Au-delà de la diffusion turbulente: la convection et ses paramétrisations

Groupe de Travail *MathsInFluids*

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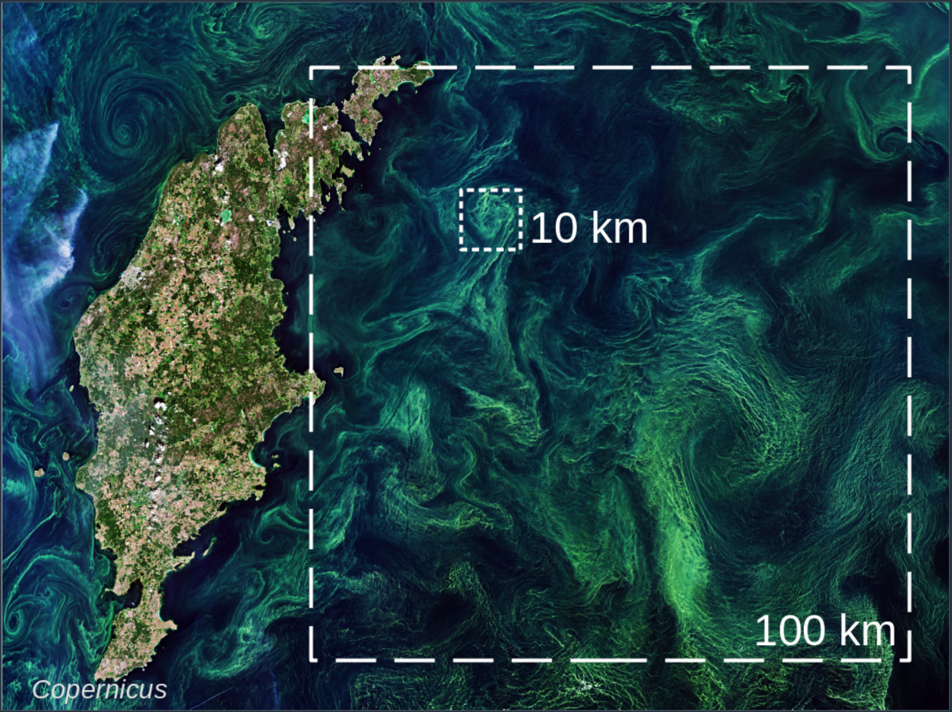
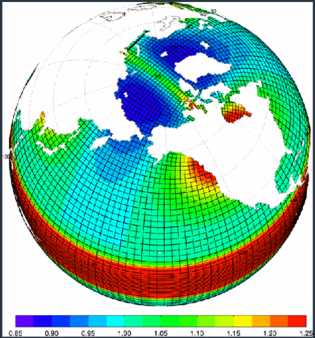
8 décembre 2023

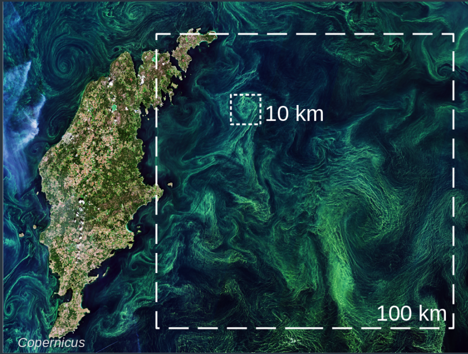
The need for parameterization in climate models


Copernicus


10 km
100 km
Copernicus

Discretization and subgrid modelling





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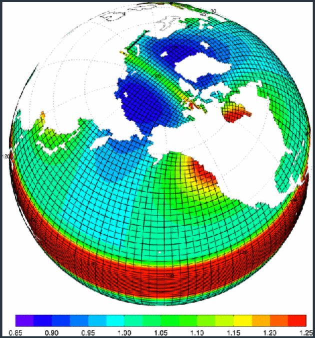

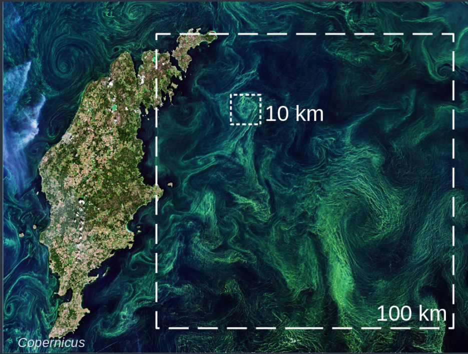

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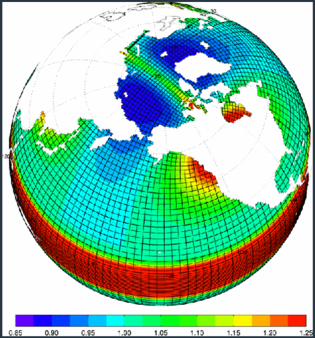

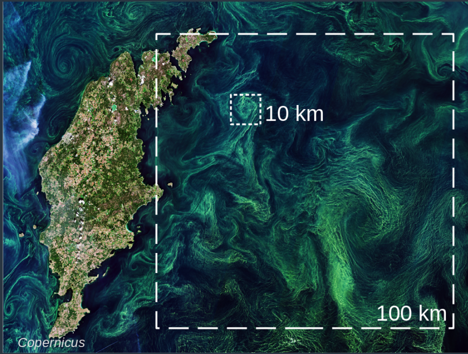

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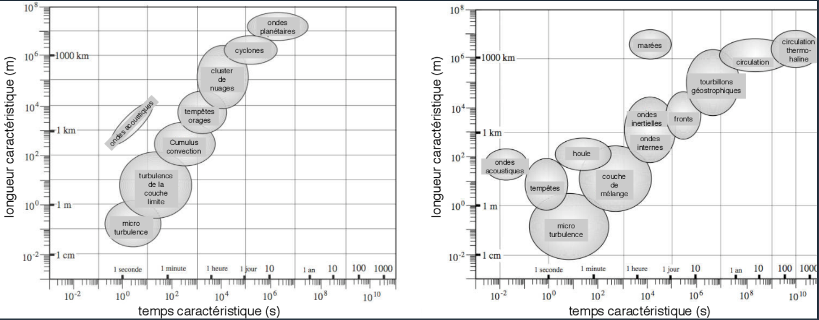
For each process: effects on large scales are **important**? yes $\Longrightarrow$ parameterization

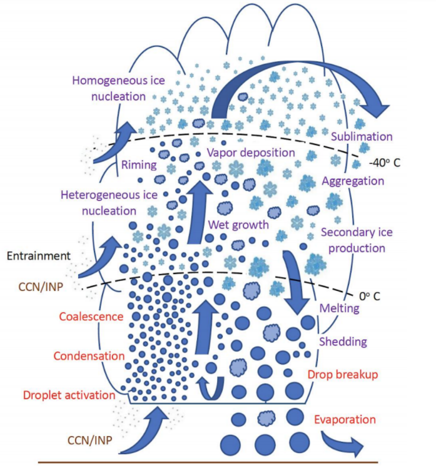
Turbulence and waves in ocean and atmosphere



adapted from von Storch et al., 1999

Apart from turbulence

Cloud micro-physics



Morrison *et al.* 2020

Radiative transfer

Wikimedia Commons

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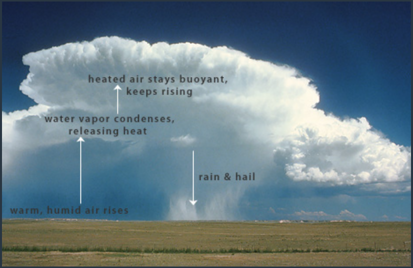
- ? scale awareness
- ? gray zone: what is resolved, what is unresolved?
- ? constraints, uncertainty quantification and tuning of "free" parameters

