

Thermal Stability of the World Ocean Thermoclines*

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ABSTRACT

Because of the strong variation with temperature of the thermal expansion coefficient of seawater, both horizontal and vertical mixing that perturb the gradients produce changes of volume, usually a decrease, that shift mass relative to the earth's gravitational field resulting in significant changes of gravitational potential energy (GPE). For sufficiently large temperature gradients, these changes may serve as energy sources to supplement the tidal, internal wave, and other processes that mix the ocean waters and limit the gradient magnitudes. The hypothesis examined here is that for overall stable ocean stratifications of temperature and salinity, the GPE conversions must be positive with respect to local perturbations by diffusion, mixing, or other disturbances. Thus, for the long-term steady state of the oceans to exist, the GPE structure must be stable, including the nonlinear effects.

A dynamical description of the conversion process for GPE changes to kinetic energy for mixing has not yet been developed. The evidence for the significance of the process is based on observational data. Examples are given for several oceans to show the limiting effects of the nonlinear Equation of State properties on the main thermoclines.

1. Introduction

The density structures in the major ocean basins are created by circulation resulting from the thermal stratification produced by equatorial heating and polar cooling. The structures are modified further by salinity stratification resulting from evaporation and precipitation and by wind-driven circulation. These processes create gradients and fronts within the ocean interiors that are smoothed by turbulent mixing powered by internal waves, gravitational instabilities, double diffusion, and other sources.

The objective of the present paper is to describe how the nonlinear effect of contraction on mixing of seawater, referred to as "cabbeling," may determine major features of the ocean's temperature and salinity structures. The hypothesis is that the oceans are at a stable minimum gravitational potential energy (GPE) with respect to local perturbations. If vertical temperature gradients exceed a limiting value, the excess GPE released due to cabbeling will enhance local mixing and shorten the timescale necessary to eliminate the "instability."

The nonlinearity of the Equation of State (EOS-80)

results from the strong dependence of the thermal expansion coefficient on temperature. The nonlinearity is strongest for freshwater at 4°C and atmospheric pressure, and decreases with increasing temperature, salinity, and pressure. The first and second derivatives of specific volume are shown in Fig. 1. The first derivative with respect to temperature varies by an order of magnitude over the oceanic range of temperature, salinity, and pressure. The variation of specific volume with salinity is nearly linear. Thus, the major contributor to the nonlinear effects is the variation of the thermal expansion with temperature as can be seen from Fig. 1d. Figure 1 also compares the standard EOS-80 values with those computed from the more recent equations of Feistel and Hagen (1995). The differences are negligible for the present discussion. The EOS-80 has been described in detail by Fofonoff (1985).

A smaller, but important, effect is the variation of compressibility with temperature. Water is more compressible at low temperatures. Thus, a cold seawater layer overlying warmer water can become unstable with an increase in pressure. This effect was described for freshwater by Eckel (1949). A similar process was described by Ekman (1934) and Gill (1973) for the Antarctic thermocline.

The effect of cabbeling on ocean processes has been studied by many authors, starting with Witte (1902). Mamayev (1975) and Fofonoff (1995) have reviewed some of the earlier papers. More recent studies are by McDougall (1987). Effects of cabbeling have been in-

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incorporated in recent numerical models by McDougall and Dewar (1998). Earlier studies of the GPE changes of a water column were made by Fofonoff (1961) and Fine et al. (1978).

2. Gravitational potential energy changes

The gravitational potential energy of a vertical column of seawater extending from a level z_0 to the surface z_s is defined as

$$\begin{aligned} \text{GPE} &= \int_{z_0}^{z_s} \rho g (z - z_0) dz = \int_{z_0}^{z_s} p dz \\ &= \int_0^M p \alpha dm, \end{aligned} \quad (1)$$

where ρ is density, α specific volume, g gravity, p pressure, and $m = \int_{z_0}^z \rho dz$ is the mass variable measured from z_0 of the water column of total mass M . By using a mass variable, the column shifts relative to the gravitational field are taken into account while conserving mass. Variations of gravity with depth are neglected.

The changes of GPE can be estimated from the hydrodynamic equations by evaluating the contributions from changes of $p\alpha$. The local volume changes produced by advection and mixing cause global shifts of the water column relative to the geopotential field and conversions between GPE and kinetic energy. Expanding $p\alpha$ in terms of adiabatically and isentropically conserved variables yields

$$\begin{aligned} \frac{Dp\alpha}{Dt} &= \frac{dp\alpha}{dt} + u_j \frac{dp\alpha}{dx_j} \\ &= \left(\frac{\partial p\alpha}{\partial \eta} \right)_{p,s} \frac{D\eta}{Dt} + \left(\frac{\partial p\alpha}{\partial p} \right)_{\eta,s} \frac{Dp}{Dt} + \left(\frac{\partial p\alpha}{\partial s} \right)_{\eta,p} \frac{Ds}{Dt}, \end{aligned} \quad (2)$$

where η is entropy and s salinity. The changes relative to an adiabatic pressure change along streamlines are given by

$$\begin{aligned} \frac{Dp\alpha}{Dt} - \left(\frac{\partial p\alpha}{\partial p} \right)_{\eta,s} \frac{Dp}{Dt} \\ = p \left[\left(\frac{\partial \alpha}{\partial \eta} \right)_{p,s} \frac{D\eta}{Dt} + \left(\frac{\partial \alpha}{\partial s} \right)_{\eta,p} \frac{Ds}{Dt} \right]. \end{aligned} \quad (3)$$

