The eddy-driven thermocline

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Abstract

The role of baroclinic eddies in transferring thermal gradients from the surface of the ocean to the abyss, and thus determining the stratification, is examined. The hypothesis is that the density differences imposed at the surface by differential heating are a source of available potential energy which can be released by mesocale eddies with horizontal scales of the order of 10 km. Eddy-fluxes balance the vertical diffusion of heat and thus determine the vertical scale of penetratation of horizontal gradients (i.e. the thermocline). This conjecture is in contrast with the current thinking that the deep stratification is determined by a balance of turbulence at smaller scales (which determine vertical mixing) and the large-scale ocean circulation.

Eddy-processes are analyzed in the context of a rapidly rotating primitive-equation flow driven by specified surface temperature, with isotropic diffusion and viscosity. Scaling-laws for the depth of the thermocline as a function of the external parameters are proposed.

Classical thermocline theories

Because the ocean is heated differentially from above, the vertical penetration scale, h, of the surface signal depends on vertical mixing, characterized by the diffusivity κ. In classical thermocline theories vertical diffusion of heat is balanced by heat transport convergence by the mean circulation. This leads to the following vertical scale of the thermocline:

$$h = \left(\frac{\kappa f L}{g\alpha\Delta T}\right)^{1/3}, \quad \text{(Welander, 1959)} \tag{1}$$

In the absence of downward Ekman pumping, L is the scale of the domain.

Here, we enquire how the scaling is altered when vertical diffusion of heat is balanced by the convergence of eddy heat transport rather than laminar upwelling. Eddy-transports are essential in the Antarctic Circumpolar Current and in sub-polar regions, where a Sverdrupian mean circulation is unavailable.

Hydrostatic flow driven by surface temperature

We examine buoyancy-driven flows in a 3-D domain on the f-plane by integrating the primitive equations. The domain area is 1000×1000 km^2 and all variables are periodic in x and y.

At the surface, $z = H$, we specify the temperature to be $T = \Delta T \cos(2\pi y/L)$ and require no stress. At the bottom, $z = 0$, impose no-flux of heat and no-slip.

The diffusivity, κ, and the viscosity, ν are isotropic in all three dimensions. The addition of a small hyperviscosity removes noise on the grid-scale in a time scale of 20 days.

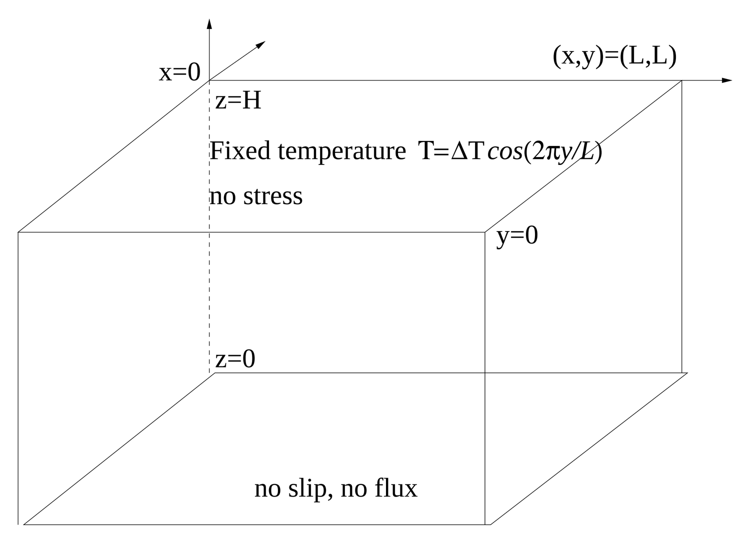

Fig. 1: The model domain. H is either 2000 or 4000 m.