Convection in stratified fluids

Atmospheric convection

Shallow / boundary layer convection

Deep convection



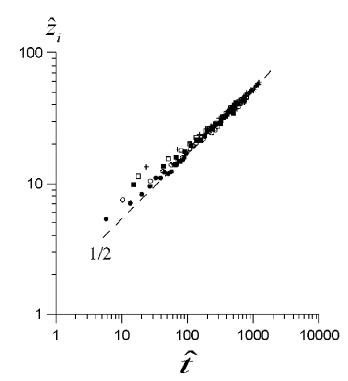
Oceanic Convection

- cooling
- evaporation
- double-diffusion
- brine rejection

+ video of Oceananigans from Ali Ramadhan

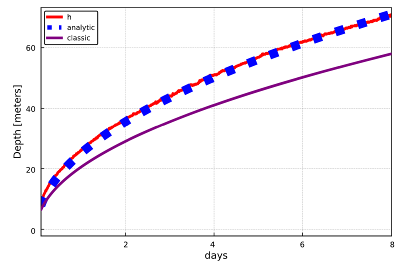
Introduction sur la convection dans l'atmosphère et dans l'océan + modèles d'ordre 0 $\longrightarrow$ *voir notes manuscrites*

Boundary layer height



Fedorovich 2004

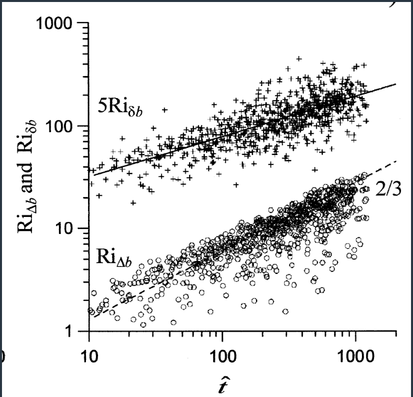
$$\hat{z}_i = \frac{h}{L_0},\ \hat{t} = tN_0$$



Souza 2020

$$h_{\text{analytic}} = \sqrt{3}\sqrt{2\frac{B_0}{N0^2}t}$$

Buoyancy jump



Fedorovich 2004

$$\mathrm{Ri}_{\Delta b} = \frac{\Delta b h}{w_*^2}$$

Limitations

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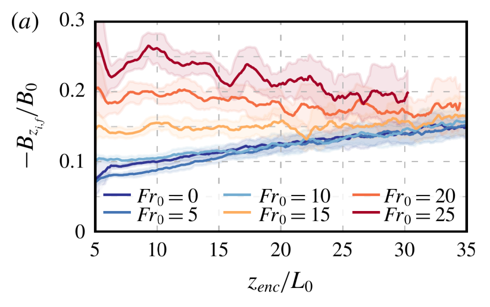
[Garcia & Mellado, 2014]: entrainment zone has a two-layer structure

$$\Delta h_1 \simeq 0.25h$$

$$\Delta h_2 \simeq 1.2L_0$$

ex: for oceanic deep convection, $B_0 \simeq 10^{-7}\mathrm{m}^2\,\mathrm{s}^{-3}$, $N_0^2 \simeq 10^{-7}\mathrm{s}^{-2}$
$L_0 \simeq 60m$

- Effects of mean wind(atmosphere) or surface stress (ocean)?



Hagshenas & Mellado 2019

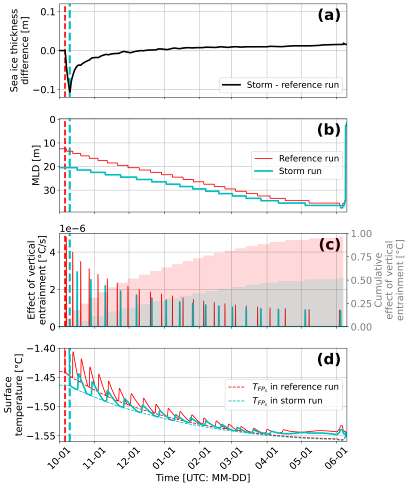
$$Fr_0 = \frac{U_0}{L_0 N_0}$$

$$\frac{B_0}{B_h} \simeq \left(0.45\left(\frac{h}{L_0}\right)^{-1} - 0.12\right)\frac{\Delta h}{0.25h}\left[0.82 + 0.18\frac{\Delta h}{0.25h}\right]$$

$$\Delta h = \frac{U_0 - u_m}{\partial_z \overline{u}}$$

$\longrightarrow$ Kelvin-Helmoltz instabilities increase vertical entrainment!

Ocean-Ice interactions



Bent et al., submitted

Proposed mechanism:
Ice formation ⟶ brine (salt) rejection ⟶ convection ⟶ entrainment of warmer water below
⟶ warming of mixed layer
⟶ inhibition of ice formation...

Summary and outlook

- simple "slab" models predicts main features
- in practice, it is not fully self-similar
- for boundary layer mathematicians: coupling surface layer (Monin-Obukhov), mixed layer and entrainment zone?

Eddy-Diffusivity Mass-Flux Models

SCM vs LES framework

A framework for the evaluation of vertical mixing parameterizations:

$\longrightarrow$ horizontal average of Large Eddy Simulations (LES – high-resolution) vs 1D Single Column Model (SCM)

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LES vs. SCM

$$\partial_t X + \nabla \cdot \boldsymbol{u} X = S_X \overset{\overline{(\cdot)} = \overline{(\cdot)}^{x,y}}{\longrightarrow} \overline{X}, \overline{w'X'} \quad \text{vs.} \quad \partial_t \overline{X}^{SCM} + \partial_z \overline{w'X'}^{SCM} = \overline{S}$$

$$X = \boldsymbol{u}, \theta, S...$$

Standard TKE parameterizations

Boussinesq hypothesis: analogy with molecular diffusion

$$\overline{w'X'} \overset{\text{param}}{=} -K_X \partial_z \overline{X}$$

Heuristic derivation: assume fluctuations X' are due to parcel motion w' at a distance l_X, then $X' \simeq l_X \partial_z \overline{X}$ and

$$\overline{w'X'} \simeq -l_X |w'| \partial_z \overline{X}$$

Motivates turbulent kinetic energy (TKE) closures

$$K_X = -c_X l_X \sqrt{k} \partial_z \overline{X}$$

Turbulent Kinetic Energy

TKE $k := \frac{1}{2}\overline{\boldsymbol{u}' \cdot \boldsymbol{u}'}$

$$\partial_t k + \partial_z T = Sh + B - \epsilon$$

where

- TKE fluxes $T := \overline{w' \frac{1}{2}\boldsymbol{u}' \cdot \boldsymbol{u}' + \frac{1}{\rho_0}\overline{w'p'} - \nu \partial_z k}$
- Shear production $Sh := -\overline{w'\boldsymbol{u}'_h} \cdot \partial_z \overline{\boldsymbol{u}}_h$
- Buoyancy production $B := \overline{w'b'}$
- viscous dissipation $\epsilon := \nu \overline{\nabla \boldsymbol{u}' : \nabla \boldsymbol{u}'}$

Commonly parameterized with eddy-diffusivity closures:

$$\partial_t k + \partial_z(-K_k \partial_z k) = K_m \partial_z \overline{\boldsymbol{u}}_h \cdot \partial_z \overline{\boldsymbol{u}}_h - K_\phi \partial_z \overline{b} - \frac{c_\epsilon}{l_\epsilon} k^{3/2}$$

Analytical results

Detailed review in the *neutral* case $(b = cst)$ in *[Chacón-Rebollo & Lewandowski 2014]* book:

- existence of weak solution for *stationary* TKE
- no such result for unsteady TKE models