Substituting for $D\eta/Dt$, the thermodynamic relationship:

$$\begin{aligned} \frac{D\eta}{Dt} &= \left(\frac{\partial \eta}{\partial T} \right)_{p,s} \frac{DT}{Dt} + \left(\frac{\partial \eta}{\partial p} \right)_{T,s} \frac{Dp}{Dt} + \left(\frac{\partial \eta}{\partial s} \right)_{T,p} \frac{Ds}{Dt} \\ &= \frac{C_p}{T} \frac{DT}{Dt} - \left(\frac{\partial \alpha}{\partial T} \right)_{p,s} \frac{Dp}{Dt} - \left(\frac{\partial \mu}{\partial T} \right)_{p,s} \frac{Ds}{Dt} \\ &= \frac{C_p}{T} \frac{D\theta}{Dt} - \left(\frac{\partial \mu}{\partial T} \right)_{p,s} \frac{Ds}{Dt}, \end{aligned}$$

where T is temperature, θ potential temperature, C_p specific heat, μ chemical potential of salt in seawater, and substituting

$$\begin{aligned} \left(\frac{\partial \alpha}{\partial s} \right)_{\eta,p} &= \left(\frac{\partial \alpha}{\partial s} \right)_{T,p} + \left(\frac{\partial \alpha}{\partial T} \right)_{p,s} \left[- \left(\frac{\partial \eta}{\partial s} \right)_{T,p} / \left(\frac{\partial \eta}{\partial T} \right)_{p,s} \right] \\ &= \left(\frac{\partial \alpha}{\partial s} \right)_{T,p} + \Gamma \left(\frac{\partial \mu}{\partial T} \right)_{p,s}, \end{aligned}$$

where

$$\Gamma = \frac{T}{C_p} \frac{\partial \alpha}{\partial T}$$

is the adiabatic temperature gradient, yields

$$\begin{aligned} \frac{Dp\alpha}{Dt} - \left(\frac{\partial p\alpha}{\partial p} \right)_{\eta,s} \frac{Dp}{Dt} \\ = p \left[\left(\frac{\partial \alpha}{\partial T} \right)_{p,s} \frac{D\theta}{Dt} + \left(\frac{\partial \alpha}{\partial s} \right)_{T,p} \frac{Ds}{Dt} \right] = \frac{D\chi}{Dt}, \end{aligned} \quad (4)$$

where the substantial derivative $D\chi/Dt$ is the rate of change of GPE due to nonadiabatic contributions.

The changes of potential temperature and salinity by mixing and diffusion can be estimated as flux divergences and local sources; that is,

$$\frac{D\theta}{Dt} = - \frac{dF_j^q}{dx_j} + q, \quad \frac{Ds}{Dt} = - \frac{dF_j^s}{dx_j}, \quad (5)$$

where $C_p F_j^q$ is the heat flux and $C_p q$ the heat produced by local dissipation of kinetic energy. The salt flux F_j^s is assumed not to be accompanied by any local sources.

Making the substitutions and rearranging the equation into a flux divergence plus source terms produces the equation for estimating nonadiabatic contributions to GPE:

$$\begin{aligned} \frac{D\chi}{Dt} &= - \frac{d}{dx_j} \left(p \frac{\partial \alpha}{\partial T} F_j^q + p \frac{\partial \alpha}{\partial s} F_j^s \right) + \frac{d}{dx_j} \left(p \frac{\partial \alpha}{\partial T} \right) F_j^q \\ &\quad + \frac{d}{dx_j} \left(p \frac{\partial \alpha}{\partial s} \right) F_j^s + p \frac{\partial \alpha}{\partial T} q \\ &= - \frac{dF_j^x}{dx_j} + \chi_s + \chi_q, \end{aligned} \quad (6)$$

where

$$\begin{aligned} F_j^x &= p \frac{\partial \alpha}{\partial T} F_j^q + p \frac{\partial \alpha}{\partial s} F_j^s \\ \chi_s &= \frac{d}{dx_j} \left(p \frac{\partial \alpha}{\partial T} \right) F_j^q + \frac{d}{dx_j} \left(p \frac{\partial \alpha}{\partial s} \right) F_j^s \\ \chi_q &= p \frac{\partial \alpha}{\partial T} q. \end{aligned} \quad (7)$$

TABLE 1. Location of Fig. 2 CTDs.

Ocean	Station File	Latitude	Longitude
(a) Arctic	94700124.CTD	89°58'N	91°41'E
(b) Antarctic	HYD13.TXT	66°53'S	68°56'W
(c) Indian	I08S0027.WCT	42°0'S	95°0'E
(d) Indian	I07N75901.CTD	0°0'	57°16'E
(e) Pacific	KF.41.CTD	34°0'N	144°0'E
(f) Atlantic	018FLIS	34°39'N	72°52'W

The fluxes F_j^χ are the contributions from the heat and salt fluxes, χ_s the local source contribution, and χ_q the positive contribution from local dissipation of kinetic energy.

