73*, 607–612.

Hull, J. R., Bushnell, D. L., Sempsrote, D. G., & Pena, A. (1989). Ammonium sulfate solar pond: Observations from small scale experiments. *Solar Energy*, *1*, 57–64.

Huppert, H. E. (1972). On the stability of a series of double diffusive layers. *Deep-Sea Research*, *18*, 1005–1022.

Huppert, H. E., & Linden, P. (1979). On heating a stable salinity gradient from below. *Journal of Fluid Mechanics*, *95*, 431–464.

Huppert, H. E., & Turner, J. S. (1972). Double-diffusive convection and its implications for the temperature and salinity structure of the ocean and Lake Vanda. *Journal of Physical Oceanography*, *2*, 456–461.

Huppert, H. E., & Turner, J. S. (1981). Double-diffusive convection. *Journal of Fluid Mechanics*, *106*, 299–329.

Kelley, D. E. (1984). Effective diffusivities within oceanic thermohaline staircases. *Journal of Geophysical Research*, *89*, 10484–10488.

Kelley, D. E. (1987a). The influence of planetary rotation on oceanic double-diffusive fluxes. *Journal of Marine Research*, *45*, 829–841.

Kelley, D. E. (1987b). Interfacial migration in thermohaline staircases. *Journal of Physical Oceanography*, *17*, 1633–1639.

Kelley, D. E. (1988). Explaining effective diffusivities within diffusive oceanic staircases. In J. C. J. Nihoul, & B. M. Jamart (Eds.), *Small-scale turbulence and mixing in the ocean* (pp. 481–502). Amsterdam: Elsevier.

Kelley, D. E. (1990). Fluxes through diffusive staircases: a new formulation. *Journal of Geophysical Research*, *95*, 3365–3371.

Kelley, D. E. (2001). Six questions about double-diffusive convection. In P. Müller, & C. Garrett (Eds.), *From stirring to mixing in a stratified ocean: 12th 'Aha Huliko' a Hawaiian Winter Workshop*. Univ. of Hawaii.

Kelley, D. E., & Van Scoy, K. A. (1999). A basin-wide estimate of vertical mixing in the upper pycnocline: spreading of bomb tritium in the North Pacific Ocean. *Journal of Physical Oceanography*, *29*, 1759–1771.

Kerpel, J., Tanny, J., & Tsinober, A. (1991). On a stable solute gradient heated from below with prescribed temperature. *Journal of Fluid Mechanics*, *223*, 83–91.

Kerr, R. M. (1996). Rayleigh number scaling in numerical convection. *Journal of Fluid Mechanics*, *310*, 139–179.

Krishnamurti, R. (1970). On the transition to turbulent convection. Part 2. The transition to time-dependent flow. *Journal of Fluid Mechanics*, *42*, 309–320.

Kunze, E. (1987). Limits on growing, finite-length salt fingers: a Richardson number constraint. *Journal of Marine Research*, *45*, 533–556.

Kunze, E. (1994). A proposed flux constraint for salt fingers in shear. *Journal of Marine Research*, *52*, 999–1016.

Kuzmina, N. P., & Rodionov, V. B. (1992). Influence of baroclinicity on formation of thermohaline intrusions in ocean frontal zones. *Izvestiya, Atmospheric and Oceanic Physics*, *28*, 804–810.

Lazier, J., & Sandstrom, H. (1978). Migrating thermal structure in a freshwater thermocline. *Journal of Physical Oceanography*, *8*, 1070–1079.

Lazier, J. R. N. (1973). Temporal changes in some fresh water temperature structures. *Journal of Physical Oceanography*, *3*, 226–229.

Ledwell, J. R., Watson, A. J., & Law, C. S. (1993). Evidence for slow mixing across the pycnocline from an open-ocean tracer-release experiment. *Nature, London*, *364*, 701–703.

Levitus, S., & Boyer, T. P. (1994). World ocean atlas 1994: temperature. Technical report, US Dept. of Commerce, NOAA Atlas NESDIS 4, Washington, D.C.

