INTERNAL WAVES

4. PROPAGATION AND DISSIPATION

- **Lesson 1 :**
  - Introduction about Ocean waves
  - Surface waves
  - Internal Waves (Introduction)
- **Lesson 2 :**
  - Internal Waves in the 2-layer model
  - Internal Waves with a continuous stratification (equations)
- **Lesson 3 :**
  - Internal Waves with a continuous stratification (solutions)
- **Lesson 4 :**
  - Generation of internal waves
  - Activities (home): Numerical simulation of internal waves
- **Lesson 5 :**
  - Propagation, dissipation and interaction of internal waves

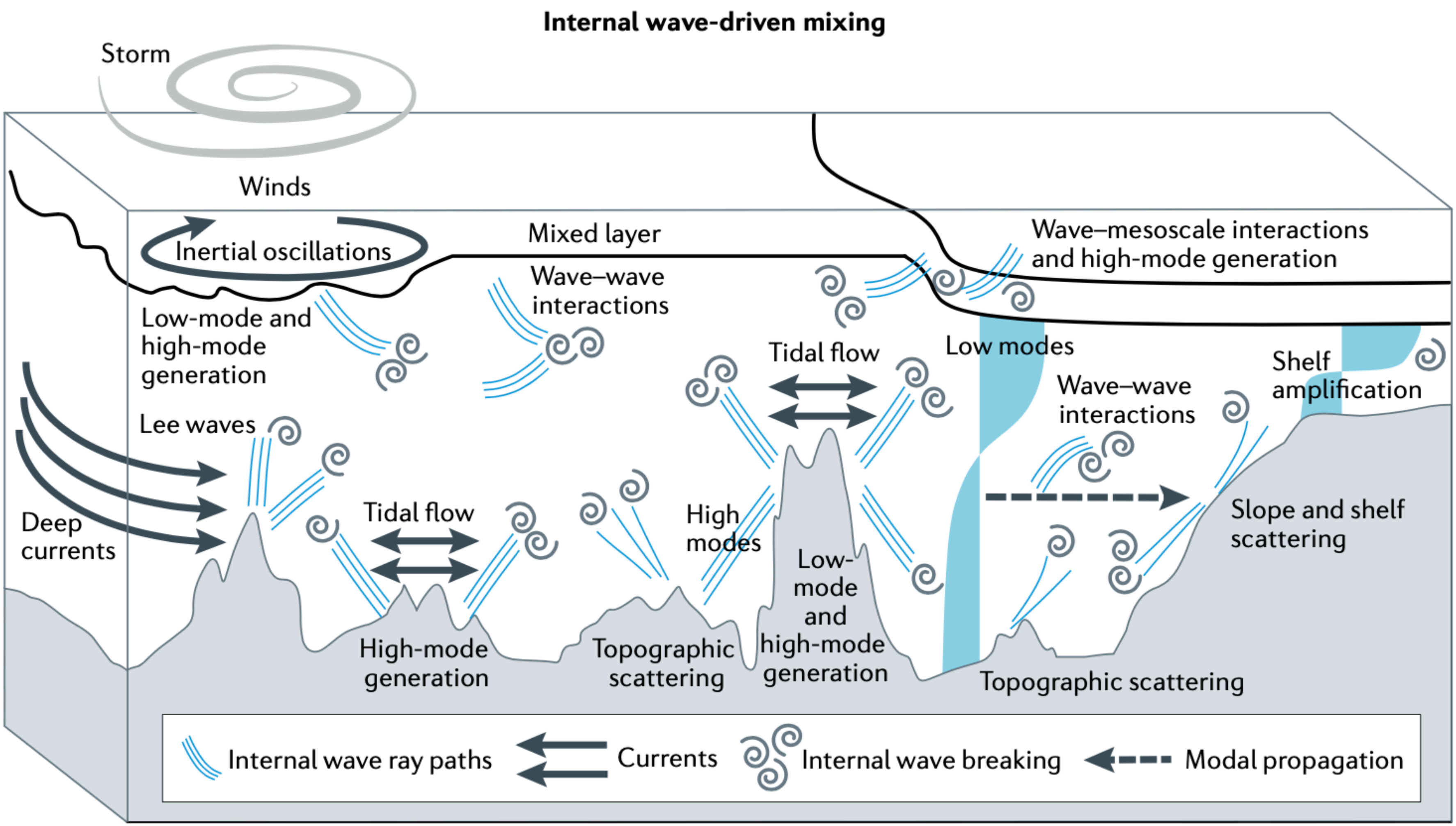
Presentations and material will be available at :

jgula.fr/Ondes/

- **1.4 : Propagation and dissipation of internal waves**

- *Waves in the Ocean*, LeBlond & Misak
- *The excitation, dissipation, and interaction of internal waves in the deep ocean,* Thorpe, 1975
- *Internal waves in the ocean*, Garrett & Munk, 1979
- *Internal Waves and Small-Scale Processes,* Munk, 1981

Internal wave generation (last lesson)



Whalen et al 2020

1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

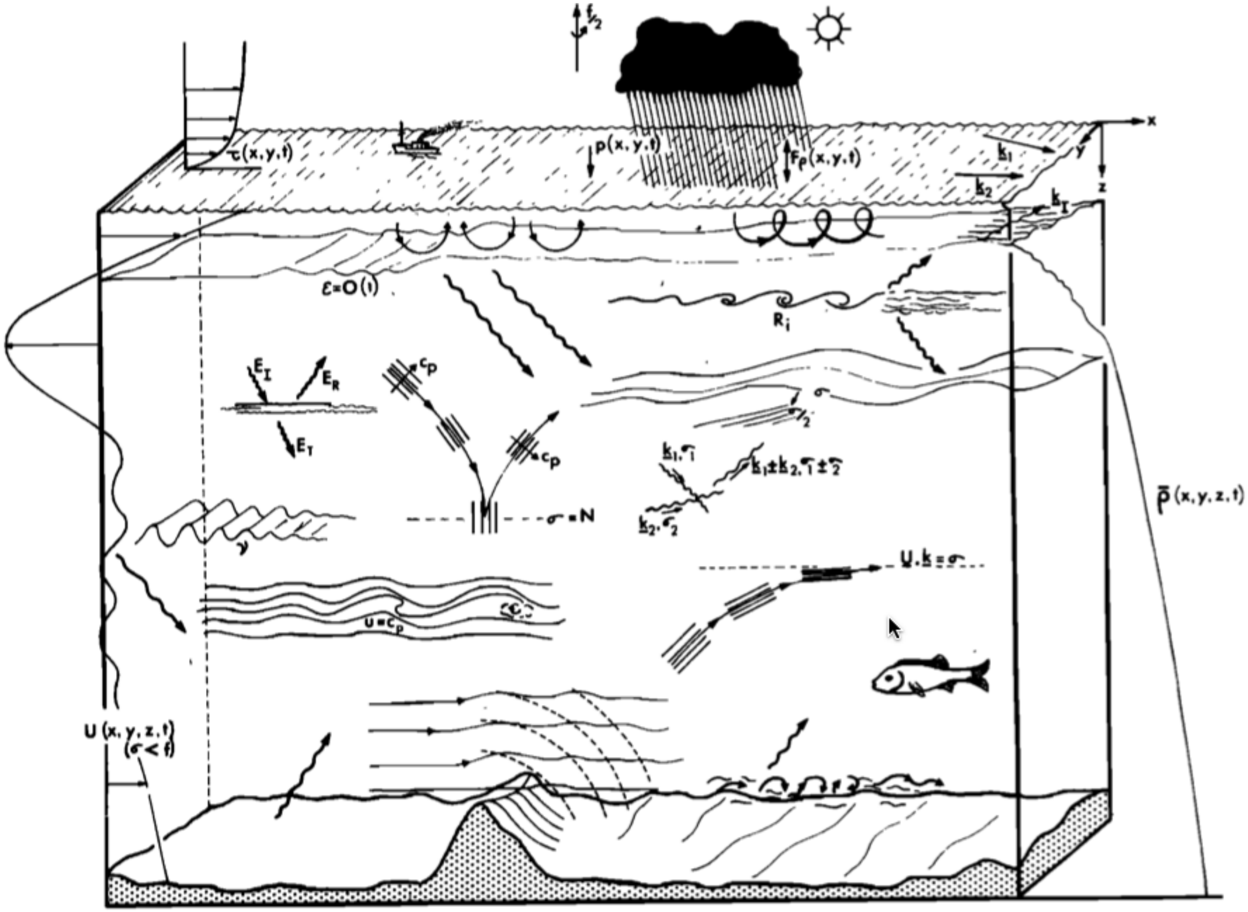


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

Propagation: notions of ray-tracing

- Let $\overrightarrow{U}(\overrightarrow{x}, t)$ be the background current and consider a wave with an intrinsic frequency ω_0 and wavenumber k. ("Intrinsic" means that the property is measured in a frame moving with the fluid)

- The frequency measured in a fixed frame (*doppler-shifted frequency*) is $\omega = \omega_0 + \overrightarrow{k} \cdot \overrightarrow{U}$ (modified dispersion relation)

- (It can be written $\omega = \overrightarrow{k} \cdot (\overrightarrow{c_0} + \overrightarrow{U})$ with $\overrightarrow{c_0}$ the intrinsic group velocity of the wave)

https://glossary.ametsoc.org/wiki/Intrinsic_wave_frequency

1.4. Propagation and dissipation of waves

Propagation: notions of ray-tracing

Ray tracing is a method coming from geometrical optics where the path of the waves is modelled by assuming that the medium is “slowly varying”, spatially and temporally, compared to the wavelength and period of the waves (*).