For an isolated segment of the water column, assuming no fluxes across the boundaries and neglecting the small contribution from local dissipation of kinetic energy, the equation reduces to

$$\frac{d\chi}{dt} \approx \chi_s = \frac{d}{dz} \left(p \frac{\partial \alpha}{\partial T} \right) F_z^q + \frac{d}{dz} \left(p \frac{\partial \alpha}{\partial s} \right) F_z^s. \quad (8)$$

The fluxes can be estimated using different models, but the minimum fluxes are assumed to be by molecular diffusion and small-scale mixing and can be approximated by

$$F_z^q \approx -k_T \frac{dT}{dz}, \quad F_z^s \approx -k_s \frac{ds}{dz}, \quad (9)$$

where k_T , k_s are the diffusion or turbulent transfer coefficients for temperature and salinity.

The equation becomes

$$\chi_s = -k_T \frac{d}{dz} \left(p \frac{\partial \alpha}{\partial T} \right) \frac{dT}{dz} - k_s \frac{d}{dz} \left(p \frac{\partial \alpha}{\partial s} \right) \frac{ds}{dz}. \quad (10)$$

At small scales, the molecular diffusion of salt is negligible in comparison with that of heat; that is, $k_T \gg k_s$. The thermal stability parameter is, therefore, defined as

$$E_{\text{ther}} = -\frac{d}{dz} \left(p \frac{\partial \alpha}{\partial T} \right) \frac{dT}{dz} \quad (11)$$

and must be positive for a stable stratification. For equal transfer rates for heat and salt by small-scale turbulent mixing and for a linear EOS, the stability expression reduces to

$$\chi_s = k\rho g \left(\frac{\partial \alpha}{\partial T} \frac{d\theta}{dz} + \frac{\partial \alpha}{\partial s} \frac{ds}{dz} \right) = kN^2, \quad (12)$$

where N^2 is the local stability parameter and k the transfer rate. For turbulent exchange, the temperature gradient is replaced by the potential temperature gradient $d\theta/dz$.

A limiting vertical diffusive temperature gradient $(dT/dz)_{\text{diff}}$ is defined by $E_{\text{ther}} = 0$, or $p\partial\alpha/\partial T$ constant with depth. For $p\partial\alpha/\partial T$ to be constant with depth, the thermal

expansion must decrease sufficiently rapidly to offset the pressure increase, indicating a strong decrease in the temperature. Thus, plotting $p\partial\alpha/\partial T$ against temperature provides a sensitive diagnostic tool to identify critical regions in the oceans. Examples are given in Fig. 2 and Table 1 for several ocean regions.

At the diffusive limit, the linear and nonlinear contributions to GPE are of equal magnitude and opposite sign. The nonlinear contribution is always negative and destabilizing. The limiting diffusive temperature gradient is obtained by solving the equation:

$$\frac{d}{dz} \left(p \frac{\partial \alpha}{\partial T} \right) = -\rho g \left[\frac{\partial \alpha}{\partial T} + p \left(\frac{\partial^2 \alpha}{\partial T \partial p} \right) \right] + p \left[\frac{\partial^2 \alpha}{\partial T^2} \left(\frac{dT}{dz} \right)_{\text{diff}} + \frac{\partial^2 \alpha}{\partial T \partial s} \frac{ds}{dz} \right] = 0, \quad (13)$$

which yields

$$\left(\frac{dT}{dz} \right)_{\text{diff}} = \frac{\rho g \left[\frac{\partial \alpha}{\partial T} + p \left(\frac{\partial^2 \alpha}{\partial T \partial p} \right) \right] - p \left(\frac{\partial^2 \alpha}{\partial T \partial s} \right) \frac{ds}{dp}}{p \left(\frac{\partial^2 \alpha}{\partial T^2} \right)}. \quad (14)$$

The diffusive limit $(dT/dz)_{\text{diff}}$ apparently sets a strong upper bound at all depths to the magnitudes of the vertical temperature gradients that can persist in the ocean. Gradients above this limit must be actively forced as, for example, in surface thermoclines below the wind mixed layer. An example from the main thermoclines of the Atlantic and Pacific subtropical basins is shown in Fig. 3. The measured gradients fluctuate about the diffusive limit within the main thermocline. The interpretation is that mixing converts sufficient GPE to provide the energy to form the “slippery” layered structures that move relative to each other to force the mixing at the interfaces. A mechanism for creating slopes at the thermocline, such as internal waves, may enhance the process. A crude analogy is a block resting on a sloping surface. When a movement produces a GPE change greater than the frictional dissipation at the interface, the surface becomes “slippery” and the block begins to slide.

A vertical average shows a balance between the energy production by cabbeling and the dissipation by diffusion within the thermocline yielding a neutral thermal stability. Thus, the nonlinear effects limit the intensity of the thermocline stratification that is formed by the convergence and divergence of the large-scale basin circulation. At this limit, the mixing is strongly powered by the nonlinear internal sources.

3. Thermal and turbulent stability parameters

For comparison to the thermal and local stability parameters, Fofonoff (1998) defined a turbulent stability