Innacuracy of eddy-diffusivity

'local' Eddy-Diffusivity closure (ED): $\overline{w'\theta'} \overset{\text{param}}{=} -K_\theta \partial_z \overline{\theta}$

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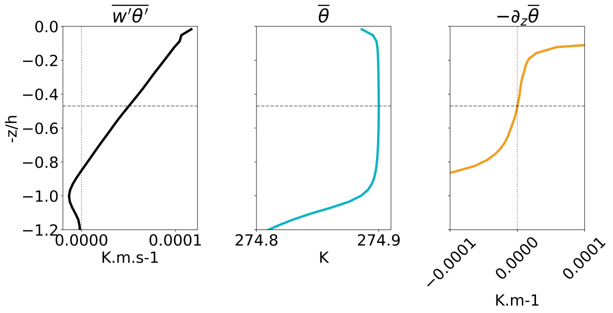

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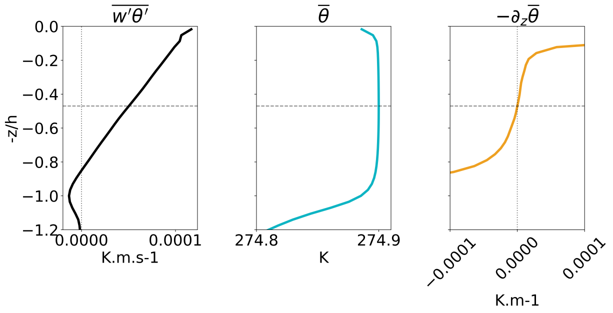

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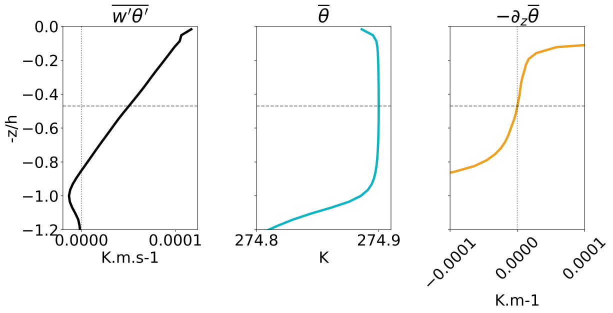

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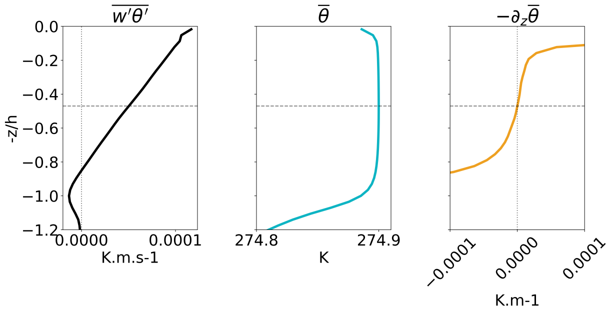

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Non-local mixing

Atmospheric parameterization of *[Troen & Mahrt 86]*

$\longrightarrow$ In the ocean: **K-profile parameterization** (KPP) *[Large et al., 94]*

$$\overline{w'\theta'} \overset{\text{param}}{=} -K_\theta(\partial_z\overline{\theta}-\gamma_\theta)$$

$$K_\theta, \gamma_\theta \overset{\text{param}}{=} \text{functions}(z; \overline{w'b'}_{sfc}, \mathrm{MLD}, Ri)$$

(MLD $=$ Mixed Layer Depth, $\simeq$ inversion height z_i in the atmosphere)

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In NEMO: TKE scheme + **Enhance Vertical Diffusion (EVD)** to $K \simeq 10m^2.s^{-1}$ when $N^2 < 0$

Eddy-diffusivity - Mass-Flux models

Combine:

- Eddy-diffusivity (ED): small scale isotropic turbulence

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First models for the atmospheric boundary layer: *[Hourdin et al. 2002, Soares et al. 2004]*

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First use for oceanic convection: *[Giordani et al. 2021]*

Formal scale separation and assumptions: *[Yano 2014]*
Unsteady plumes and multifluid models: *[Tan et al. 2018, Thuburn et al. 2018]*

In my presentation

Focus on methodological aspects

- **derivation** of EDMF schemes for ocean models:

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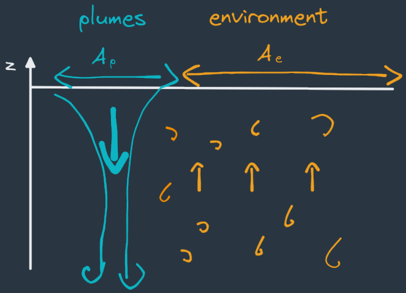
- **energy budgets**
- **sensitivity** to parameters and **calibration**

What can be formaly done?

Two-domain decomposition



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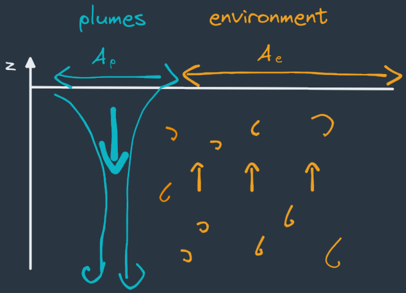

$$\int_A X\,\mathrm{d}x\mathrm{d}y = \frac{A_p}{A}\int_{A_p} X\,\mathrm{d}x\mathrm{d}y + \frac{A_e}{A}\int_{A_e} X\,\mathrm{d}x\mathrm{d}y$$

$$\updownarrow$$

$$\overline{X} = a_p X_p + a_e X_e$$

a_p, a_e: *fractional* areas of subdomains $a_p + a_e = 1$

Decomposition of fluxes

Apply this decomposition to fluxes:

$$\overline{w'X'} = \underbrace{a_p(w_p - \overline{w})(\theta_p - \overline{\theta})}_{\text{coherent structures}} + \underbrace{a_p \overline{w'_p \theta'_p}^p}_{\text{subdomain turbulence}}$$

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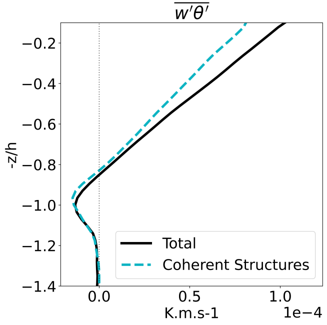

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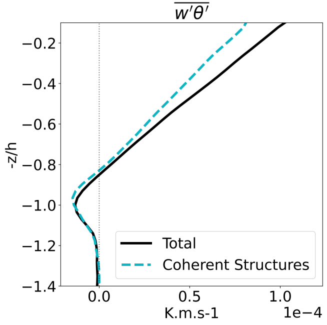

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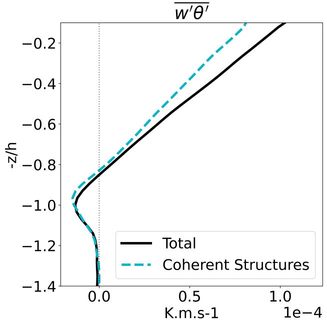

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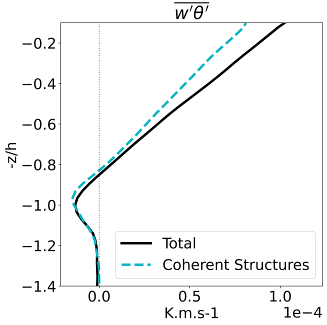

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How to parameterize a_p, w_p, X_p?