General dispersion relation :

$$\omega(\overrightarrow{x}, t) = \Omega_0[\overrightarrow{k}(\overrightarrow{x}, t), \lambda(\overrightarrow{x}, t)] + \overrightarrow{k} \cdot \overrightarrow{U}$$

where Ω_0 is the *intrinsic* dispersion relation and $\lambda = N(\overrightarrow{x}, t) + f(\overrightarrow{x})$ for internal waves and $\overrightarrow{U}$ is the background velocity.

(*) *This corresponds to the WKB (Wentzel–Kramers–Brillouin) approximation.*

1.4. Propagation and dissipation of waves

Propagation: notions of ray-tracing

The ray paths (corresponding to the wave energy), as well as the variations of the wavenumber k and the frequency ω along the ray path are then given by:

$$\begin{aligned}
\frac{dx_i}{dt} &= \frac{\partial \Omega_0}{\partial k_i} + U_i \\
\frac{dk_i}{dt} &= -\frac{\partial \Omega_0}{\partial \lambda}\frac{\partial \lambda}{\partial x_i} - k_j \frac{\partial U_j}{\partial x_i} \\
\frac{d\omega}{dt} &= \frac{\partial \Omega_0}{\partial \lambda}\frac{\partial \lambda}{\partial t} + k_j \frac{\partial U_j}{\partial t}
\end{aligned}$$

for x_i in x, y, z

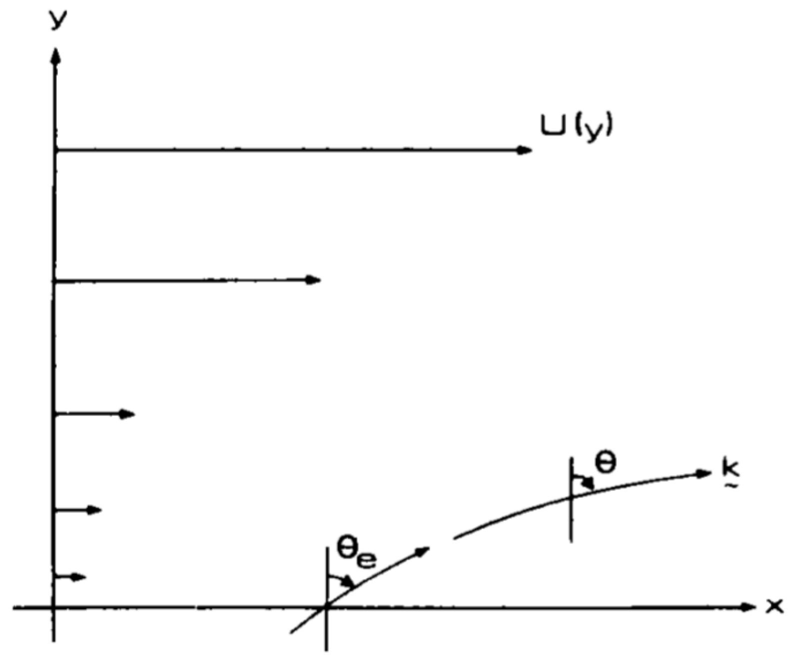
1.4. Propagation and dissipation of waves

Propagation: notions of ray-tracing

Example : we consider a long gravity wave in a horizontal plane (x, y) :

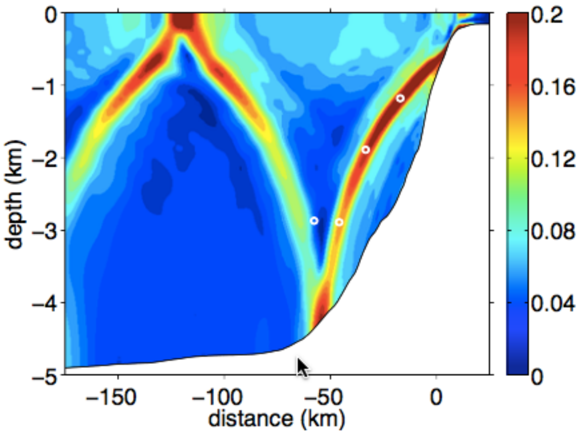
- The wave frequency is $\Omega_0(k_x, k_y, H) = k\sqrt{gH}$ (intrinsic dispersion relation) with H the water depth

- Now consider a horizontal background flow $U(y)$ and assume H is constant and $U_e = 0$ ("$_e$" for "entry" in the current shear)



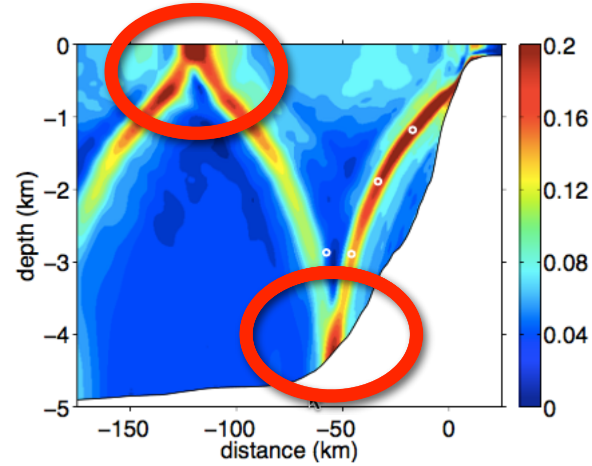
1. Derive the equations for the ray path
2. Get the propagation angle θ, as a function of θ_e, $U(y)$ and $\sqrt{gH}$

1.4. Propagation and dissipation of waves



What happens after internal wave generation?

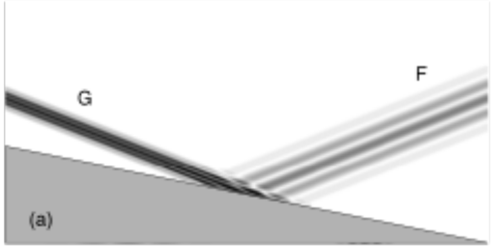
1.4. Propagation and dissipation of waves



What happens after internal wave generation?

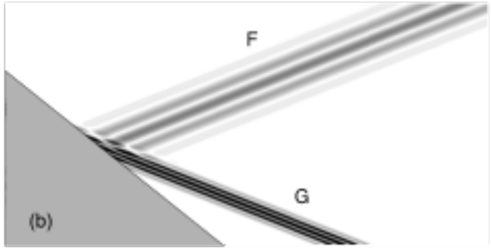
A. Wave reflection at the surface and bottom.

In reflection from a rigid boundary, frequency is conserved, and the waves retain their inclination to the horizontal



$$\tan\theta = \pm\left(\frac{\omega^2 - f^2}{N^2 - \omega^2}\right)^{1/2}$$

(See LeBlond & Mysak, p 54)