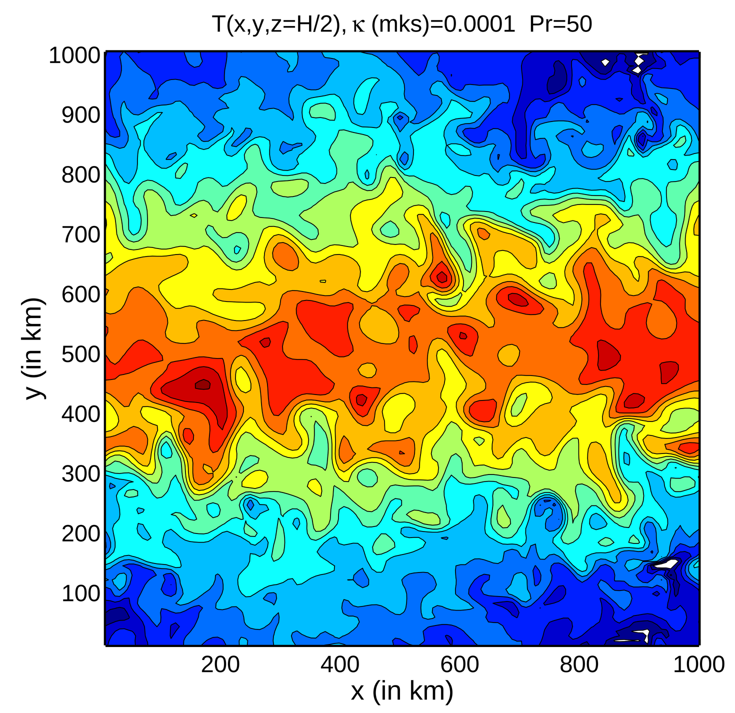

Fig. 2: A snapshot of the temperature at mid-depth: the mean temperature gradient is stirred by baroclinic eddies.

The zonally averaged fields

The statistically steady state has a shallow zonally averaged thermocline, maintained by a balance of vertical diffusion and convergence of eddy heat transport:

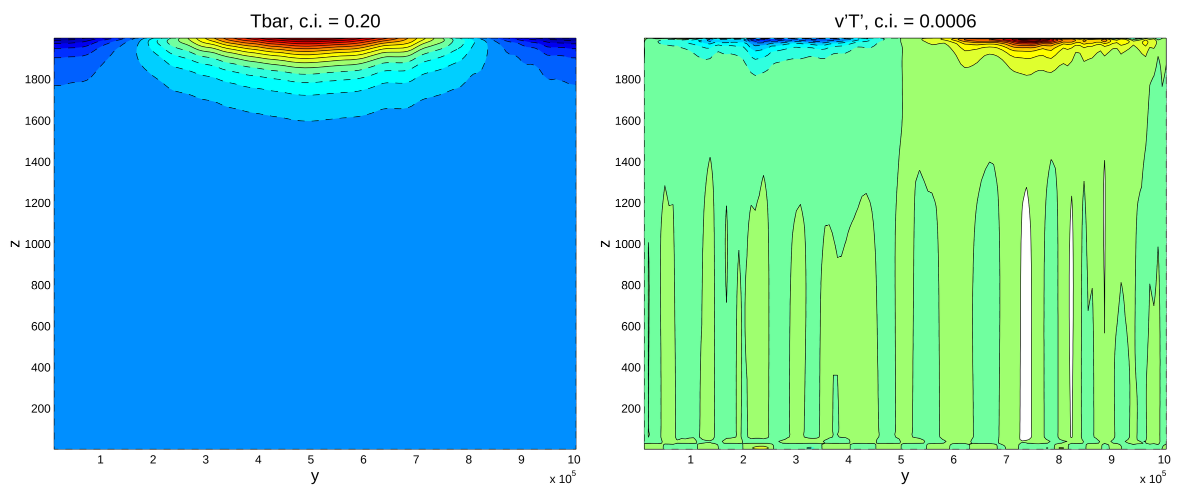

Fig. 3: The mean temperature, $\bar{T}$, and eddy heat transport, $\overline{v'T'}$ for $\kappa = 8 \times 10^{-4}$ m^2 s^{-1} and Prandtl=50.

There are two separate aspects of the mean temperature field: the horizontally averaged temperature $\langle T \rangle$ and the zonally varying mean temperature $\theta(y,z) = \bar{T} - \langle T \rangle$. They obey different dynamics:

$$\begin{aligned} \left(\overline{v'T'}\right)_y &= \kappa\theta_{zz}, \\ \langle w'T' \rangle &= \kappa \langle T \rangle_z. \end{aligned} \tag{2}$$

We focus on θ since the vertical structure of $\langle T \rangle$ depends on convective motions and secondary circulations not well resolved here.