Plume averaged equations

Start from unaveraged (Boussinesq) tracer equation

$$\partial_t X + \nabla \cdot \boldsymbol{u} X = S_X$$

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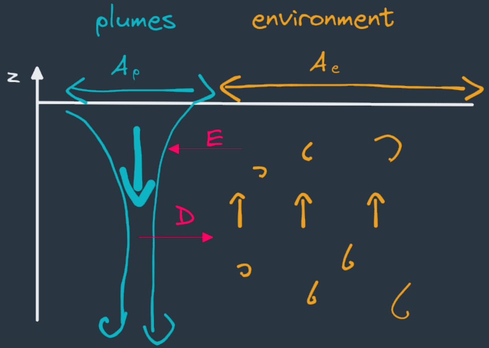
Average over plume area A_p+ upstream approximation $X_{\mathrm{bdary}} = \begin{cases} X_p & \text{if } u_{\mathrm{bdary}} > 0 \\ X_e & \text{if } u_{\mathrm{bdary}} < 0 \end{cases}$

$$\partial_t a_p X_p + \partial_z (a_p w_p X_p + a_p \overline{w_p' X_p'}) = E X_e - D X_p + a_p \overline{S_X}^p$$



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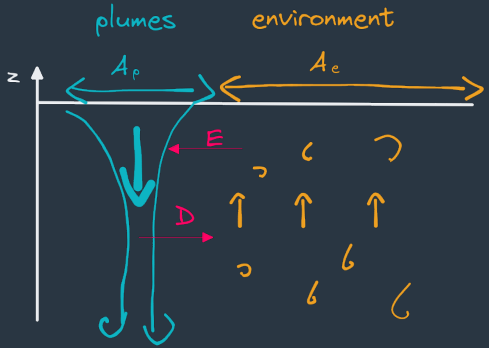

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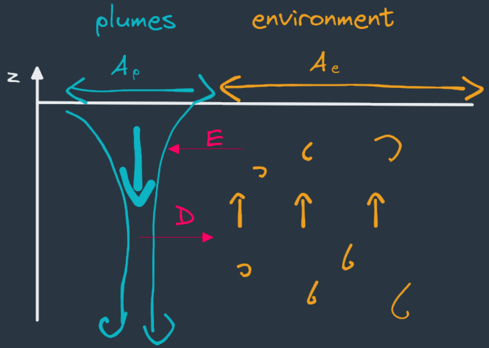

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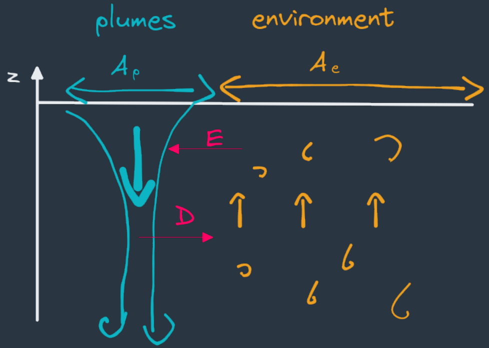
$$\partial_t a_p X_p + \partial_z (a_p w_p X_p + a_p \overline{w_p' X_p'}) = E X_e - D X_p + a_p \overline{S_X}^p$$



- $-a_p w_p$: mass-flux (> 0)
- E: entrainment (> 0)
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Plume averaged equations

Start from unaveraged (Boussinesq) tracer equation

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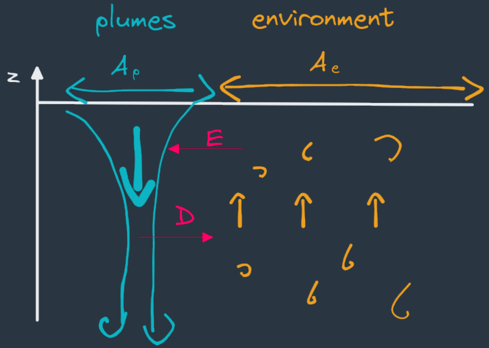

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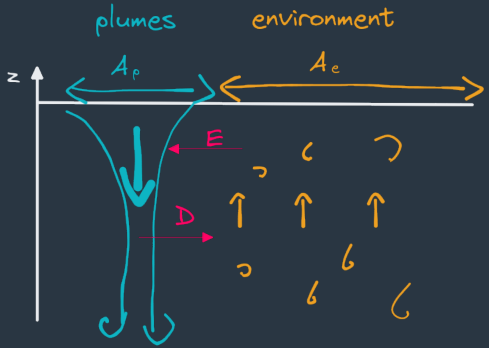

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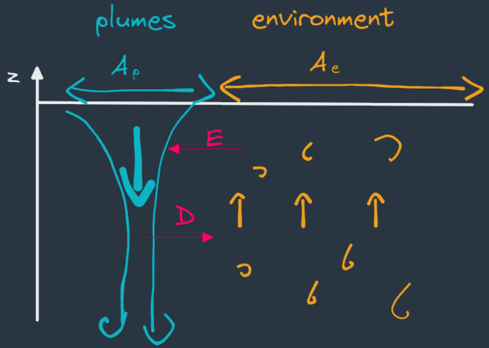

- $-a_p w_p$: mass-flux (> 0)
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- D: detrainment (> 0)
- Area conservation:

$$\partial_t a_p + \partial_z (a_p w_p) = E - D$$

Steady-plume hypothesis

$$\frac{\partial_t(a_p X_p)}{\partial_z(a_p w_p X_p)} \ll 1$$

$$\text{order of magn. } \frac{\mathrm{d}h/\mathrm{d}t}{W} \ll 1$$

ex: for idealized free convection ($W = w_*$) or pure shear ($W = u_*$)

$$\frac{\mathrm{d}h/\mathrm{d}t}{W} \propto (N_0 t)^{-1/3}$$

Small area limit

Standard (implicit) assumptions:

- small plume area: $a_e \Longrightarrow 1,\ a_p \Longrightarrow 0$

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Summary:

$$\begin{aligned}
\partial_t \overline{X} &= -\partial_z \overline{w' X'} + \overline{S}_X \\
\overline{w' X'} &= a_p w_p (X_p - \overline{X}) + \overline{w_e' X_e'} \\
a_p w_p \partial_z X_p &= -E(X_p - \overline{X}) + a_p S_{X,p}
\end{aligned}$$

What has to be closed/paremeterized?

Closure

- Eddy-diffusivity: local turbulence in environment

$$\overline{w'_e X'_e} \overset{\text{param}}{=} -K_X \partial_z \overline{X}$$

$+$ TKE closure for K_X

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- Entrainment/detrainment: several propositions, for example:

[Gregory 2001, Rio et al. 2010]

$$\begin{cases} E &= a_p \beta_1 \max(0, \partial_z w_p) \\ D &= -a_p \beta_2 \min(0, \partial_z w_p) - a_p w_p \delta_0 \end{cases}$$

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- Pressure gradient

$$a_p S_{w,p} = a_p \left(\frac{1}{\rho_0} \partial_z p^\dagger \right)_p = (a-1) a_p B_p + (b-1)(-E w_p) + b' a_p w_p^2$$

$$a_p S_{\boldsymbol{u}_h,p} = a_p \left(\frac{1}{\rho_0} \nabla_h p^\dagger \right)_p = a_p w_p C_u \partial_z \overline{\boldsymbol{u}}_h$$

a, b, b', C_u are parameters

Putting it all together

Resolved fields:

$$\partial_t \overline{X} = -\partial_z (\underbrace{-K_X \partial_z \overline{X}}_{\text{ED}} + \underbrace{a_p w_p (X_p - \overline{X})}_{\text{MF}}) + \overline{S_X}$$