1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

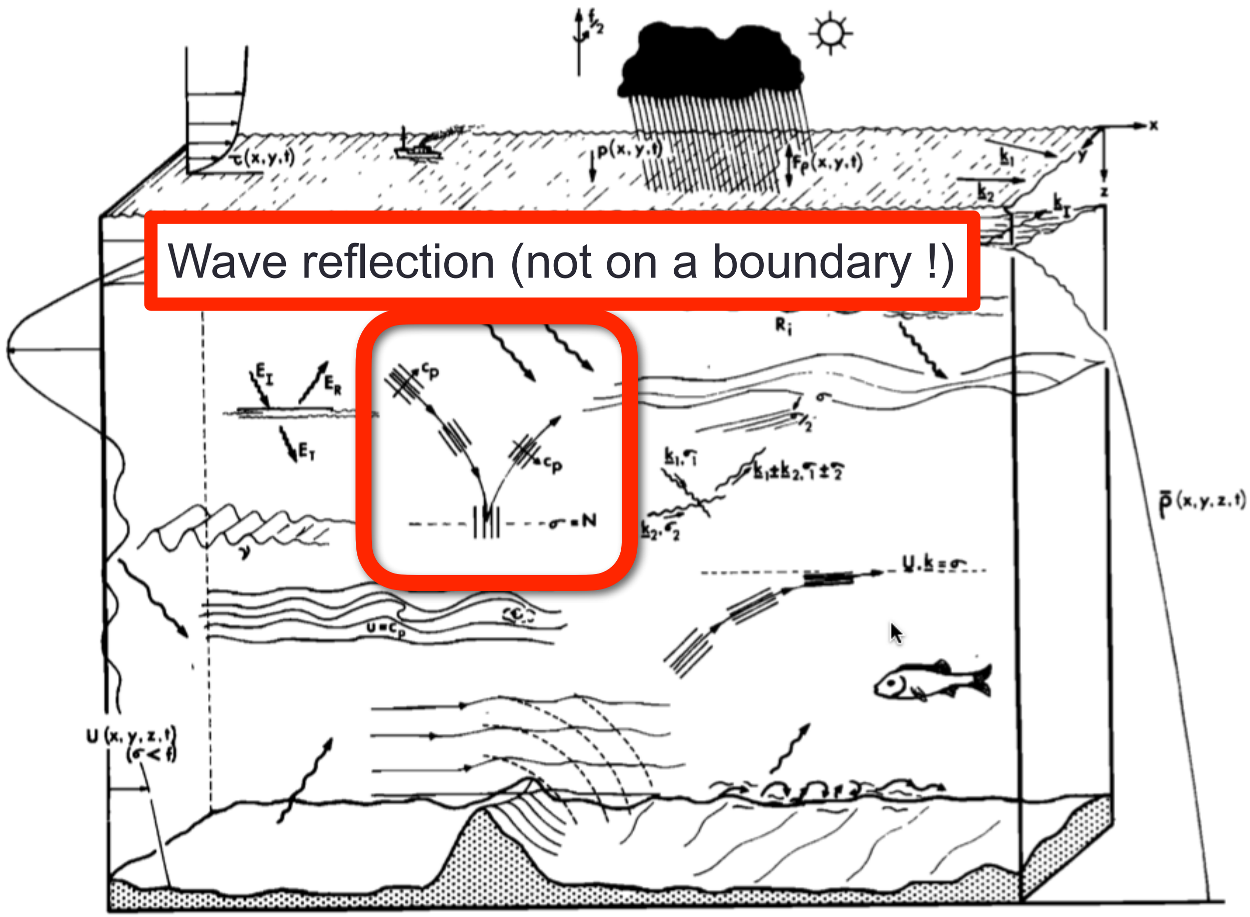


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

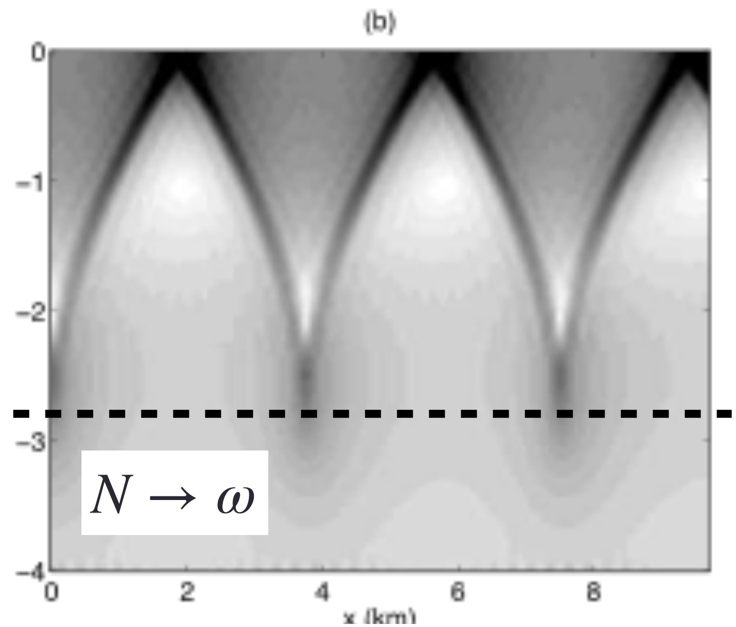
B. Wave reflection in the interior

Remember the dispersion relation in the continuously stratified case :

$$\omega^2 = \frac{N^2k^2 + f^2m^2}{k^2 + m^2}$$

That is valid is the cases :

$N \leq \omega \leq |f|$ or $|f| \leq \omega \leq N$



If a wave propagates into a region where $N \to \omega$ (frequency of the waves), it is reflected and its group velocity becomes vertical near this surface.

Note: they may be trapped into a waveguide if this happens below and above.

1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

Wave reflection AND refraction (not on a boundary !)

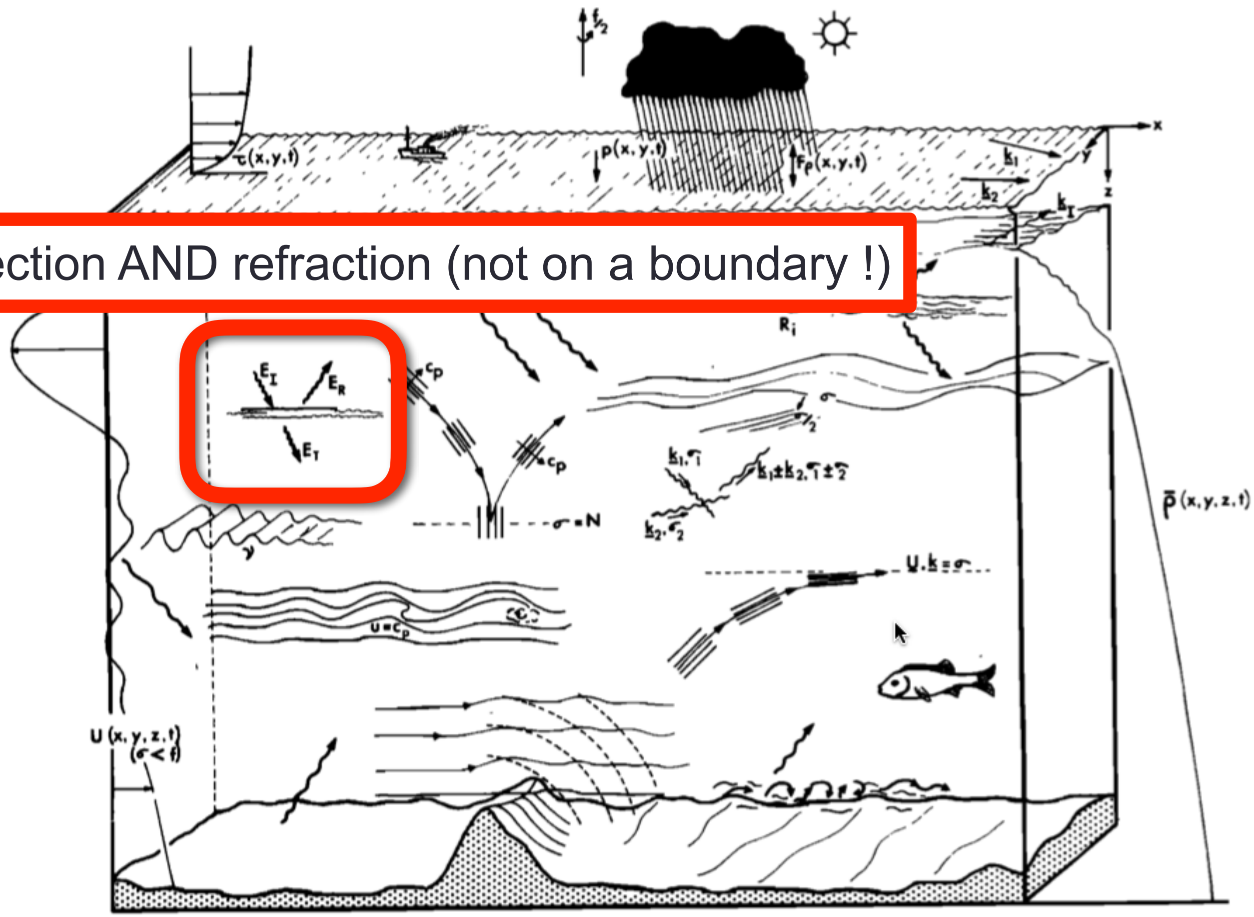
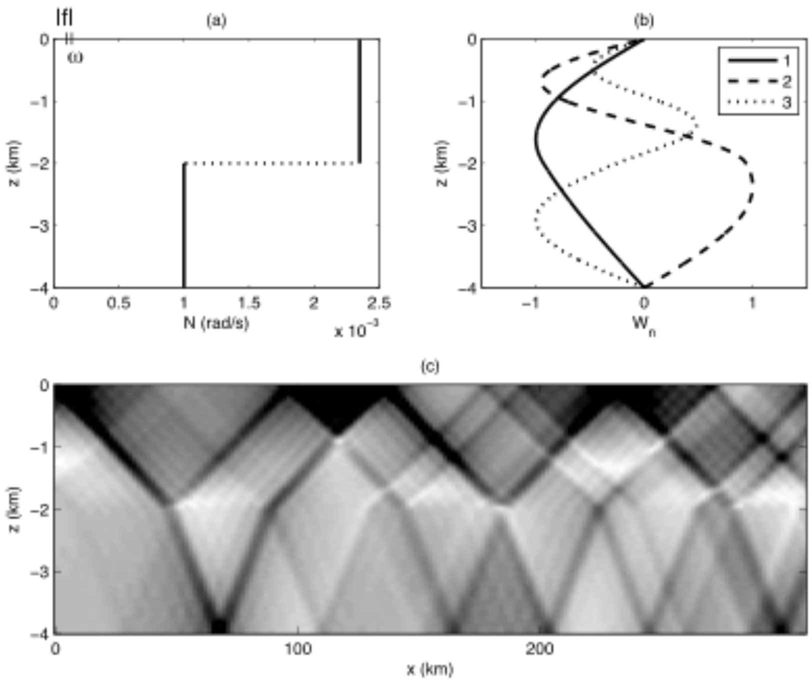


Fig. 5. Physical processes affecting internal waves.