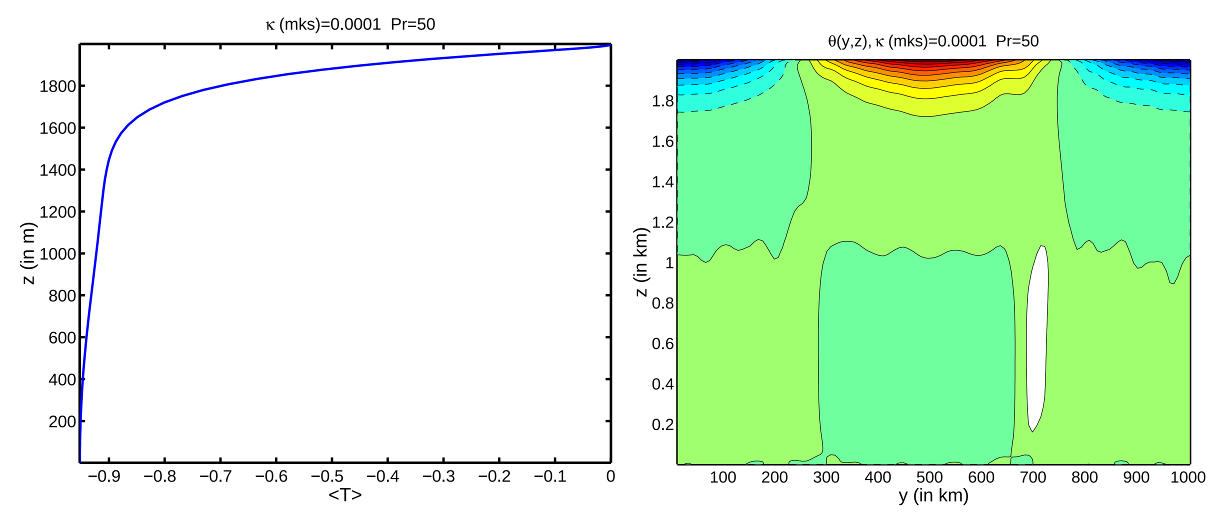

Fig. 4: $\langle T \rangle$ and θ for a typical set of parameters.

A scalar advected by the barotropic flow

The field θ compares very well with that obtained advecting a passive scalar identically forced at the surface but advected only by the vertically averaged velocities. This is because the baroclinic velocity vector is in thermal wind balance and thus largely orthogonal to the temperature gradient.

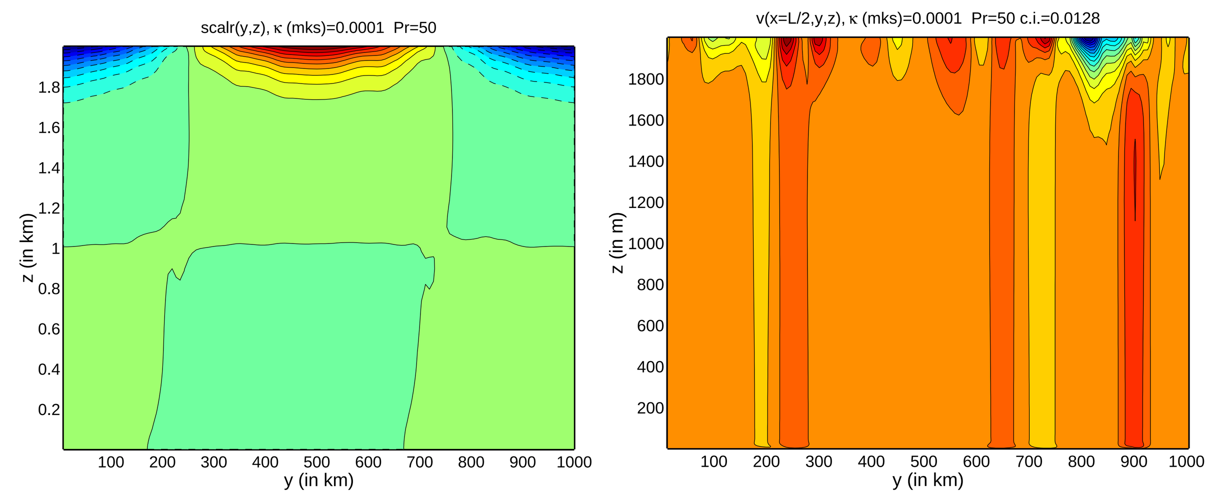

Fig. 5: Left: The zonally averaged passive scalar advected by the barotropic component is almost identical to θ which is advected by the 3-D velocity. Right: A snapshot of v for a section along the middle longitude.