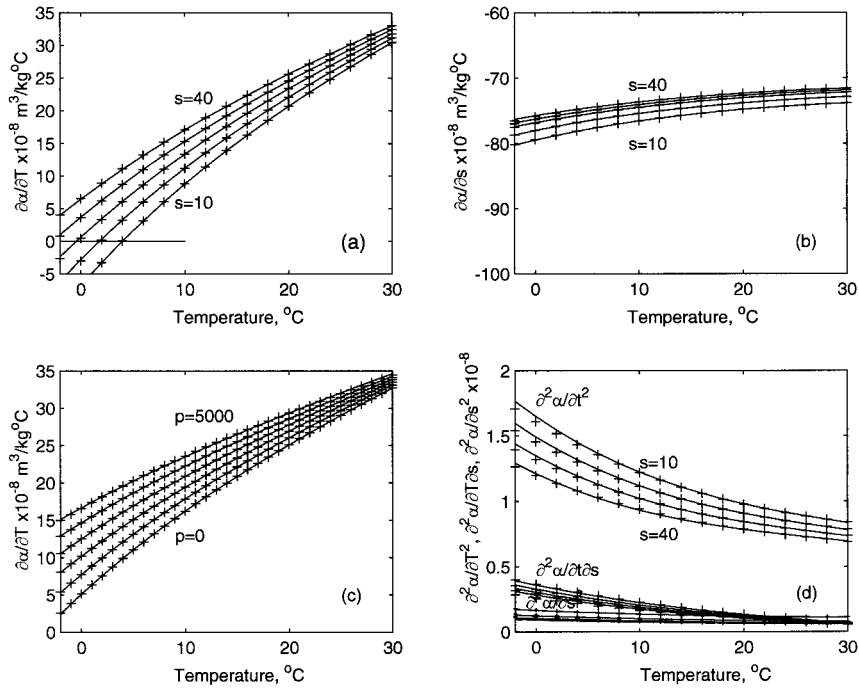


FIG. 1. Magnitudes of the first- and second-order derivatives of specific volume with respect to temperature and salinity. The lines are calculated from EOS-80 and the crosses (+) from the Feistel–Hagen (1995) equations of state of seawater. Variations are plotted with respect to temperature $-2(2)40$, salinity $10(10)40$, and pressure $0(1000)5000$. (a) Thermal variation of specific volume $\partial\alpha/\partial t$. (b) Saline variation $\partial\alpha/\partial s$. (c) Pressure variation of $\partial\alpha/\partial t$. (d) Variation of the second derivatives with respect to temperature and salinity. The same geometric scale is used for the temperature and salinity units.

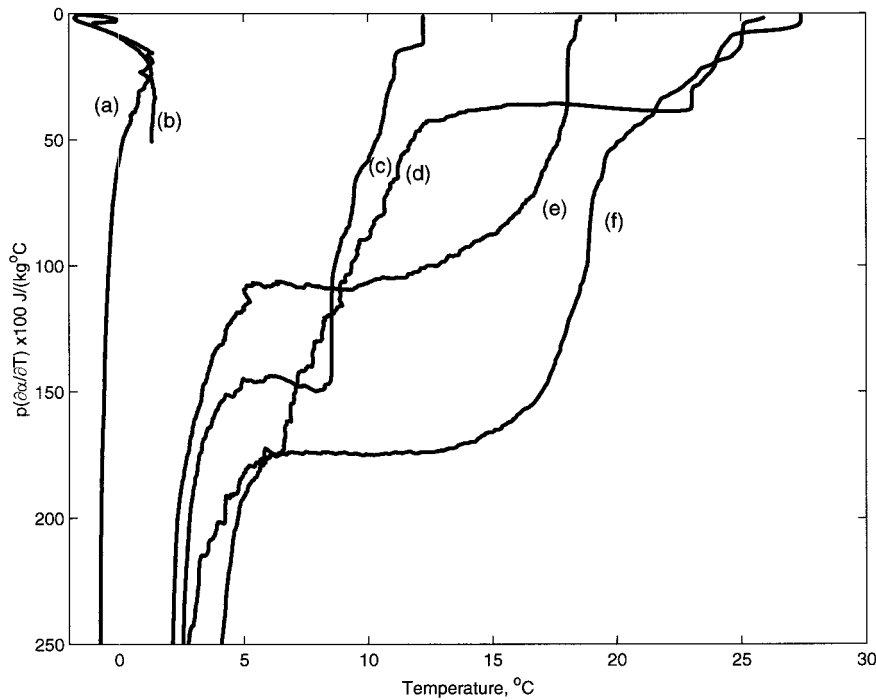


FIG. 2. Examples from several oceans showing thermoclines at or near the critical thermal gradient. The CTD station locations (a–f) are given in Table 1. The polar stations have destabilizing temperature gradients of the same magnitude as the capping instability.