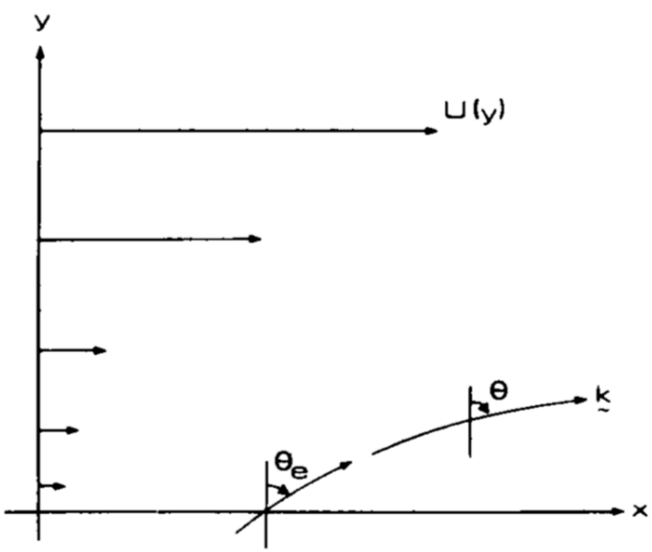
1.4. Propagation and dissipation of waves

C. Wave reflection/refraction

- by a density jump



- by the horizontal shear of a background velocity



1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

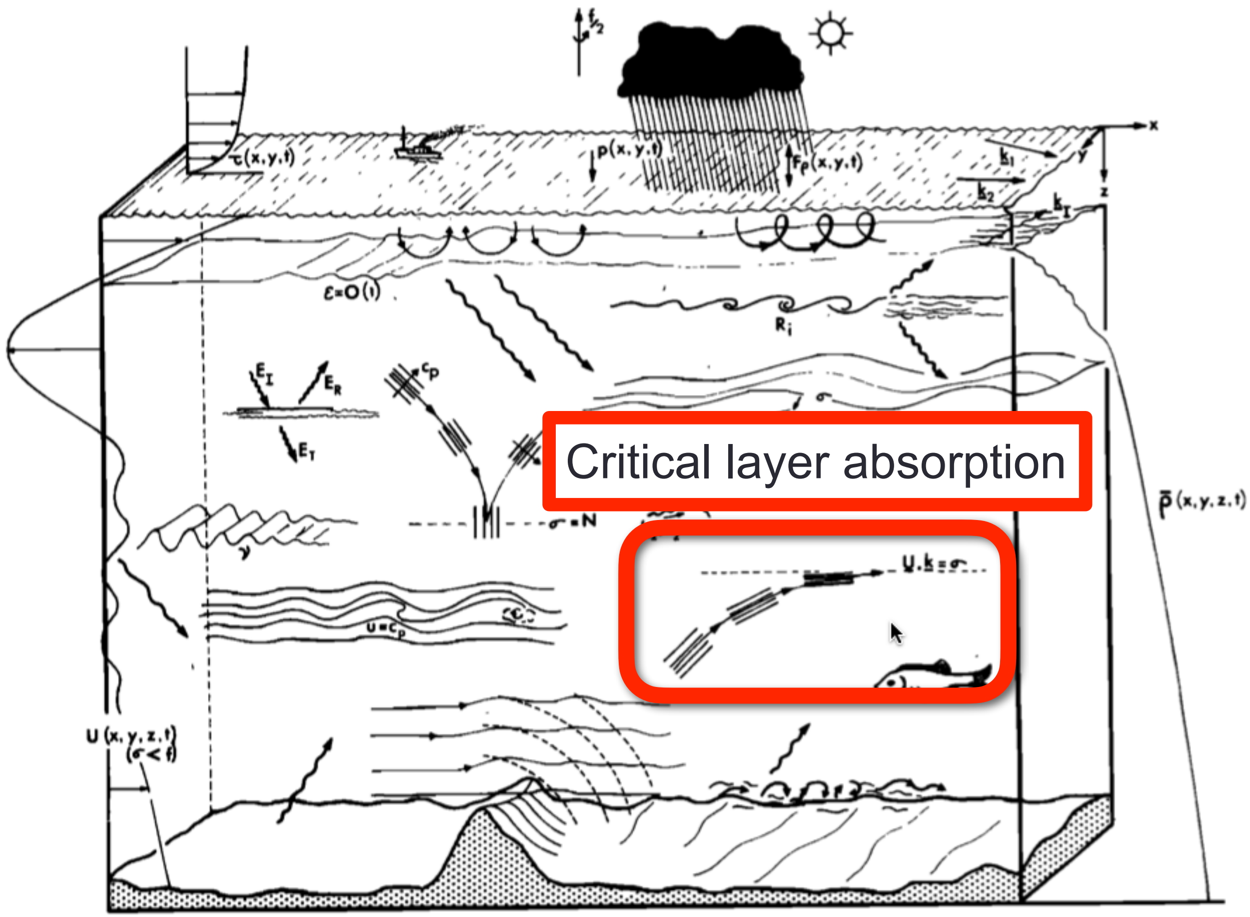


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

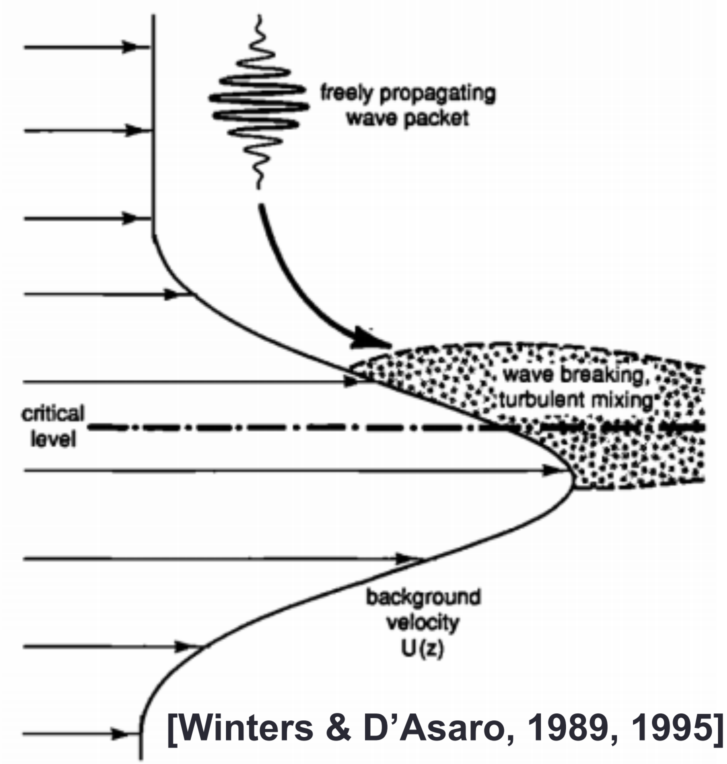
D. Critical layer absorption

A critical layer is a region where the velocity of the flow is equal to the phase speed of the waves **[Bretherton & Garrett, 1968]**

$$\omega = \vec{k} \cdot \vec{U}$$

At such level, energy is transferred to the mean flow, turbulence is generated, and the wave is strongly attenuated.

(See LeBlond & Mysak, p 387)



[Winters & D'Asaro, 1989, 1995]

1.4. Propagation and dissipation of waves

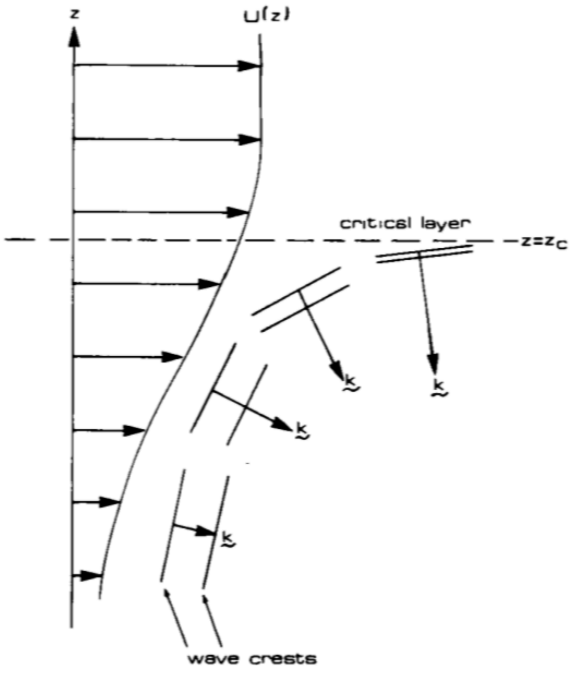
D. Critical layer absorption

- Activity:

We consider an internal wave propagating upward in a vertically sheared flow $\vec{U} = [U(z), 0, 0]$
(with no rotation for simplicity)