Levitus, S., Burgett, R., Boyer, T. P. (1994). World ocean atlas 1994: salinity. Technical report, US Dept. of Commerce, NOAA Atlas NESDIS 3, Washington, D.C.

Linden, P. F. (1971). Salt fingers in the presence of grid generated turbulence. *Journal of Fluid Mechanics*, *49*, 611–624.

Linden, P. F. (1974a). A note on the transport across a diffusive interface. *Deep-Sea Research*, *21*, 287.

Linden, P. F. (1974b). Salt fingers in a steady shear flow. *Geophysics & Fluid Dynamics*, *6*, 1–27.

Linden, P. F., & Shirtcliffe, T. G. L. (1978). The diffusive interface in double-diffusive convection. *Journal of Fluid Mechanics*, *87*, 417–432.

Linden, P. F., & Weber, J. E. (1977). The formation of layers in a double-diffusive system with a sloping boundary. *Journal of Fluid Mechanics*, *81*, 757–773.

Lueck, R. G. (1987). Microstructure measurements in a thermohaline staircase. *Deep-Sea Research*, *34*, 1677–1688.

Marmorino, G. O. (1974). *Equilibrium heat and salt transport through a diffusive, thermohaline interface*. Master's thesis, Oregon State University.

Marmorino, G. O. (1989). Substructure of oceanic salt finger interfaces. *Journal of Geophysical Research*, *94*, 4891–4904.

Marmorino, G. O. (1991). Intrusions and diffusive interfaces in a salt finger staircase. *Deep-Sea Research*, *38*, 1431–1454.

Marmorino, G. O., & Caldwell, D. R. (1976). Heat and salt transport through a diffusive thermohaline interface. *Deep-Sea Research*, *23*, 59–67.

May, B. D., & Kelley, D. E. (1997). Effect of baroclinicity on double-diffusive interleaving. *Journal of Physical Oceanography*, *27*, 1997–2008.

May, B. D., & Kelley, D. E. (2001). Dynamics of thermohaline intrusions in the Arctic Ocean. *Journal of Geophysical Research*, *106*(16), 16.

May, B. D., & Kelley, D. E. (2002). Contrasting the interleaving in two baroclinic ocean fronts. *Dynamics of Atmospheres and Oceans*, *36*, 23–42.

McDougall, T. J. (1981). Double-diffusive convection with a non-linear equation state: Part II: laboratory experiments and their interpretation. *Progress in Oceanography*, *10*, 91–121.

McDougall, T. J. (1983). Greenland Sea Bottom Water formation: a balance between advection and double diffusion. *Deep-Sea Research*, *30*, 1109–1117.

McDougall, T. J. (1984). Fluid dynamical implications for massive sulphide deposits of hot saline fluid flowing into a submarine depression from below. *Deep-Sea Research*, *31*, 145–170.

McDougall, T. J., & Taylor, J. R. (1984). Flux measurements across a finger interface at low values of the stability ratio. *Journal of Marine Research*, *42*, 1–14.

MEDRIFF Consortium (1995). Three brine lakes discovered in the seafloor of the eastern Mediterranean. *EOS, Transactions of the American Geophysical Union*, *76*, 313–318.

Melling, H., Lake, R. A., Topham, D. R., & Fissel, D. B. (1984). Oceanic thermal structure in the western Canadian Arctic. *Continental Shelf Research*, *3*, 223–258.

Merryfield, W. J. (2000). Origin of thermohaline staircases. *Journal of Physical Oceanography*, *30*, 1046–1068.

Merryfield, W. J. (2002). Intrusions in double-diffusively stable Arctic waters: evidence for differential mixing? *Journal of Physical Oceanography*, *32*, 1452–1459.

Merryfield, W. J., Holloway, G., & Gargett, A. E. (1998). Differential vertical transport of heat and salt by weak stratified turbulence. *Geophysical Research Letters*, *25*, 2773–2776.

Merryfield, W. J., Holloway, G., & Gargett, A. E. (1999). A global ocean model with double-diffusive mixing. *Journal of Physical Oceanography*, *29*, 1124–1142.