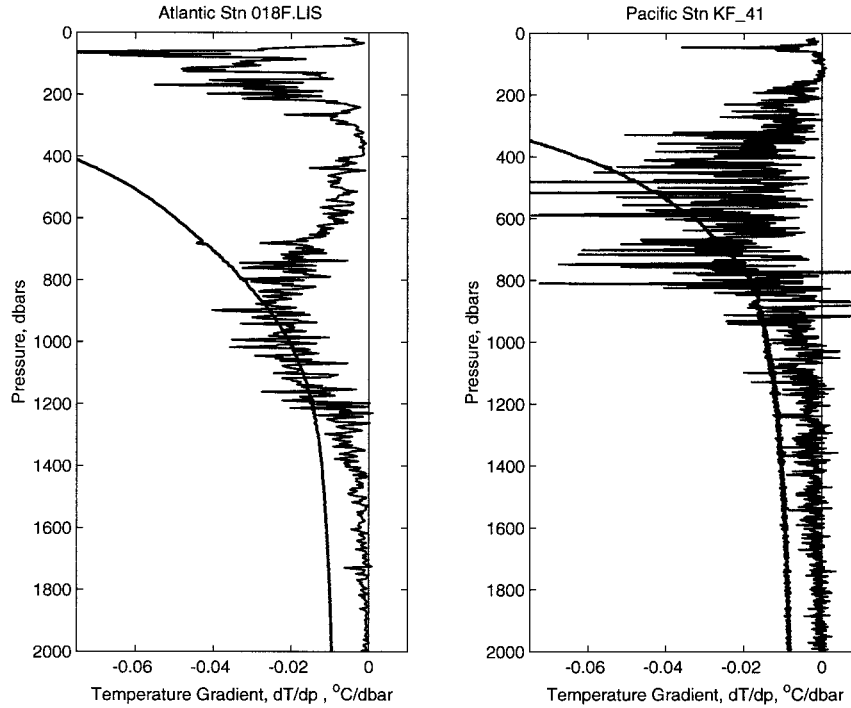


FIG. 3. Vertical temperature gradients compared with the diffusive limit $(dT/dp)_{diff}$ for Station 018FLIS from the North Atlantic and Station KF_41.CTD from the North Pacific subtropical basins.

parameter E_{tot} based on a vertical adiabatic exchange of two mass elements followed by mixing of the elements with a mass fraction ϵ of surrounding water. The parameter can be expressed as

$$E_{tot} = E_{loc} + E_{ext} - \frac{\epsilon}{1 + \epsilon} E_{mix}, \quad (15)$$

where

$$E_{loc} = N^2 = \rho g \left(\frac{\partial \alpha}{\partial T} \frac{d\theta}{dz} + \frac{\partial \alpha}{\partial s} \frac{ds}{dz} \right), \quad (16)$$

$$E_{ext} = -gp \left(\frac{\partial \kappa}{\partial T} \frac{d\theta}{dz} + \frac{\partial \kappa}{\partial s} \frac{ds}{dz} \right), \quad (17)$$

where

$$\kappa = \alpha^{-1} (\partial \alpha / \partial p)_a$$

is the adiabatic compressibility coefficient and

$$E_{mix} = p \left[\frac{\partial^2 \alpha}{\partial T^2} \left(\frac{d\theta}{dz} \right)^2 + 2 \frac{\partial^2 \alpha}{\partial T \partial s} \frac{d\theta}{dz} \frac{ds}{dz} + \frac{\partial^2 \alpha}{\partial s^2} \left(\frac{ds}{dz} \right)^2 \right] \quad (18)$$

is the cabbeling contribution.

The thermal stability parameter can be expressed in a similar form:

$$E_{ther} = ET_{loc} + ET_{ext} - ET_{mix}, \quad (19)$$

where

$$ET_{loc} = \rho g \frac{\partial \alpha}{\partial T} \frac{dT}{dz}$$

$$ET_{ext} = \rho g p \frac{\partial^2 \alpha}{\partial T \partial p} \frac{dT}{dz}$$

$$ET_{mix} = p \left[\frac{\partial^2 \alpha}{\partial T^2} \left(\frac{dT}{dz} \right)^2 + \frac{\partial^2 \alpha}{\partial T \partial s} \frac{ds}{dz} \frac{dT}{dz} \right].$$

The major difference between the thermal and turbulent stability parameters is that E_{tot} can have stabilizing or destabilizing contributions from the vertical salinity stratification and the mixing ratio parameter ϵ that determines the magnitude of the nonlinear cabbeling contribution. A reviewer pointed out that the turbulent stability parameter reduces to the diffusive stability parameter for complete mixing, $\epsilon \rightarrow \infty$, in the absence of a salinity gradient. This may lead to an alternative interpretation of the thermal stability parameter.

An example from the North Atlantic is shown in Fig. 4. For a mixing ratio $\epsilon = 1.0$, both the thermal and turbulent parameters approach neutral values. For both parameters to be equal to zero, we must have

$$E_{tot} - E_{ther} = 0.0 \approx \rho g \left[\frac{1}{1 + \epsilon} \frac{\partial \alpha}{\partial T} \frac{d\theta}{dz} + \frac{\partial \alpha}{\partial s} \frac{ds}{dz} \right], \quad (20)$$