1. Write the vertical component of the group speed

2. Explain what happens when the wave reaches a critical layer



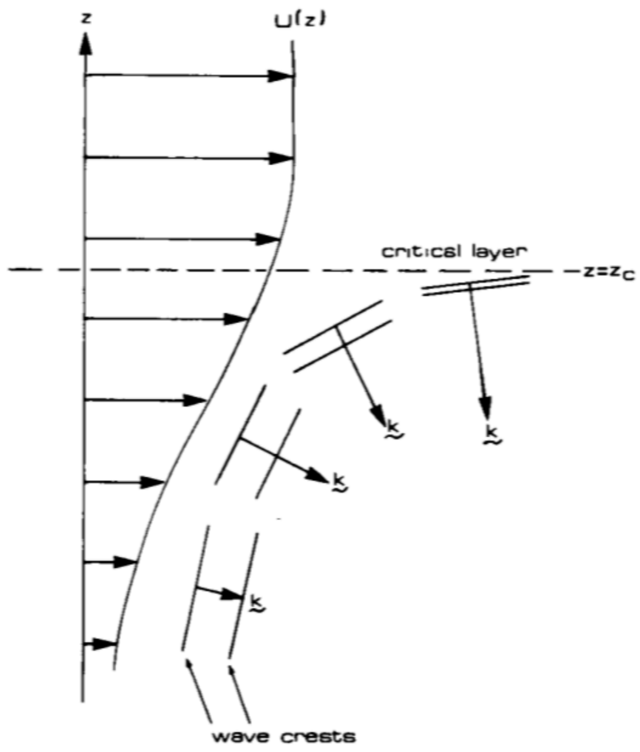
[LeBlond & Misak]

1.4. Propagation and dissipation of waves

D. Critical layer absorption

- For small wave in a presence of a vertical shear: If the background flow increases as a small wave approaches, it is stretched and rotated due to the shear until its group velocity is nearly horizontal, wavelength increases infinitely, lending the wave's energy to the background.

- For large amplitude waves, the steepness of the wave causes breaking before absorption occurs.

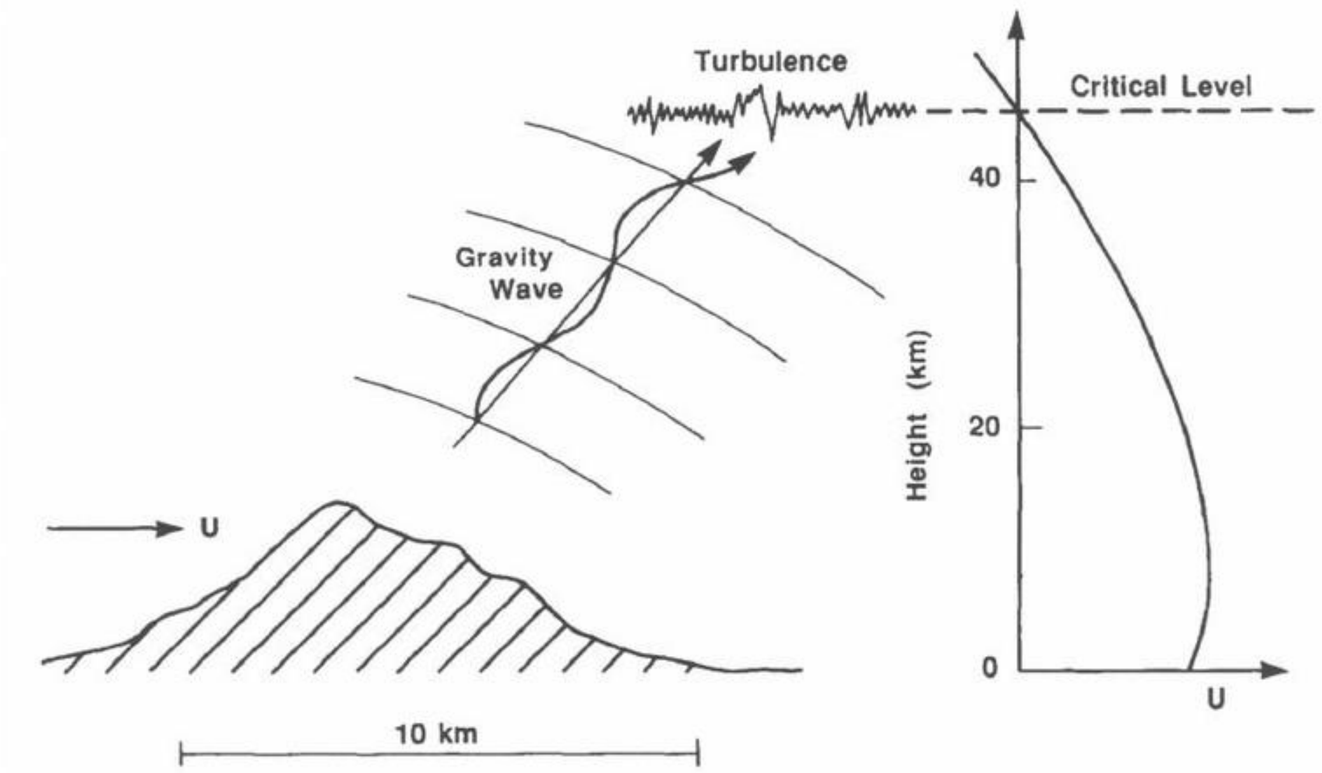


[LeBlond & Misak]

1.4. Propagation and dissipation of waves

D. Critical layer absorption

- Short internal Lee waves have a high probability of encountering critical layers and be absorbed in the lower 1 km above ocean bottom:



1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

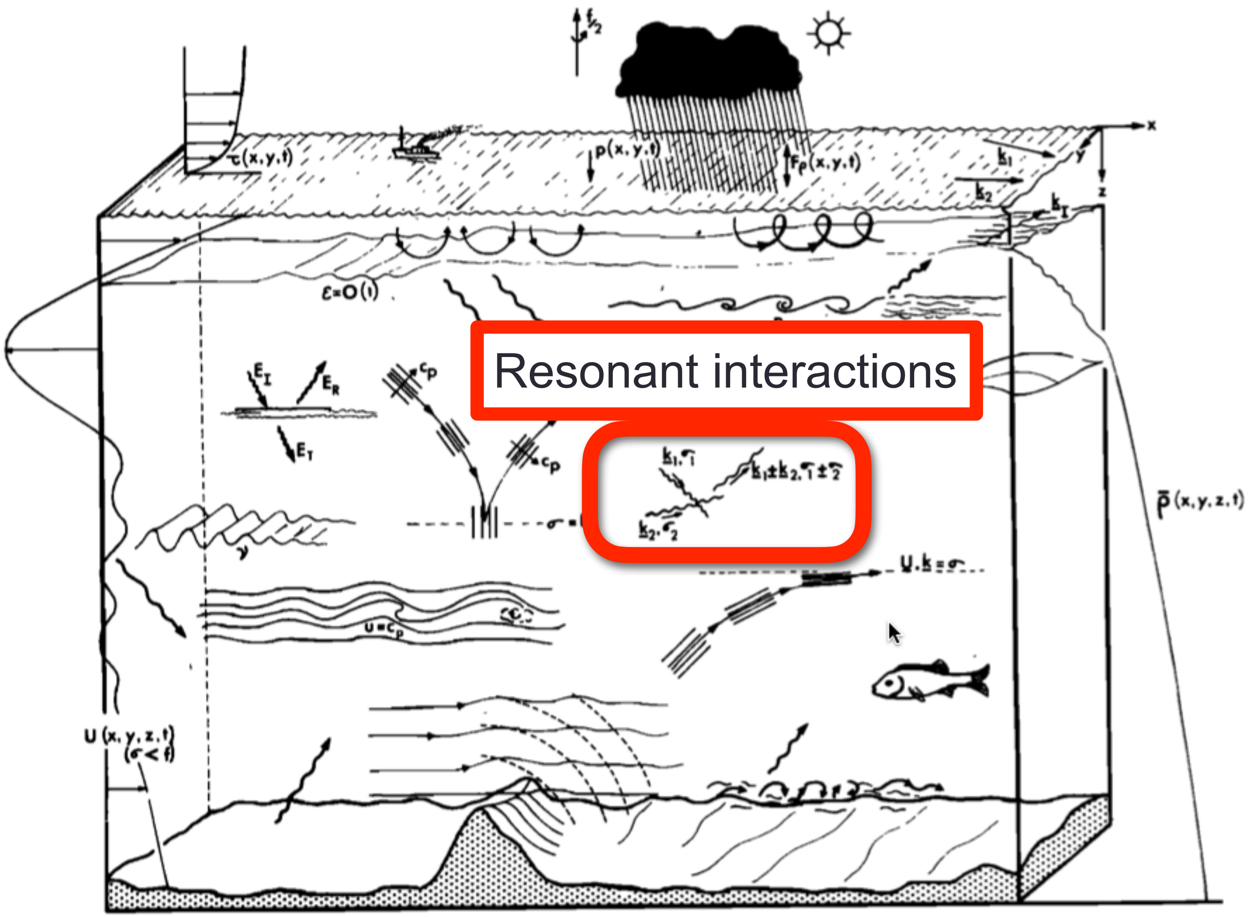


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

E. Resonant interactions

A resonant second-order interaction between three internal waves may occur if :