Molemaker, M. J., & Dijkstra, H. (1997). The formation and evolution of a diffusive interface. *Journal of Fluid Mechanics*, *331*, 199–229.

Muench, R. D., Fernando, H. J. S., & Stegen, G. R. (1990). Temperature and salinity staircases in the northwestern Weddell sea. *Journal of Physical Oceanography*, *20*, 295–306.

Muench, R. D., Rudels, B., Bjork, G., & Schauer, U. (1997). Basin scale double-diffusive layering in the Arctic Ocean. In Marine Turbulence Revisited, 29th International Liege Colloquium on Ocean Hydrodynamics, Liege May 5–9, 1997. Elsevier.

Munk, W. H. (1966). Abyssal recipes. *Deep-Sea Research*, *13*, 707–730.

Munk, W. H., & Wunsch, C. (1998). Abyssal recipes II: energetics of tidal and wind mixing. *Deep-Sea Research*, *45*, 1977–2010.

Murray, J. W., Top, Z., & Özsoy, E. (1991). Hydrographic properties and ventilation of the Black Sea. *Deep-Sea Research*, *38*(2), S663–S689.

Narusawa, V. (1986). Structure of diffusive interface of double diffusive convection. ASME paper 86-HT-3.

Neal, V. T., Neshyba, S., & Denner, W. (1969). Thermal stratification in the Arctic Ocean. *Science*, *166*, 373–374.

Neshyba, S., Neal, V., & Denner, W. (1971). Temperature and conductivity measurements under ice island T-3. *Journal of Geophysical Research*, *76*, 8107–8120.

Newell, T. A. (1984). Characteristics of double-diffusive interface at high density-stability ratios. *Journal of Fluid Mechanics*, *149*, 385–401.

Newman, F. C. (1976). Temperature steps in Lake Kivu: A bottom heated saline lake. *Journal of Physical Oceanography*, *6*, 157–163.

Osborn, T. R. (1973). Temperature microstructure in Powell Lake. *Journal of Physical Oceanography*, *3*, 302–307.

Östlund, H. G. (1974). Expedition Odysseus 65: Radiocarbon age of Black Sea water. In E. Degens and D. Ross (Eds.), *The Black Sea—geology, chemistry and biology*. The American Association of Petroleum Geologists, Memoir No. 20.

Östlund, H. G., & Dyrssen, D. (1986). Renewal rates of the Black Sea deep water. In *The chemical and physical oceanography of the Black Sea*, Univ. of Göteborg, Rep. on the Chemistry of the Sea XXXIII, Presented in the Meeting on the Chemical and Physical Oceanography of the Black Sea, Göteborg, Sweden, June 1986. University of Göteborg.

Özsoy, E., & Besiktepe, S. (1995). Sources of double diffusive convection and impacts on mixing in the Black Sea. In A. Brandt, & H. J. S. Fernando (Eds.), Double-diffusive convection (pp. 261–274). American Geophysical Union, Monograph No. 94.

Özsoy, E., & Ünlüata, U. (1997). Oceanography of the Black Sea: A review of some recent results. *Earth Sciences Review*, *42*(4), 231–272.

Özsoy, E., & Ünlüata, U. (1998). The Black Sea. In A. R. Robinson, & K. Brink (Eds.), (pp. 889–914). *The global coastal ocean: Regional studies and synthes*, 11. John Wiley and Sons.

Özsoy, E., Top, Z., White, G., & Murray, J. W. (1991). Double diffusive intrusions, mixing and deep convective processes in the Black Sea. In E. Zdar, & J. Murray (Eds.), *Black Sea Oceanography* (pp. 17–42). *NATO ASI series C*, 351. Kluwer Academic Publishers.

Özsoy, E., Ünlüata, U., & Top, Z. (1993). The Mediterranean water evolution, material transport by double diffusive intrusions, and interior mixing in the Black Sea. *Progress in Oceanography*, *31*, 275–320.

Padman, L. (1994). Momentum fluxes through sheared oceanic thermohaline steps. *Journal of Geophysical Research*, *99*, 22491–22499.