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Plume equations:

$$\partial_z (a_p w_p) = E - D \qquad \textit{(continuity)}$$

$$a_p w_p \partial_z \phi_p = -E(\phi_p - \overline{\phi}) + \overline{S_{\phi,p}} \qquad \textit{(salinity, temperature)}$$

$$a_p w_p \partial_z \boldsymbol{u}_{p,h} = -E(\boldsymbol{u}_{p,h} - \overline{\boldsymbol{u}}_h) - a_p w_p C_u \partial_z \overline{\boldsymbol{u}}_h \qquad \textit{(hor. velocities)}$$

$$a_p w_p \partial_z w_p = a a_p B_p - b E w_p + b' a_p w_p^2$$

$+$ boundary conditions

$+$ parameters: $\boldsymbol{C} = (\beta, a, b, b', C_u, C_v, a_p^0)$

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$+$ TKE equation! to close K_X

Energy budgets

Motivation

Caricatural pratice in modelling:

`Use TKE scheme developped by X`

`Use mass-flux  scheme developped by Y`

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Caricatural pratice in modelling:

```
Use TKE scheme developped by X
Use mass-flux  scheme developped by Y
```

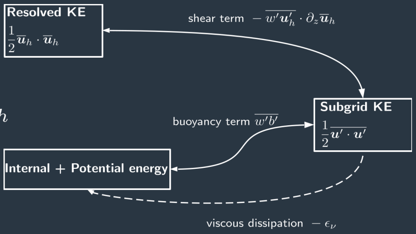
But mass-flux leads to energy transfers, so it should be carefully taken into account into TKE!

Resolved internal and potential energy reads

$$E_i + E_p = z(g - \overline{b}) + \overline{h(p_0, \phi)} - \frac{p_0}{\rho_0}$$

KE, TKE and IE+PE budget

$$\begin{cases} \partial_t E_k + \partial_z T_{E_k} = & \overline{w' \boldsymbol{u}_h'} \cdot \partial_z \overline{\boldsymbol{u}}_h \\ \partial_t k + \partial_z T_k = & -\overline{w' \boldsymbol{u}_h'} \cdot \partial_z \overline{\boldsymbol{u}}_h \\ & + \overline{w'b'} - \overline{\epsilon}_\nu \\ \partial_t (E_i + E_p) + \partial_z T_h = & -\overline{w'b'} + \overline{\epsilon}_\nu \end{cases}$$



Standard ocean model approximations

$E_i + E_p$ for seawater (and linear EOS)

$$\partial_t \left(c_p \overline{\theta} - z\overline{b} \right) = \overline{\epsilon}_\nu - \partial_z \left(c_p \overline{w'\theta'} - z\overline{w'b'} \right) - \overline{w'b'} \tag{1}$$

[McDougall 2003]: $\overline{\epsilon}_\nu \longrightarrow$ heating of $10^{-3}K$ per 100 years...
neglecting $\overline{\epsilon}_\nu \implies$ decoupling of $\overline{\theta}/\overline{B}$ equations

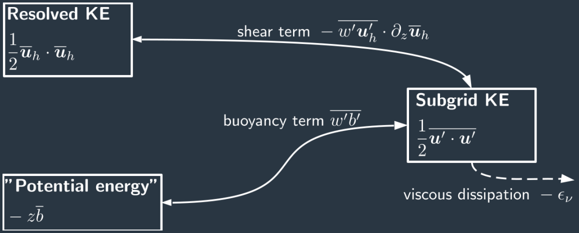
$$\partial_t \overline{\theta} = -\partial_z \overline{w'\theta'} \tag{2}$$

$$\partial_t(-z\overline{b}) = -\partial_z(-z\overline{w'b'}) - \overline{w'b'} \tag{3}$$

Ocean SCM energy budget

Oceanic "Potential" Energy: $\tilde{E}_p = -z\overline{b}$

$$\begin{cases} \partial_t E_k + \partial_z T_{E_k} & = \overline{w' \boldsymbol{u}'_h} \cdot \partial_z \overline{\boldsymbol{u}}_h \\ \partial_t k + \partial_z T_k & = -\overline{w' \boldsymbol{u}'_h} \cdot \partial_z \overline{\boldsymbol{u}}_h + \overline{w'b'} - \overline{\epsilon}_\nu \\ \partial_t \tilde{E}_p & = -\partial_z (-z\overline{w'b'}) - \overline{w'b'} \end{cases}$$



An consistent TKE equation

To close the resolved+unresolved energy budget, we need:

$$\partial_t k + \partial_z T_k = - \underbrace{\left(-K_u(\partial_z \overline{\boldsymbol{u}}_h) + a_p w_p(\boldsymbol{u}_{h,p} - \overline{\boldsymbol{u}}_h)\right) \cdot \partial_z \overline{\boldsymbol{u}}_h}_{\text{shear term}}$$

$$\underbrace{-K_b \partial_z \overline{b} + a_p w_p (b_p - \overline{b})}_{\text{buoyancy term}}$$

$$- \epsilon_\nu$$

$$\overline{T_k} \underset{\mathrm{EDMF}}{=} \underbrace{-K_k \partial_z k}_{ED} + \underbrace{a_p w_p \left(k_p - k + \frac{1}{2} \|\boldsymbol{u}_p - \overline{\boldsymbol{u}}\|^2 \right)}_{MF}$$

with plume inner TKE $k_p = 1/2\overline{\boldsymbol{u}_p' \cdot \boldsymbol{u}_p'}$

Plume TKE

Main equilibrium

$$\begin{aligned}
\partial_t(a_p k_p) + \partial_z(a_p w_p k_p) \quad = \quad & -a_p \overline{w_p' \boldsymbol{u}_{h,p}'} \cdot \partial_z \boldsymbol{u}_{h,p} + a_p \overline{w_p' b_p'} \\
& + E\left(k_e + \frac{1}{2}\|\boldsymbol{u}_e - \boldsymbol{u}_p\|^2\right) - D k_p \\
& - \partial_z \left(a_p \overline{w_p' \frac{\boldsymbol{u}_p' \cdot \boldsymbol{u}_p'}{2}} + a_p \overline{\boldsymbol{u}_p' \cdot \frac{1}{\rho_0} (\nabla p^\dagger)_p'} \right) \\
& - a_p (\epsilon_\nu)_p
\end{aligned}$$

Plume TKE

Full equation

$$\begin{aligned}\partial_t(a_p k_p) + \partial_z(a_p w_p k_p) \quad = \quad & -a_p \overline{w_p' \boldsymbol{u}_{h,p}'} \cdot \partial_z \boldsymbol{u}_{h,p} + a_p \overline{w_p' b_p'} \\ & + E\left(k_e + \frac{1}{2}\|\boldsymbol{u}_e - \boldsymbol{u}_p\|^2\right) - D k_p \\ & -\partial_z\left(a_p \overline{w_p' \frac{\boldsymbol{u}_p' \cdot \boldsymbol{u}_p'}{2}} + a_p \overline{\boldsymbol{u}_p' \cdot \frac{1}{\rho_0}(\nabla p^\dagger)_p'}\right) \\ & -a_p(\epsilon_\nu)_p\end{aligned}$$