$$\vec{k_1} \pm \vec{k_1} \pm \vec{k_3} = 0$$

$$\omega_1 \pm \omega_2 \pm \omega_3 = 0$$

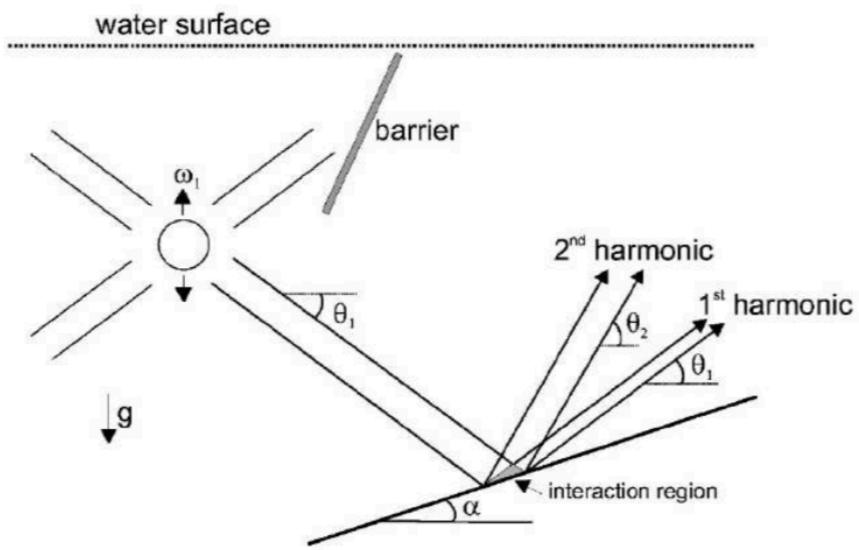
Under this special circumstance, **non-linear terms efficiently transfer energy from one scale to another.**

See LeBlond & Mysak, p 333

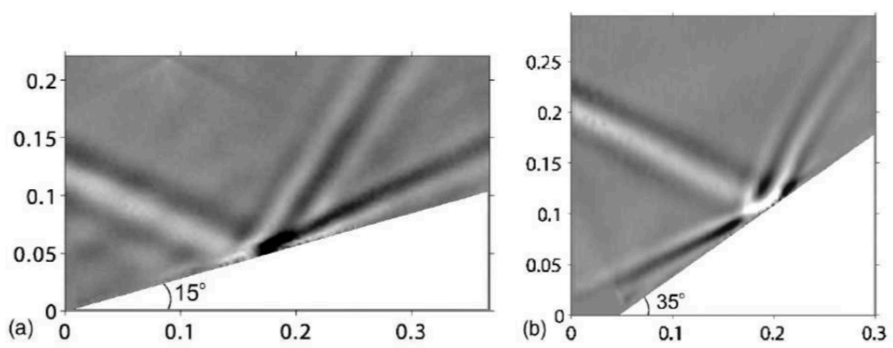
1.4. Propagation and dissipation of waves

E. Resonant interactions

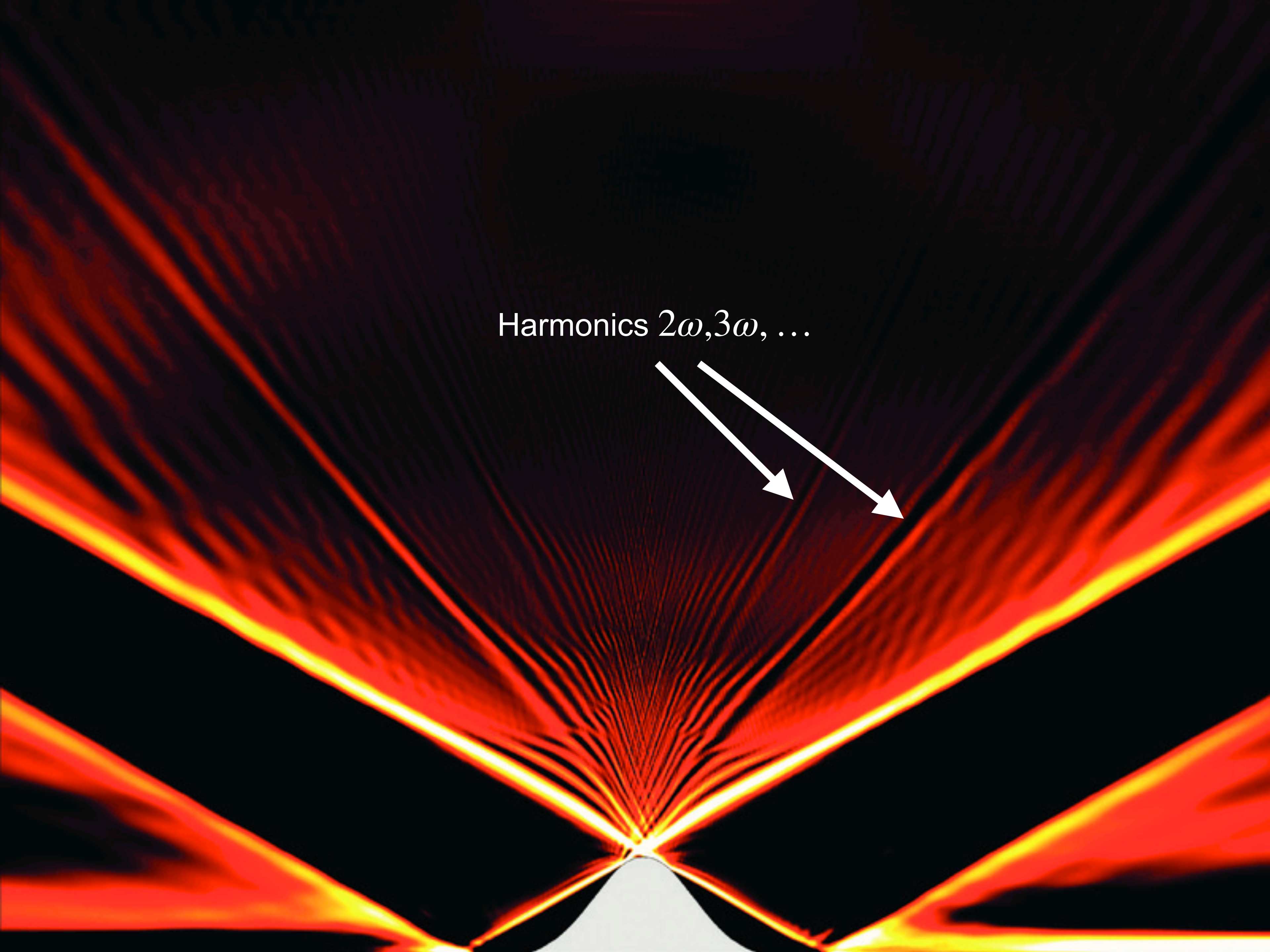
Generation of higher harmonics at boundaries (or density gradients):



Results from laboratory experiments:
[Peacock and Tabei, 2005]



see textbook by Gerkema, p 190



Harmonics $2\omega, 3\omega, \ldots$

1.4. Propagation and dissipation of waves

E. Resonant interactions

Generation of higher harmonics at the bottom: *(see Gerkema, p 190)*

$$
\begin{aligned}
J(\nabla^2\psi,\psi) &= -u\nabla^2 w + w\nabla^2 u = -\mu^{-1}F(\mu^2+1)F'' + F\mu^{-1}(\mu^2+1)F'' = 0\\
J(v,\psi) &= uv_x + wv_z = -i\,\frac{f}{\omega}\mu^{-1}FF' + i\,\frac{f}{\omega}\mu^{-1}FF' = 0\\
J(b,\psi) &= ub_x + wb_z = -i\,\frac{N^2}{\omega}FF' + i\,\frac{N^2}{\omega}FF' = 0\,.
\end{aligned}
$$

Hence a beam taken in isolation F (with G=0) also fulfills the non-linear equation. The same holds for G (if F=0).

However, wherever F and G occur together (reflection at the bottom), nonlinear effects will, in general, set in.