Padman, L., & Dillon, T. M. (1987). Vertical fluxes through the Beaufort Sea thermohaline staircase. *Journal of Geophysical Research*, *92*, 799–806.

Padman, L., & Dillon, T. M. (1988). On the horizontal extent of the Canada Basin thermohaline steps. *Journal of Physical Oceanography*, *18*, 1458–1462.

Padman, L., & Dillon, T. M. (1989). Thermal microstructure and internal waves in the Canada Basin diffusive staircase. *Deep-Sea Research*, *36*, 31–542.

Perkin, R. G., & Lewis, E. L. (1984). Mixing in the West Spitsbergen current. *Journal of Physical Oceanography*, *14*, 1315–1325.

Phillips, O. M. (1972). Turbulence in a strongly stratified fluid—is it unstable? *Deep-Sea Research*, *19*, 79–81.

Posmentier, E. S. (1977). The generation of salinity finestructure by vertical diffusion. *Journal of Physical Oceanography*, *7*, 298–299.

Robertson, R., Padman, L., & Levine, M. (1995). Fine structure, microstructure, and vertical mixing processes in the upper ocean in the western Weddell Sea. *Journal of Geophysical Research*, *100*, 18517–18535.

Ruddick, B. (1997). Differential fluxes of heat and salt: implications for circulation and ecosystem modelling. *Oceanography*, *10*(3), 122–127.

Ruddick, B. R., McDougall, T. J., & Turner, J. S. (1989). The formation of layers in a uniformly stirred density gradient. *Deep-Sea Research*, *36*, 597–609.

Rudels, B. (1991). The diffusive interface at low stability: the importance of non-linearity and turbulent entrainment. *Tellus A*, *43*, 153–167.

Schmitt, R. W. (1979). Flux measurements on salt fingers at an interface. *Journal of Marine Research*, *37*, 419–436.

Schmitt, R. W. (1994). Double diffusion in oceanography. *Annual Review of Fluid Mechanics*, *26*, 255–285.

Schmitt, R. W., Perkins, H., Boyd, J. D., & Stalcup, M. C. (1987). C-SALT: an investigation of the thermohaline staircase in the western tropical north atlantic. *Deep-Sea Research*, *34*, 1655–1665.

Shirtcliffe, T. (1973). Transport and profile measurements of the diffusive interface in double diffusive convection with similar diffusivities. *Journal of Fluid Mechanics*, *57*, 27–43.

Shirtcliffe, T. G. L., & Calhaem, I. M. (1968). Measurements of temperature and electrical conductivity in Lake Vanda, Victoria Land, Antarctica. *New Zealand Journal of Geological Geophysics*, *11*, 976–981.

Sommeria, J. (1999). The elusive 'ultimate state' of thermal convection. *Nature, London*, *398*, 294–295.

Stamp, A. P., Hughes, G., Griffiths, R. W., & Nokes, R. I. (1998). Coherent circulation due to coupling of double-diffusive convection and interfacial waves. *Journal of Fluid Mechanics*, *372*, 231–271.

Staneva, J. E., & Stanev, E. V. (1997). Cold intermediate water formation in the Black Sea: analysis on numerical model simulations.

In E. Özsoy, & A. Mikaelyan (Eds.), *Sensitivity to change: Black Sea, Baltic Sea and North Sea*. *NATO ASI Series*, Kluwer Academic Publishers.

Stanley, D. J., & Blanpied, C. (1980). Late quaternary water exchange between the eastern Mediterranean and the Black Sea. *Nature, London*, *285*, 537–541.

Stenni, B., & Longinelli, A. (1990). Stable isotope study of water, gypsum and carbonate samples from the Bannock and Tyro Basins, Eastern Mediterranean. *Marine Chemistry*, *31*, 123–135.

Stern, M. E. (1960). The 'salt-fountain' and thermohaline convection. *Tellus*, *2*, 172–175.