Plume TKE

Full equation

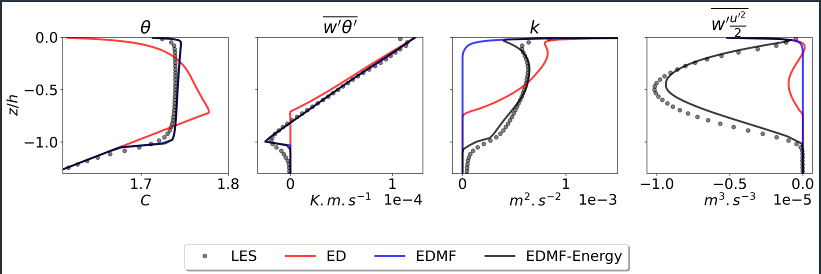
$$\begin{aligned}\partial_t(a_p k_p) + \partial_z(a_p w_p k_p) &= -a_p\overline{w_p'\boldsymbol{u}_{h,p}'}\cdot\partial_z\boldsymbol{u}_{h,p} + a_p\overline{w_p'b_p'} \\ &+E\left(k_e + \frac{1}{2}\|\boldsymbol{u}_e - \boldsymbol{u}_p\|^2\right) - Dk_p \\ &-\partial_z\left(a_p\overline{w_p'\frac{\boldsymbol{u}_p'\cdot\boldsymbol{u}_p'}{2}} + a_p\overline{\boldsymbol{u}_p'\cdot\frac{1}{\rho_0}(\nabla p^\dagger)_p'}\right) \\ &-a_p(\epsilon_\nu)_p\end{aligned}$$

Small area $a_p \ll 1$

$$a_p w_p \partial_z k_p = -E\left(k_p - k\right) + E\left(\frac{1}{2}(\boldsymbol{u}_p - \overline{\boldsymbol{u}})^2\right) - a_p(\epsilon_\nu)_p$$

Free convection into linear stratification

Surface cooling $Q_0 = -500\,\mathrm{W\,m^{-2}}$, initial stratification $1\mathrm{K}/1000\mathrm{m}$



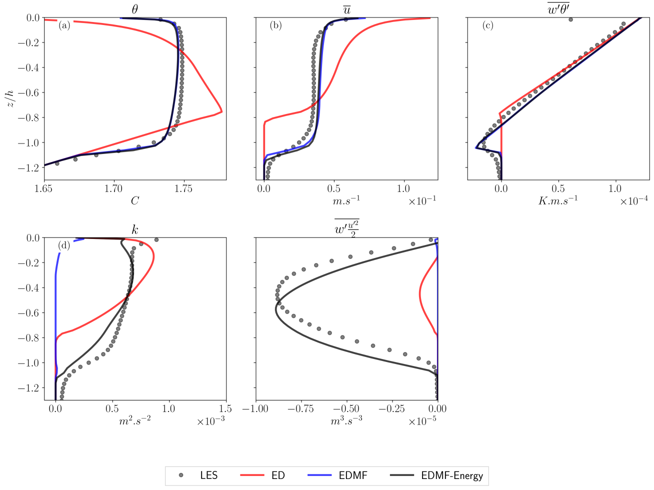
LES= high resolution reference, ED = Eddy-Diffusivity only,

EDMF = EDMF + Standard TKE equation, EDMF-Energy = EDMF + new TKE equation

- good representation of turbulent statistics
- not really improving $\overline{\theta}$ or $\overline{w'\theta'}$ because ED is small?

Surface Cooling and Wind stress

Surface cooling $Q_0 = -500\,\mathrm{W\,m^{-2}}$, wind stress $u_*^2 = 0.05\,\mathrm{m^2\,s^{-2}}$, initial stratification $1\mathrm{K}/1000\mathrm{m}$



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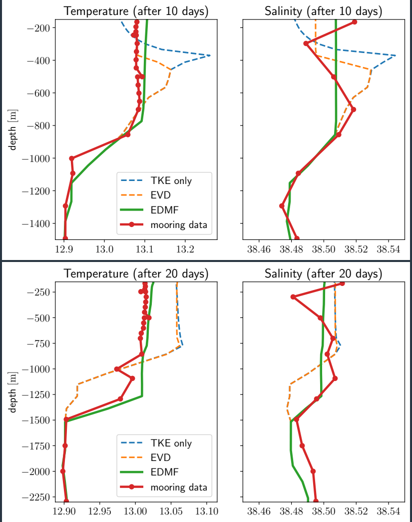
ASICS-MED Campaign

Real data during violent events of convection: January-March 2013

Data provided by Hervé Giordani, and corrected fluxes by *[Caniaux et al. 2017]*

ASICS-MED Campaign

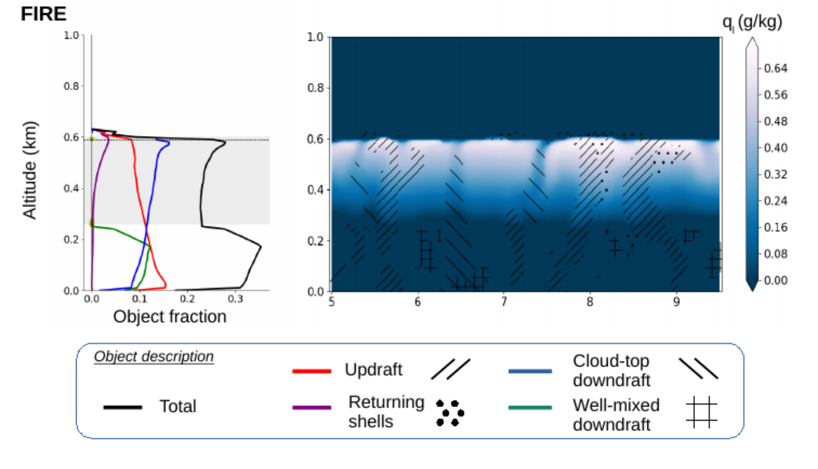


Summary:

- eddy-diffusivity not satisfied for convection
- non-local coherent structures responsible for mixing
- multifluid approach (EDMF) provide a satifactory modelling and key advantages:
    - formal derivation from first principles + hypothesis
    - flexible to incorporate other processes
    - derive energy budgets
    - scale-awareness

Perspectives

- incorporate more coherent structures (downdrafts, returning shells...) *[Brient et al 2023]*
-



Perspectives

- illustrate energetic consistency with atmospheric cases
- incorporate more coherent structures (downdrafts, returning shells...) *[Brient et al 2023]*
- relax $a_p \ll 1$ *[Honnert et al. 2016]*
- relax steady plume hypothesis *[Tan et al. 2018, Thuburn et al. 2018]*
- unified description of boundary layer convection and deep convection *[Suselj et al 2019, Hourdin et al 2019]*
- uncertainty quantification, sensitivity and tuning *[Souza et al 2020, Hourdin et al 2021]*

Coriolis effects during convection

Plume dynamics

Turnover (=lifetime) time: $\tau = \frac{h}{w_*}$ vs rotation time $\frac{1}{f}$ Natural Rossby number

$$Ro_* = \left(\frac{1/f}{\tau}\right)^{3/2} = \frac{l_{\mathrm{rot}}}{h} = \left(\frac{B_0}{f^3 h^2}\right)^{1/2}$$

$l_{\mathrm{rot}} = (B_0/f^3)^{1/2}$: scale at which convection is affected by rotation

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$l_{\text{rot}} = (B_0/f^3)^{1/2}$: scale at which convection is affected by rotation For a given heat flux: $\frac{B_0^{\text{atmosphere}}}{B_0^{\text{ocean}}} \simeq 10^5$

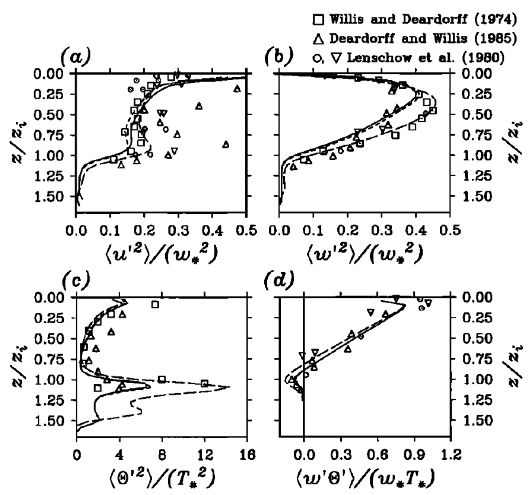
Oceanic Deep Convection:

$Ro_* \simeq 0.01 - 1, l_{\text{rot}} = 500 - 1000\text{m}, \tau = 10^4 - 10^5\text{s}$

Atmospheric Convection:

$Ro_* \simeq 10 - 100, l_{\text{rot}} = 10^5 - 10^6\text{m}, \tau = 10 - 30\text{min}$

Plume dynamics



Denbo & Skyllingstad JGR 1996

plain: full Coriolis
short dashed: trad. Coriolis

long dashed: no Coriolis

⟶ plumes are less energetics and less penetrative!