1.4. Propagation and dissipation of waves

E. Resonant interactions

Generation of higher harmonics at the bottom: *(see Gerkema, p 190)*

Non-linear equations:

$$\begin{aligned}
\nabla^2\psi_t + J(\nabla^2\psi,\psi) - f v_z + b_x &= 0 \\
v_t + J(v,\psi) + f\psi_z &= 0 \\
b_t + J(b,\psi) - N^2\psi_x &= 0\,,
\end{aligned}$$

We consider weakly nonlinear waves, monochromatic (with frequency ω) at lowest order, and write the fields in a formal expansion in which ε, a measure of the intensity of the wave, serves as the small parameter:

$$\begin{aligned}
\psi &= \epsilon\{\Psi\exp(-i\omega t)+\text{c.c.}\} + \epsilon^2\{\Psi_0 + [\Psi_2\exp(-2i\omega t)+\text{c.c.}]\} + \cdots \\
v &= \epsilon\{V\exp(-i\omega t)+\text{c.c.}\} + \epsilon^2\{V_0 + [V_2\exp(-2i\omega t)+\text{c.c.}]\} + \cdots \\
b &= \epsilon\{\Gamma\exp(-i\omega t)+\text{c.c.}\} + \epsilon^2\{\Gamma_0 + [\Gamma_2\exp(-2i\omega t)+\text{c.c.}]\} + \cdots
\end{aligned}$$

1.4. Propagation and dissipation of waves

E. Resonant interactions

Generation of higher harmonics at the bottom: *(see Gerkema, p 190)*

Lower order:

$$
\begin{aligned}
-i\omega\nabla^2\Psi - fV_z + \Gamma_x &= 0 \\
-i\omega V + f\Psi_z &= 0 \\
-i\omega\Gamma - N^2\Psi_x &= 0\,,
\end{aligned}
$$

Which gives the well-known equation and solution:

$$
(N^2-\omega^2)\Psi_{xx} - (\omega^2 - f^2)\Psi_{zz} = 0\,,
$$

$$
\Psi = F(\xi_+) + G(\xi_-)\,,
$$

1.4. Propagation and dissipation of waves

E. Resonant interactions

Generation of higher harmonics at the bottom: *(see Gerkema, p 190)*

Order ϵ^2, mean field:

$$\begin{aligned}[J(\nabla^2\Psi,\Psi^*)+\mathrm{c.c.}]-fV_{0,z}+\Gamma_{0,x} &= 0\\ [J(V,\Psi^*)+\mathrm{c.c.}]+f\Psi_{0,z} &= 0\\ [J(\Gamma,\Psi^*)+\mathrm{c.c.}]-N^2\Psi_{0,x} &= 0\,.\end{aligned}$$

Which gives :

$$\Psi_0 = \frac{i}{\omega}J(\Psi,\Psi^*)\,.$$

Using previous low order solution:

$$\Psi_0 = \frac{2}{\omega}(\mu_+-\mu_-)\mathrm{Im}[F'(\xi_+)G'(\xi_-)^*]\,.$$

This expression confirms that no nonlinear contributions arise from an interaction of one plane internal wave (F, say) with itself; only junctions of plane waves, involving both F and G, provide nonlinear terms.

1.4. Propagation and dissipation of waves

E. Resonant interactions

<u>Generation of higher harmonics at the bottom</u>: *(see Gerkema, p 190)*

Order ϵ^2, second harmonic:

$$\begin{aligned} -i\omega_2\nabla^2\Psi_2 + J(\nabla^2\Psi,\Psi) - fV_{2,z} + \Gamma_{2,x} &= 0 \\ -i\omega_2 V_2 + J(V,\Psi) + f\Psi_{2,z} &= 0 \\ -i\omega_2\Gamma_2 + J(\Gamma,\Psi) - N^2\Psi_{2,x} &= 0\,, \end{aligned}$$

Which gives :

$$(N^2-\omega_2^2)\Psi_{2,xx} - (\omega_2^2 - f^2)\Psi_{2,zz} = 3i\omega J(\nabla^2\Psi,\Psi)\,.$$

The left-hand side of this equation describes the propagation of free waves at frequency 2ω; the right-hand side, the nonlinear forcing by the lowest-order terms.

The left-hand-side can be written:

$$J(\nabla^2\Psi,\Psi) = -2\mu(1+\mu^2)\left[F'''(\xi_+)G'(\xi_-) - G'''(\xi_-)F'(\xi_+)\right].$$

1.4. Propagation and dissipation of waves

E. Resonant interactions

Generation of higher harmonics at the bottom: *(see Gerkema, p 190)*

We can solve it with

$$F(\xi_+) = \int_0^\infty dk\, a(k)\, e^{ik\xi_+} \, .$$

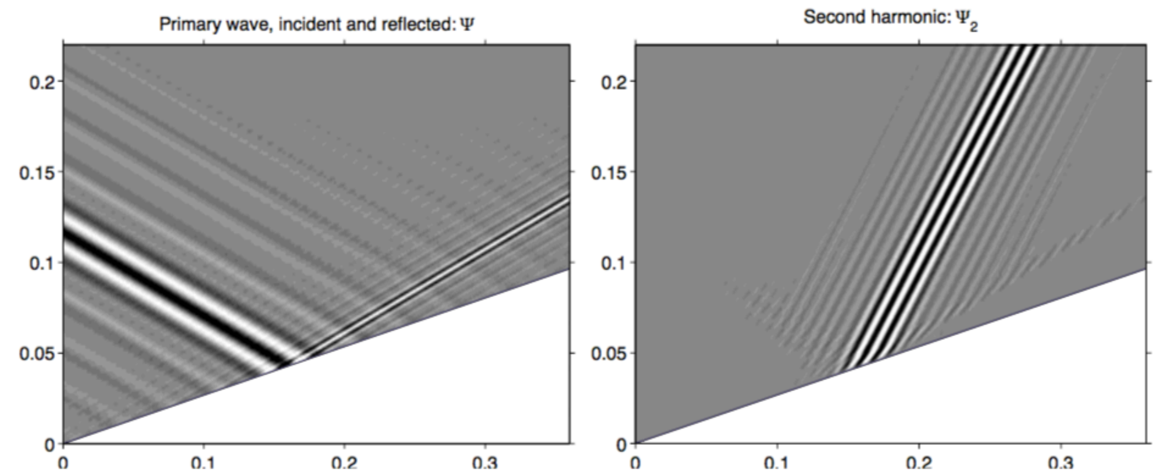


Fig. 9.5: On the left, the primary incident and reflected waves (9.37); the incident wave enters from the left. On the right, the nonlinearly generated second harmonic, (9.41). Parameters as in Figure 9.4a.

1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

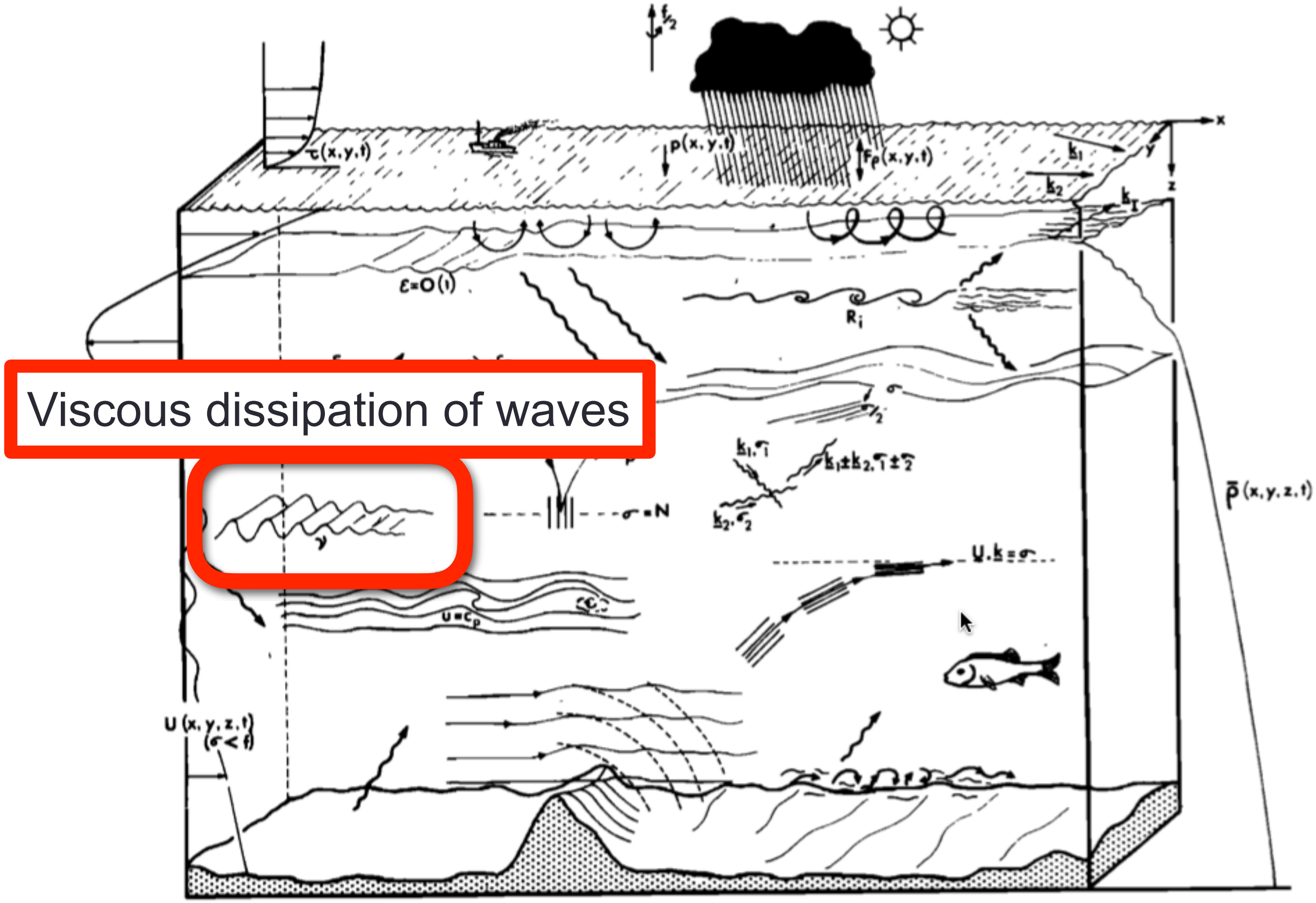


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

F. Viscous dissipation of waves

Internal waves are damped by viscosity. But viscosity acts mostly on small scales.