Stommel, H., Arons, A. B., & Blanchard, D. (1956). An oceanographical curiosity: the perpetual salt fountain. *Deep-Sea Research*, *3*, 152–153.

Swallow, J., & Crease, J. (1965). Hot salty water at the bottom of the Red Sea. *Nature, London*, *205*, 165–166.

Tait, R. I., & Howe, M. R. (1968). Some observations of thermohaline stratification in the deep ocean. *Deep-Sea Research*, *15*, 275–280.

Takao, S., & Narusawa, V. (1980). An experimental study of heat and mass transfer across a diffusive interface. *International Journal of Heat Mass Transactions*, *23*, 1283–1285.

Tanny, J., & Tsinober, A. (1988). The dynamics and structure of double-diffusive layers in sidewall-heating experiments. *Journal of Fluid Mechanics*, *196*, 135–156.

Taylor, J. (1988). The fluxes across a diffusive interface at low values of the density ratios. *Deep-Sea Research*, *35*, 555–567.

Top, Z., Östlund, H. G., Pope, L., & Grall, C. (1991). Helium with tritium in the Black Sea: a comparison with the 1975 observations. *Deep-Sea Research*, *38*(2), S747–S760.

Tsinober, A. B., Yahalom, Y., & Shlien, D. J. (1983). A point source of heat in a stable salinity gradient. *Journal of Fluid Mechanics*, *135*, 199–217.

Turner, J. S. (1965). The coupled turbulent transport of salt and heat across a sharp density interface. *International Journal of Heat Mass Transfer*, *8*, 759–767.

Turner, J. S. (1968a). The behavior of a stable salinity gradient heated from below. *Journal of Fluid Mechanics*, *33*, 183–200.

Turner, J. S. (1968b). The influence of molecular diffusivity on turbulent entrainment across a density interface. *Journal of Fluid Mechanics*, *33*, 639–656.

Turner, J. S. (1969). A physical interpretation of hot brine layers in the Red Sea. In E. T. Degens, & D. A. Ross (Eds.), *Hot brines and recent heavy metal deposits in the Red Sea* (pp. 164–172). Springer Pres.

Turner, J. S. (1974). Double-diffusive phenomena. *Annual Review of Fluid Mechanics*, *6*, 37–56.

Turner, J. S. (1978). Double-diffusive intrusions into a density gradient. *Journal of Geophysical Research*, *83*, 2887–2901.

Turner, J. S. (1985). Multicomponent convection. *Annual Review of Fluid Mechanics*, *17*, 11–43.

Turner, J. S., & Stommel, H. (1964). A new case of convection in the presence of combined vertical salinity and temperature gradients. *Proceedings of the National Academy of Science of the USA*, *52*, 49–53.

Venables, W. N., & Ripley, B. D. (1999). *Modern applied statistics with S-PLUS*. New York: Springer Press.

Voorhis, A. D., & Dorson, D. L. (1975). Thermal convection in the Atlantis II hot brine pool. *Deep-Sea Research*, *22*, 167–175.

Walsh, J., & Crane, R. G. (1992). A comparison of GCM simulations of Arctic climate. *Geophysical Research Letters*, *19*, 29–32.

Woods, A. W. (1991). Boundary-driven mixing. *Journal of Fluid Mechanics*, *226*, 625–654.

Zangrando, F., & Fernando, H. J. S. (1991). A predictive model for the migration of double-diffusive interfaces. *Journal of Solar Energy Engineering*, *113*, 59–65.

Zhang, J., & Schmitt, R. W. (2000). The impact of salt fingering on the thermohaline circulation under mixed boundary conditions. *Journal of Physical Oceanography*, *30*, 1223–1231.

Zhang, J., Schmitt, R. W., & Huang, R. X. (1998). Sensitivity of the GFDL modular ocean model to parameterization of double-diffusive processes. *Journal of Physical Oceanography*, *28*, 589–605.

Zolotarev, V. G., Sochel'Nikov, V. V., & Malovitskiy, Y. P. (1979). Results of heat-flow measurements in the Black and Mediterranean Sea basins. *Oceanology*, *19*(6), 701–705.