⟶ how to parameterize this effect?

Non-traditional Coriolis

Full Coriolis

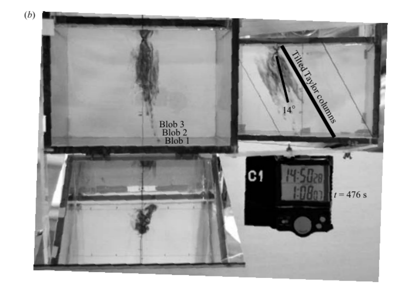
$$\boldsymbol{f} = 2\Omega \begin{pmatrix} 0 \\ \cos\theta \\ \sin\theta \end{pmatrix} = \begin{pmatrix} 0 \\ f^{\perp} \\ f \end{pmatrix}$$

$$\begin{aligned} \frac{D}{Dt}u + f^{\perp}w - fv &= -\frac{1}{\rho_0}\partial_x p \\ \frac{D}{Dt}v + fu &= -\frac{1}{\rho_0}\partial_y p \\ \frac{D}{Dt}w - f^{\perp}u &= -\frac{1}{\rho_0}\partial_z p + b \end{aligned}$$

Convection is non-hydrostatic ($O(u/w) = 1$) so non-traditional terms should be retained

Plume scale effects

[Straneo 2002, Sheremet 2004, Wirth&Barnier 2006]: point source-plumes develop in the direction between the acceleration due to gravity and rotation vectors and shifts eastward



Sheremet JFM 2004

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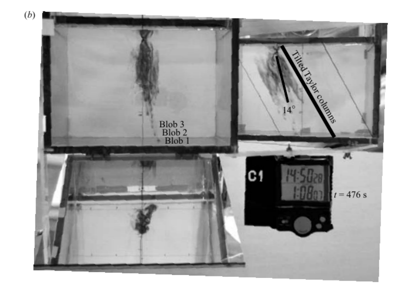


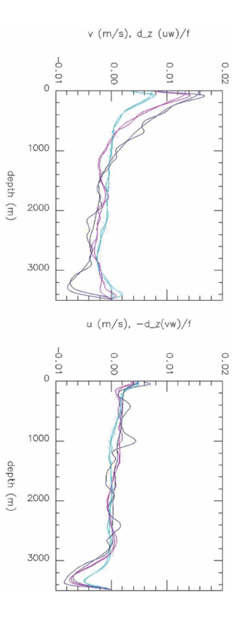
Sheremet JFM 2004

Simple model: buoyant parcel into lighter environment $b_{\text{parcel}} < b_{\text{env}}$

$$\frac{\mathrm{d}}{\mathrm{d}t}\boldsymbol{U}(t) + \boldsymbol{f} \times \boldsymbol{U} = (b_{\text{parcel}} - b_{\text{env}})\boldsymbol{e}_z$$

Analogous of charged particle in an electric and magnetic field!

Mean effects



[Wirth&Barnier 2008]: Homogeneous buoyancy loss B_0 with full rotation

Tilted convection $\implies$ assymetric pertubations $\overline{w'u'} \neq 0 \ \overline{w'v'} \neq 0$

On average, creates a mean circulation of $O(1\,\mathrm{cm\,s^{-1}})$

$$\overline{v}^{x,y,t} = \frac{1}{f}\partial_z \overline{w'u'}^{x,y,t}$$

$$\overline{u}^{x,y,t} = -\frac{1}{f}\partial_z \overline{w'v'}^{x,y,t}$$

Add full Coriolis to EDMF plume equations?

Convection vs. baroclinic instability

With horizontal buoyancy gradients, submesocale baroclinic instability at scales $O(1\,\text{km})$ can occur $\implies$ restratification by baroclinic eddies

vs

Convective plumes $O(1\,\text{km})$ overturn $\implies$ destratification

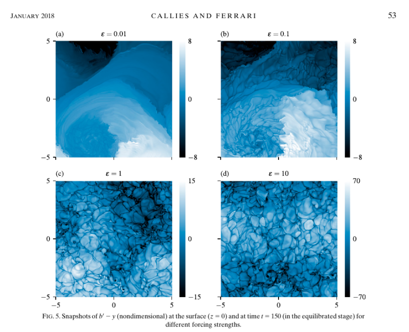
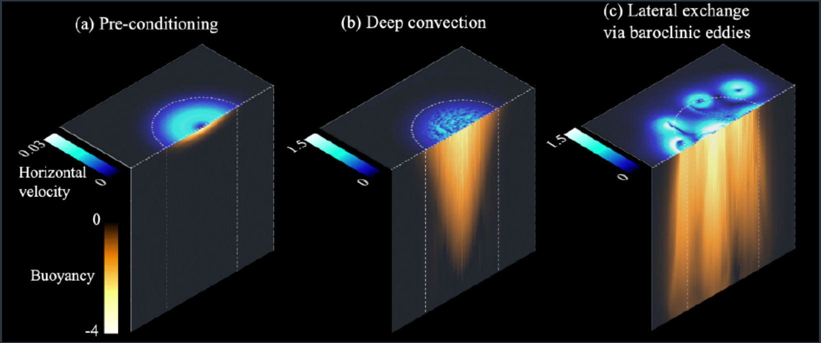


FIG. 5. Snapshots of $b' - y$ (nondimensional) at the surface ($z = 0$) and at time $t = 150$ (in the equilibrated stage) for different forcing strengths.

$$\epsilon = \frac{\text{destabilizingflux}}{\text{baroclinicflux}} = \frac{B_0}{(\partial_y b_0 h)^2/f}$$

Convective patch



Vreugdenhil&Bayen 2020, adapted from Sohail et al. 2020

Inertial Oscillations in a Zero Order Model

Wind storm blowing over the ocean

Start again from horizontal momentum budget

$$\partial_t \boldsymbol{u} + \boldsymbol{f} \times \boldsymbol{u} = -\partial_z \overline{w' \boldsymbol{u}'}$$

Integrating over the mixed layer: $\boldsymbol{U}_m := \frac{1}{h} \int_{-h^-}^{0} \boldsymbol{u} \, \mathrm{d}z$

$$h \partial_t \boldsymbol{U}_m + \boldsymbol{f} \times (h \boldsymbol{U}_m) = \overline{w' \boldsymbol{u}'}_0 - \overline{w' \boldsymbol{u}'}_h$$

Integrating over the entrainment jump:

$$\overline{w' \boldsymbol{u}'}_h = -\Delta \boldsymbol{u} \partial_t h$$

If fluid is at rest in the stratified zone $\Delta \boldsymbol{u} = 0 - \boldsymbol{U}_m$:

$$\partial_t (h \boldsymbol{U}_m) + \boldsymbol{f} \times (h \boldsymbol{U}_m) = \overbrace{\overline{w' \boldsymbol{u}'}_0}^{u_*^2}$$

Solution in complex notation: $\underline{U}_m(t) = \frac{1}{h(t)} \frac{u_*^2}{if} (1 - e^{-ift})$