neglecting small compressibility terms. In terms of the density ratio,

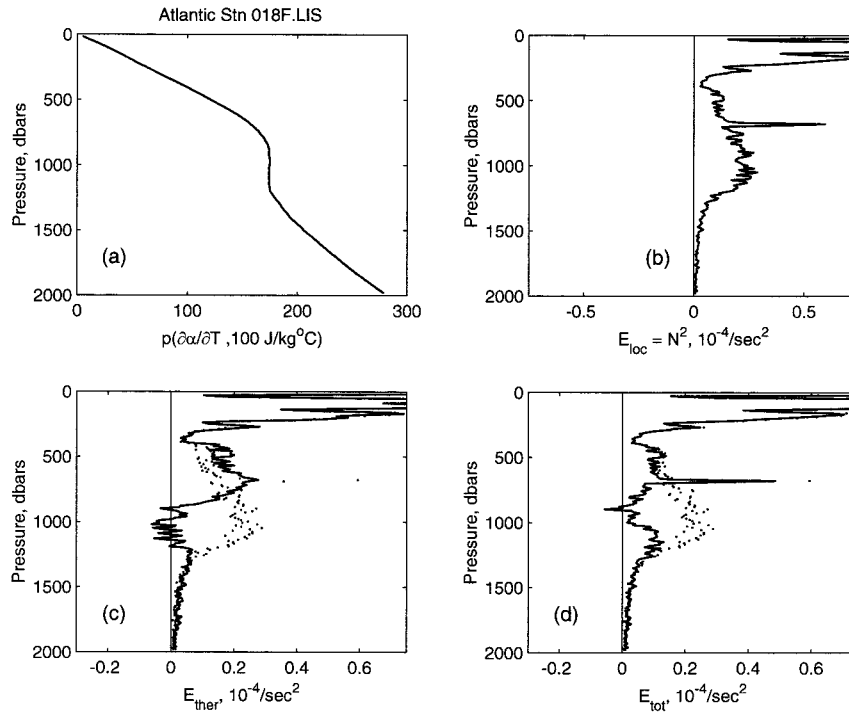


FIG. 4. Comparison of the thermal and turbulent stability parameters for Station 018FLIS in the North Atlantic subtropical basin near the Gulf Stream. The local stability profile, E_{loc} , is repeated in (c) and (d) for comparison. The mixing ratio is $\epsilon = 1.0$.

$$R_p = -\left(\frac{\partial\alpha}{\partial T} \frac{d\theta}{dz}\right) / \left(\frac{\partial\alpha}{\partial s} \frac{ds}{dz}\right)$$

(Schmitt 1981), the relationship is

$$R_p \approx 1 + \epsilon. \quad (21)$$

For both thermal and turbulent parameters to be zero, the GPE sum of the destabilizing salinity gradient and the cabbeling must balance the stabilizing temperature gradient. For $\epsilon = 1.0$, these are equal in magnitude, yielding a density ratio $R_p = 2$. For lower salinity gradients, the mixing ratio would be greater than one.

Another example from the Pacific Ocean is shown in Fig. 5. The stability parameters show neutral or negative values at the thermocline where the density ratio is about 4.0. For this region, cabbeling provides about 75% of the GPE conversion. The temperature fluctuations about the diffusive limit are greater in magnitude than the Atlantic example, where cabbeling provides 50% of the destabilizing energy. The density ratios for the Atlantic and Pacific stations are shown in Fig. 6.

In polar regions, where destabilizing temperature gradients occur, the vertical temperature gradients can increase until the nonlinear cabbeling contribution to GPE is of the same order of magnitude as the destabilizing temperature gradient. The combination of cabbeling and the destabilizing temperature gradient produces a “slippery” layer which enables intrusive layers to form more readily.

It is interesting to note that the cabbeling magnitude does not exceed, on the average, the GPE contribution from the temperature gradient. Thus, for the polar regions a density ratio of 0.5 is found for the intrusive layers. Examples are shown for the Arctic and Antarctic Oceans in Fig. 7.

A useful diagnostic parameter can be formed from the ratio of the thermal and local stability parameters. Defining Q as

$$Q = E_{ther}/E_{loc}. \quad (22)$$

allows the evaluation of contributions of cabbeling, and temperature and salinity gradients to the overall stability of the ocean structure. Figure 8 shows the Q profiles for the stations listed in Table 1. Because the critical thermocline gradients occur at the maxima of E_{loc} , the parameter Q emphasizes the regions where cabbeling is significant. At other depths, values greater than one indicate a destabilizing salinity gradient found, for example, in the deep waters of the Atlantic and Indian Oceans. The Pacific station shows a linear shift between the destabilizing temperature gradient and the stabilizing salinity gradient. This could be a significant feature to help understand deep water mixing.

4. Conclusions

Based on observations, the nonlinear cabbeling effect plays a prominent role in the limiting of temperature

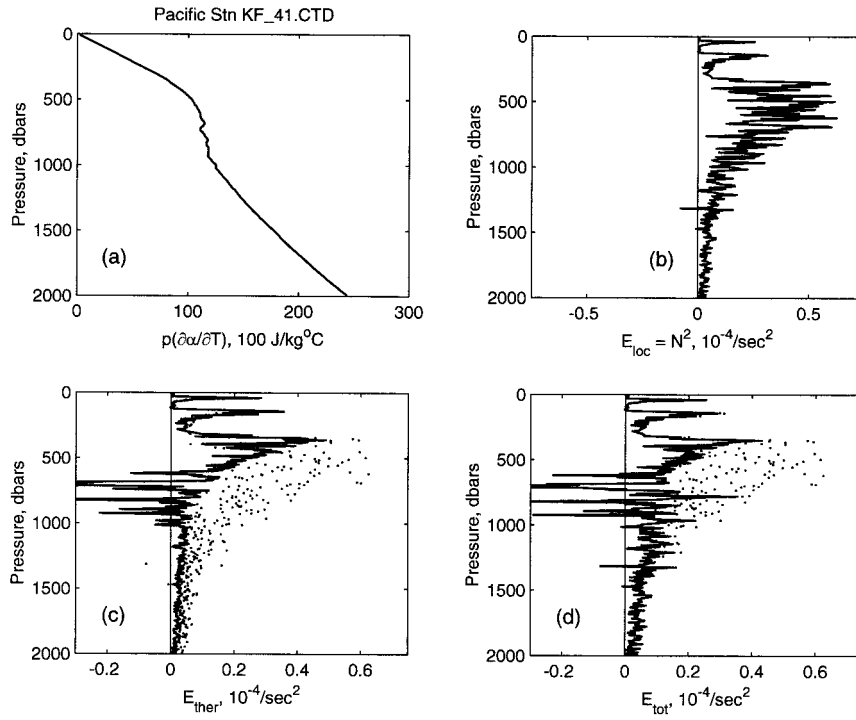


FIG. 5. Comparison of the thermal and turbulent stability parameters for station KF.41.CTD in the North Pacific subtropical basin. The local stability profile, E_{loc} , is repeated in (c) and (d) for comparison. The mixing ratio is $\epsilon = 3.0$.