The passive scalar, $S = \bar{S} + S'$, satisfies

$$\begin{aligned} S'_t + US'_x + \overbrace{\overline{V'S}_y}^{O(V'\bar{S}/L)} &= \overbrace{\kappa S'_{zz}}^{O(\kappa S'/h^2)} + O(V'S'), \\ \left(\overline{V'S'}\right)_y &= \kappa \bar{S}_{zz}, \end{aligned} \tag{3}$$

where U, V is the vertically averaged flow. The scalar variance equation then reads

$$+\overline{V'S'}\bar{S}_y = \kappa \overline{S'S'_{zz}}, \tag{4}$$

so that in statistical steady state the vertical penetration scale of the eddies is the same as the thermocline depth, h. The transport is then $\overline{V'S'} \sim -\bar{S}_y \overline{V'^2} h^2/\kappa$, and thus proportional to the variance of the barotropic flow. Using this scaling argument and the zonally averaged scalar balance we find

$$h \sim (\kappa L/V')^{1/2}. \tag{5}$$

The amplitude of the barotropic eddies, V', requires detailed knowledge of the statistics of baroclinic eddies. In the large Prandtl number limit considered here, the barotropic eddies do not undergo an inverse cascade, so we lack a theoretical framework that relates the baroclinic and barotropic eddy amplitudes. Our numerical results show that in this regime the depth of the thermocline is independent of the Prandtl number, ν/κ, and of the oceanic depth, H, and increases as $\sim \kappa^{1/3}$.

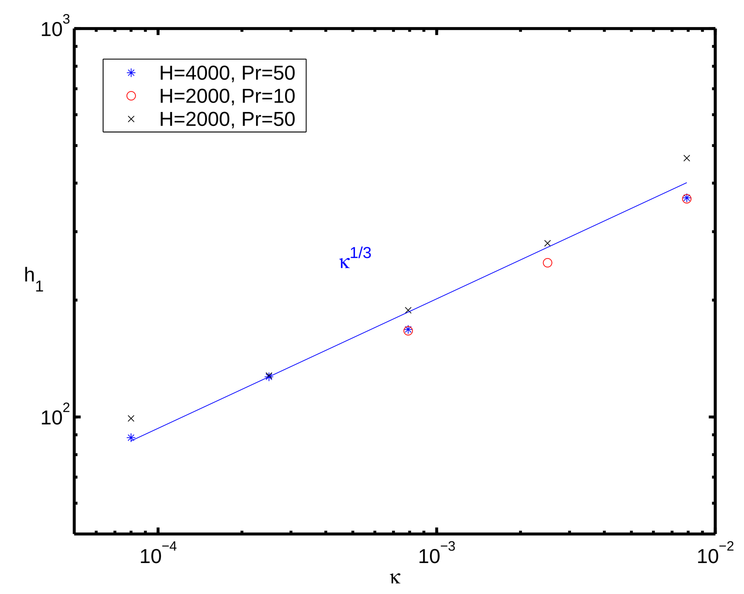

Fig. 6: The depth of the thermocline defined as $h_1 \equiv \left(\int_0^H \theta^2\, dz / \int_0^H \theta_z{}^2\, dz\right)^{1/2}$, as a function of the diffusivity, κ. There is no dependence on H or Pr.

Our numerical results further indicate that the dependence on the imposed horizontal temperature gradient, ΔT, is the same as that in the laminar theory of Welander's. We thus conclude that despite the fundamental difference in the two scenarios, the parameter dependence is the same in the eddy-driven and in the laminar Sverdrup case.