Inertial Oscillations

$\underline{U}_m(t) = \frac{1}{h(t)} \frac{u_*^2}{if}(1 - e^{-ift})$

Oscillations on the time scale $1/f \simeq 10\text{hours}$ at midlatitude

Dominant energetic mode of internal waves

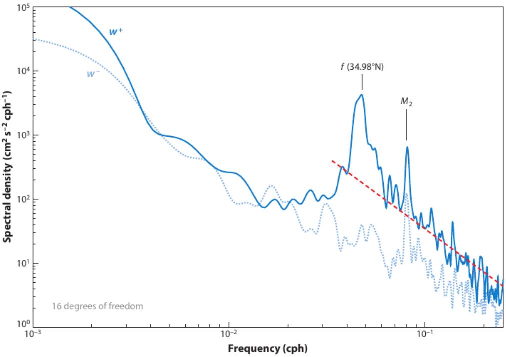


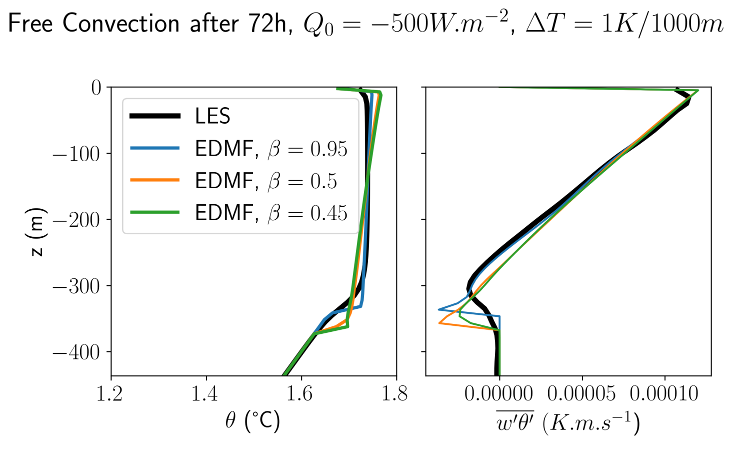
Figure 1

Rotary velocity spectrum at 261-m depth from current-meter data from the WHOI699 mooring gathered during the WESTPAC1 experiment (mooring at 6,149-m depth.) The solid blue line (w^+) is clockwise motion, and the dashed blue line (w^-) is counterclockwise motion; the differences between these emphasize the downward energy propagation that often dominates the near-inertial band. The dashed red line is the line $E_0 N \omega^{-p}$ with $N = 2.0$ cycles per hour (cph), $E_0 = 0.096$ cm^2 s^{-2} cph^{-2}, and $p = 2.25$, which is quantitatively similar to levels in the Cartesian spectra presented by Fu (1981) for station 5 of the Polygon Mid-Ocean Experiment (POLYMODE) II array.

Alford et al Ann Rev Mar Sci 2016

Parameter estimation and sensitivity

Sensitivity to parameters



Sensitivity analysis: method

- 7 parameters: $\boldsymbol{C} = (\beta, a, b, b', C_u, C_v, a_p^0)$

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- decompose variance into each parameter contribution *[Sobol 2001]*

$$\mathbf{Var}\left[\theta(\boldsymbol{C})\right] = \mathbf{Var}\left[\mathbf{E}\left[\theta(\boldsymbol{C})|C_1\right]\right] + \ldots + \mathbf{Var}\left[\mathbf{E}\left[\theta(\boldsymbol{C})|C_7\right]\right]$$

$+$ higher order terms

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$$+ \text{ higher order terms}$$

- two variances:

$$\mathbf{Var}_{z,t}\left[\theta(z, t; \boldsymbol{C})\right] = \mathbf{E}\left[(\theta(z, t; \boldsymbol{C}) - \mathbf{E}\theta(z, t; \boldsymbol{C}))^2\right]$$
$$\mathbf{Var}_{L^2}\left[\theta(\boldsymbol{C})\right] = \mathbf{E}\left[\int_0^T \int_{-H}^0 (\theta(z, t; \boldsymbol{C}) - \mathbf{E}\theta(z, t; \boldsymbol{C}))^2 \,\mathrm{d}z\mathrm{d}t\right]$$

Sensitivity analysis: results

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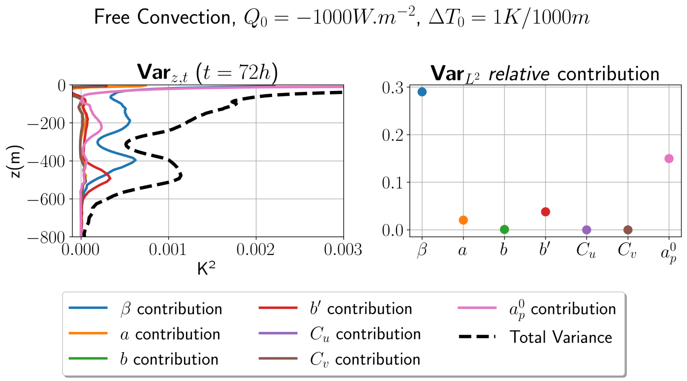
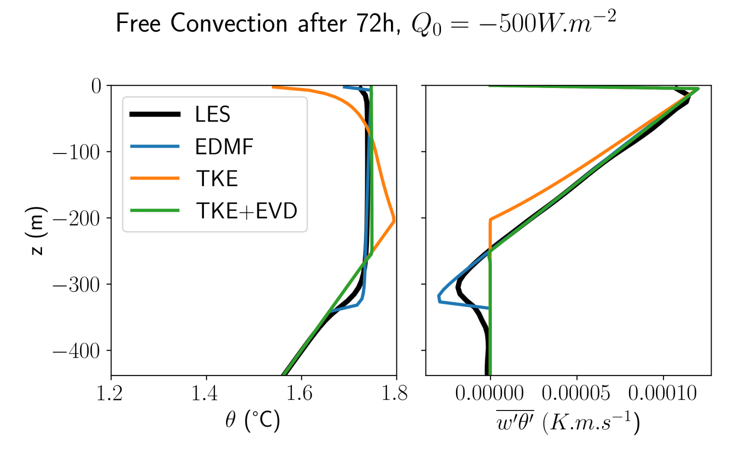


$\longrightarrow$ entrainment coefficient β and plume surface area a_p^0 are the most critical parameters!

Comparison with TKE and EVD schemes



- TKE alone does not mix enough

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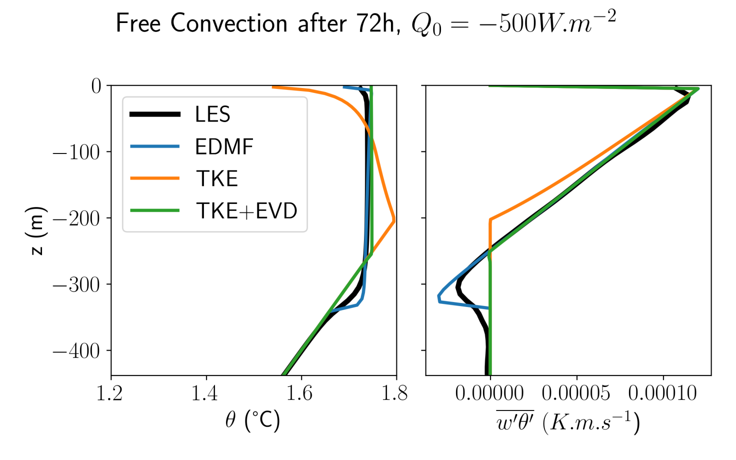

- TKE alone does not mix enough
- TKE+EVD is not penetrative (not true with wind)

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- couple to parameterization of restratification by submesocale eddies