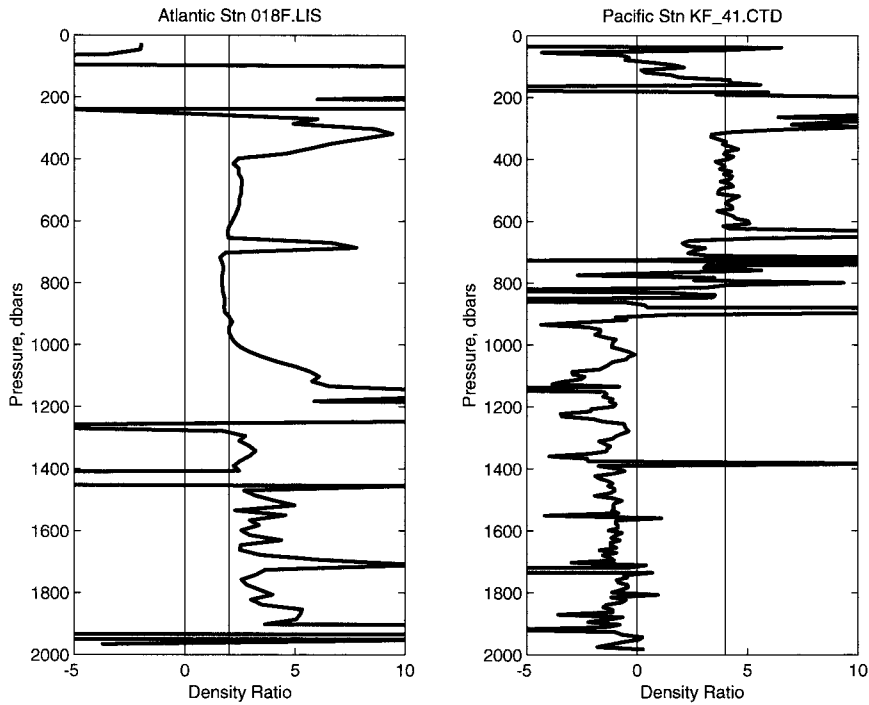


FIG. 6. Density ratios for Atlantic (018FLIS) and Pacific (KF.41.CTD) stations.

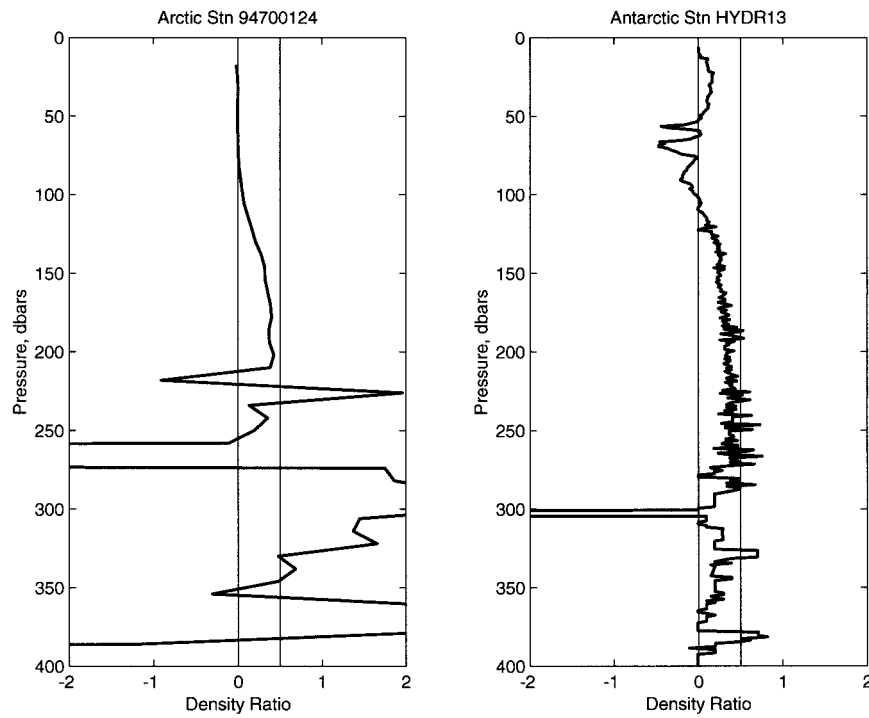


FIG. 7. Density ratios for Arctic (94700124.CTD) and Antarctic (HYD13.TXT) stations.