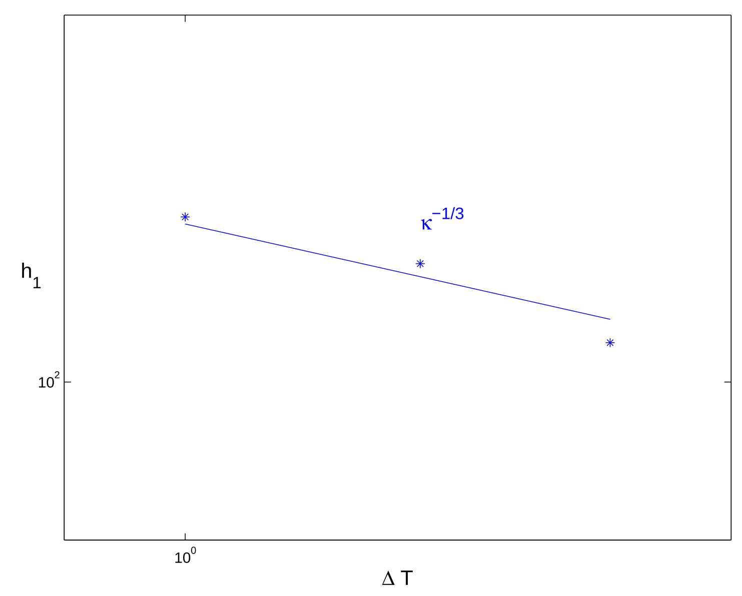

Fig. 7: The depth of the thermocline, h_1, as a function of the imposed temperature difference, ΔT.

Two Prandtl number regimes

In the large Prandtl number regime, the amplitude of the baroclinic eddies is determined by the global energy balance which reads

$$\underbrace{\nu\int_0^H \langle {u'_z}^2 + {v'_z}^2 \rangle\, dz}_{\text{Energy sink}} = \underbrace{\kappa g\alpha[\langle T\rangle_{z=H} - \langle T \rangle_{z=0}]}_{\text{Energy source}} = -\kappa g\alpha T_{abyss}.$$

Thus the baroclinic kinetic energy of the eddies per unit depth is given by

$$\mathrm{B}CKE = g\alpha T_{abyss}\, h^2 (H\, Pr)^{-1}. \tag{6}$$

In the small Prandtl number regime, dissipation occurs in the bottom boundary layer, dominated by the barotropic flow. Then the amplitude of the barotropic eddies is determined by

$$\mathrm{B}TKE = g\alpha T_{abyss} \sqrt{\kappa\, f}\, Pr^{-1/2}. \tag{7}$$

The abyssal temperature T_{abyss}, is of the same order as the horizontal temperature at the surface, ΔT.

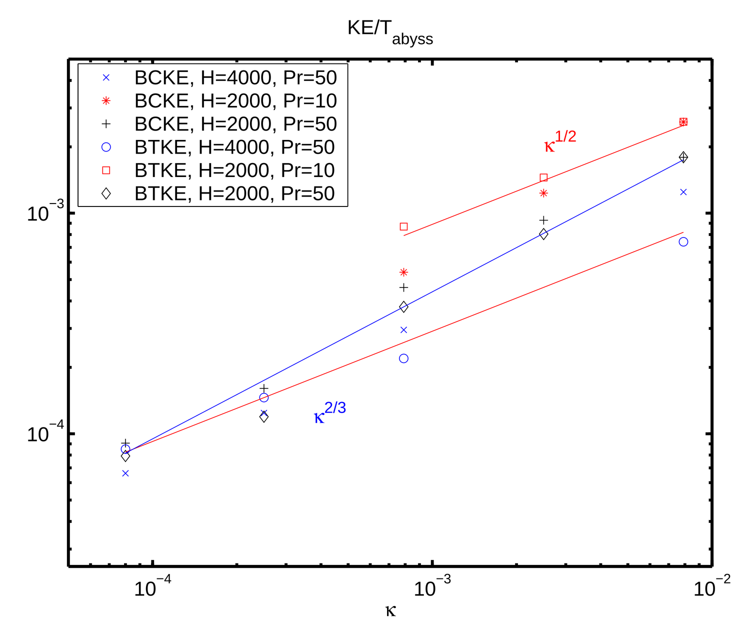

Fig. 8: The kinetic energy per unit depth, normalized by the abyssal temperature. The baroclinic kinetic energy satisfies the scaling (6) for the black and blue sequences. The barotropic kinetic energy for the red sequence satisfies (7).

Conclusion

- The eddies alone can produce a shallow thermocline, where eddy-driven meridional heat transport convergence balances downward heat diffusion.
- The eddy-driven transport is performed by the barotropic component of the flow, which is nonlinearly forced by the baroclinic eddies. We currently lack a framework for describing the statistics of the barotropic flow ($k^{-5/3}$ is inappropriate here).
- The vertical penetration scale of the surface signal depends on $\kappa^{1/3}$, as in Welander's original scaling, but for very different reasons.