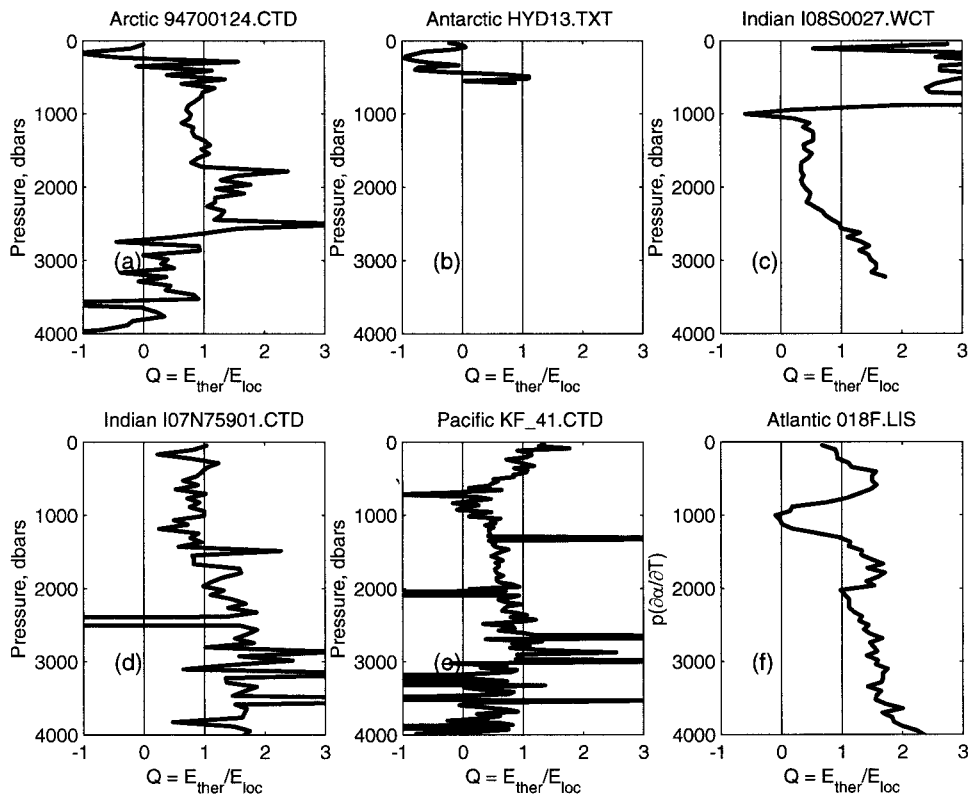


FIG. 8. Diagnostic ratio $Q = E_{ther}/E_{loc}$ for ocean stations listed in Table 1.

gradients in the ocean at all depths in all the major oceans. The critical diffusive temperature gradient determines the vertical gradient magnitude at which conversion of GPE by a local perturbation begins to exceed the work necessary to produce the perturbation. At this gradient magnitude, sufficient energy is converted to create the microstructure necessary to maintain the mixing although the mechanisms for this conversion remain unclear.

The effect is pronounced for the main thermoclines of the world's oceans where it acts to limit the vertical gradients. The convergent circulations that intensify the thermoclines are balanced by formation of layered "slippery" structures within the thermoclines that draw energy for mixing from the cabbelling process. At the critical vertical gradients, the energy for mixing is sufficient even in the absence of external mixing forces such as internal waves, tides, etc. An equivalent turbulent mixing parameter in numerical models would have to be dependent on the cabbelling process.

Clearly, the dynamical description of the cabbelling process and its effect on ocean structures need to be developed and examined in greater detail.

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REFERENCES

- Eckel, O., 1949: Über die Mischungsarbeit von stabil geschichteten Wassermassen. *Arch. Meteor.*, **1A**, 264–269.
- Ekman, V. M., 1934: Georg Wust. "Das Bodenwasser und die Gliederung der Atlantischen Tiefsee." *Wiss. Ergebn. d. Deutsch. Atlant. Exped. 1925–1927*. herausgeg. von Defant, band VI, Teil 1, Lief 1. Berlin und Leipzig, 1932. *J. Cons., Cons. Int. Explor. Mer*, **9**, 102–104.
- Feistel, R., and E. Hagen, 1995: On the *GIBBS* thermodynamic potential of seawater. *Progress in Oceanography*, Vol. 36, Pergamon, 249–327.
- Fine, R. A., C. N. Mooers, and F. J. Millero, 1978: Effects of nonlinear pressure-volume-temperature properties on the potential energy distribution in the Atlantic Ocean. *Deep-Sea Res.*, **25**, 15–22.
- Fofonoff, N. P., 1961: Energy transformations in the sea. *Fish. Res. Board Can. Man. Rep. Ser.*, **109**, 1–82.
- , 1985: Physical properties of seawater: A new salinity scale and equation of state for seawater. *J. Geophys. Res.*, **90**, 3332–3342.
- , 1995: Dynamical effects of cabbelling on thermocline structure (in Russian, with English abstract). *Okeanologia*, **35**, 824–832.
- , 1998: Nonlinear limits to ocean thermal structure. *J. Mar. Res.*, **56**, 793–811.
- Gill, A. E., 1973: Circulation and bottom water production in the Weddell Sea. *Deep-Sea Res.*, **20**, 111–140.
- Mamayev, O. I., 1975: *Temperature–Salinity Analysis of World Ocean Water*. Elsevier Oceanography Series, No. 11, Elsevier, 374 pp. (Translated by R. J. Burton.)
- McDougall, T. J., 1987: Thermobaricity, cabbelling and water mass conversion. *J. Geophys. Res.*, **92**, 5448–5464.
- , and W. K. Dewar, 1998: Vertical mixing and cabbelling in layered models. *J. Phys. Oceanogr.*, **28**, 1458–1480.
- Schmitt, R., 1981: Form of the temperature–salinity relationship in the Central Water: Evidence for double-diffusive mixing. *J. Phys. Oceanogr.*, **11**, 1015–1026.
- Witte, E., 1902: Zur theorie der stromkabbellungen. *Gaea: Natur und Leben*, **38**, 484–487.