- *Short internal waves* (< 100 m) are strongly damped
- *Long internal waves* (about 200 km) can propagate for distances of 2000 km or more, but will be damped before being able to establish cross-ocean baroclinic standing waves.
- *Very long internal waves* (> 200 km) are damped by bottom friction

1.4. Propagation and dissipation of waves

F. Viscous dissipation of waves

- Activity:

1. Write Boussinesq equations without rotation, and include viscous terms using constant eddy viscosity diffusion coefficients for momentum:

$$K_{Mh} = K_{Mv} = K_M$$

And density:

$$K_{Bh} = K_{Bv} = K_B$$

2. Write an equation for the vertical velocity only

3. Find the dispersion relation for a solution of the form

$$w = w_0 \exp(\omega t + i(k_x x + k_z z))$$

(where imaginary part of ω is the frequency, and its real part the damping rate)

1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

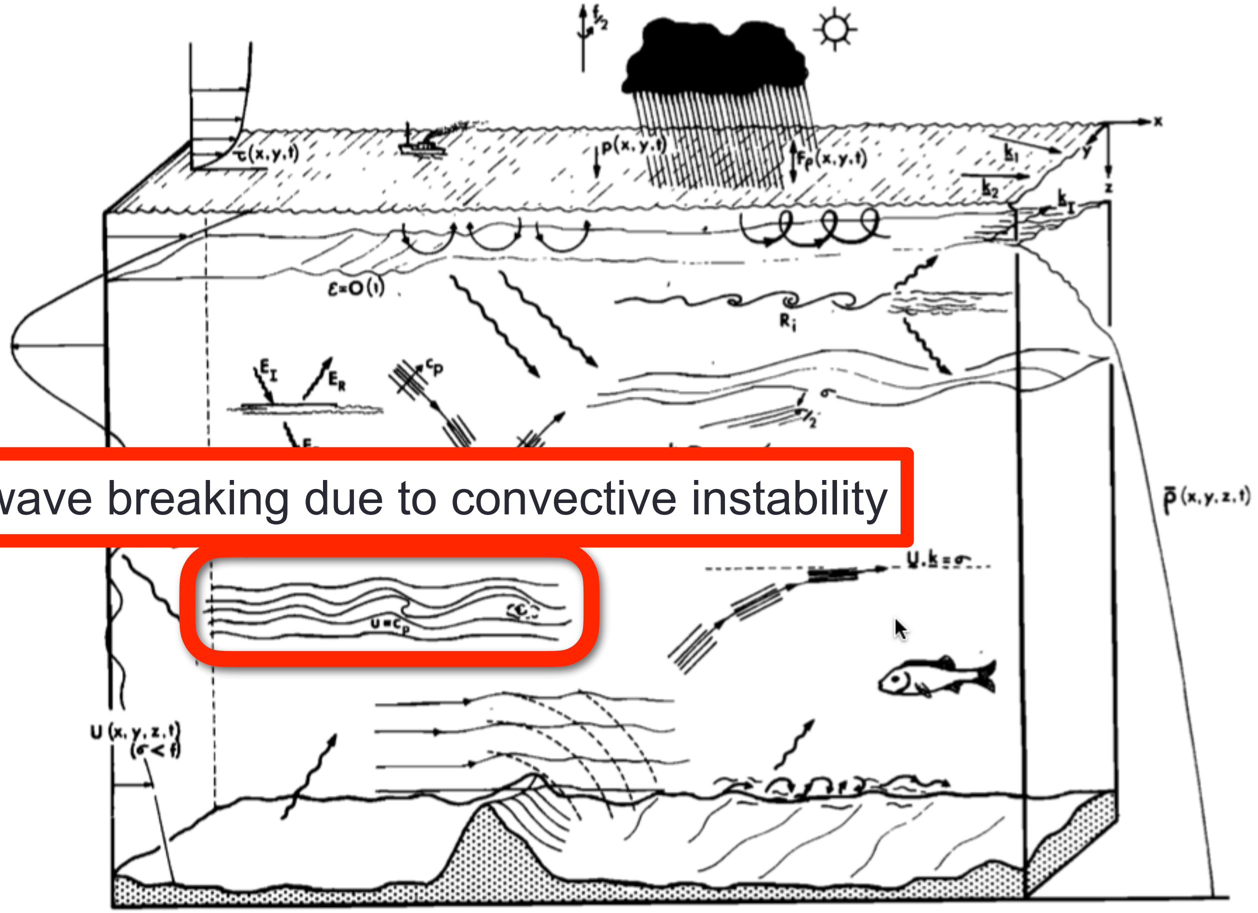


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

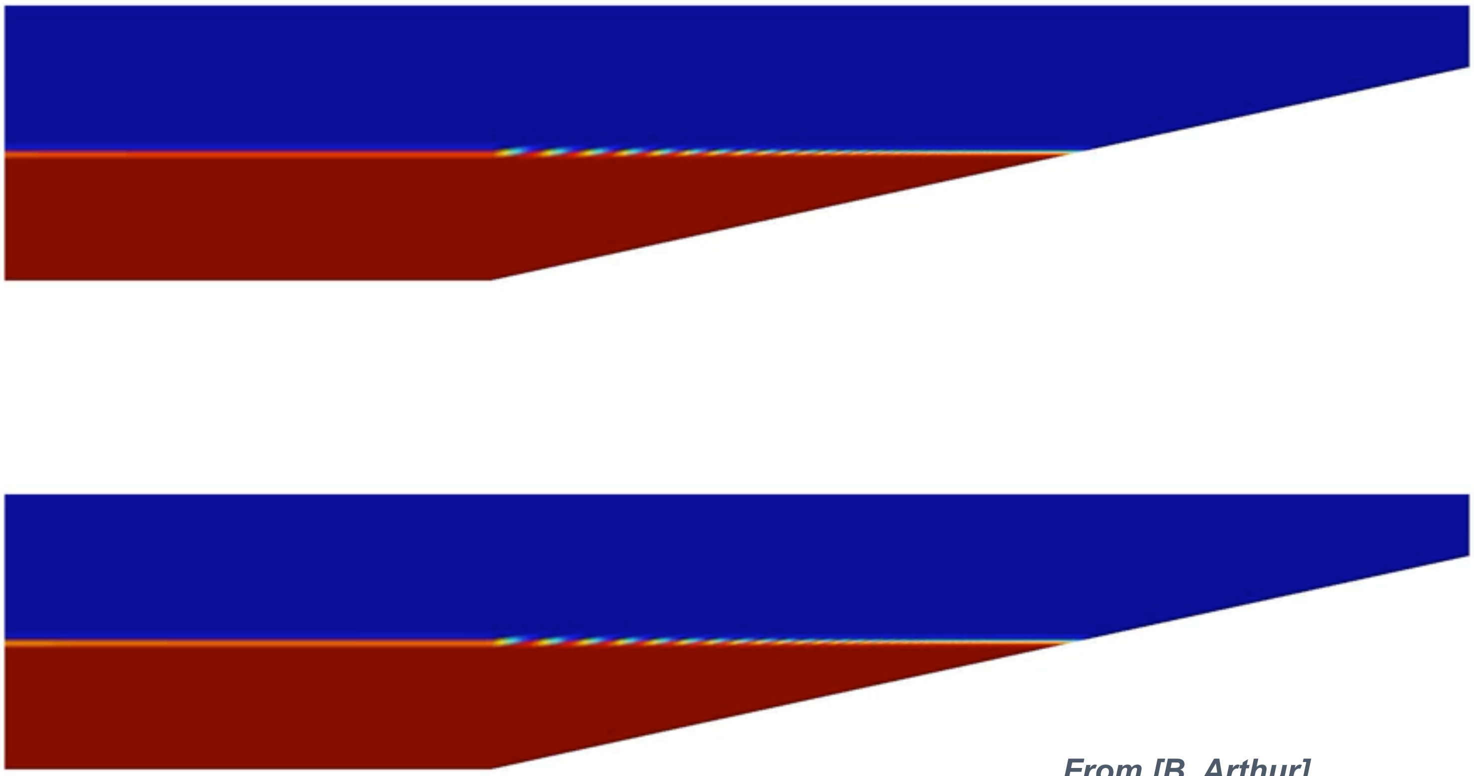
G. Wave breaking due to convective instability

- Unstable density gradients may occur where the local horizontal particle speed exceeds the phase speed of the wave.

The crest (or trough) overtakes the rest of the wave, and the wave rolls over similarly to the surf on the sea surface upon approaching a beach.

1.4. Propagation and dissipation of waves

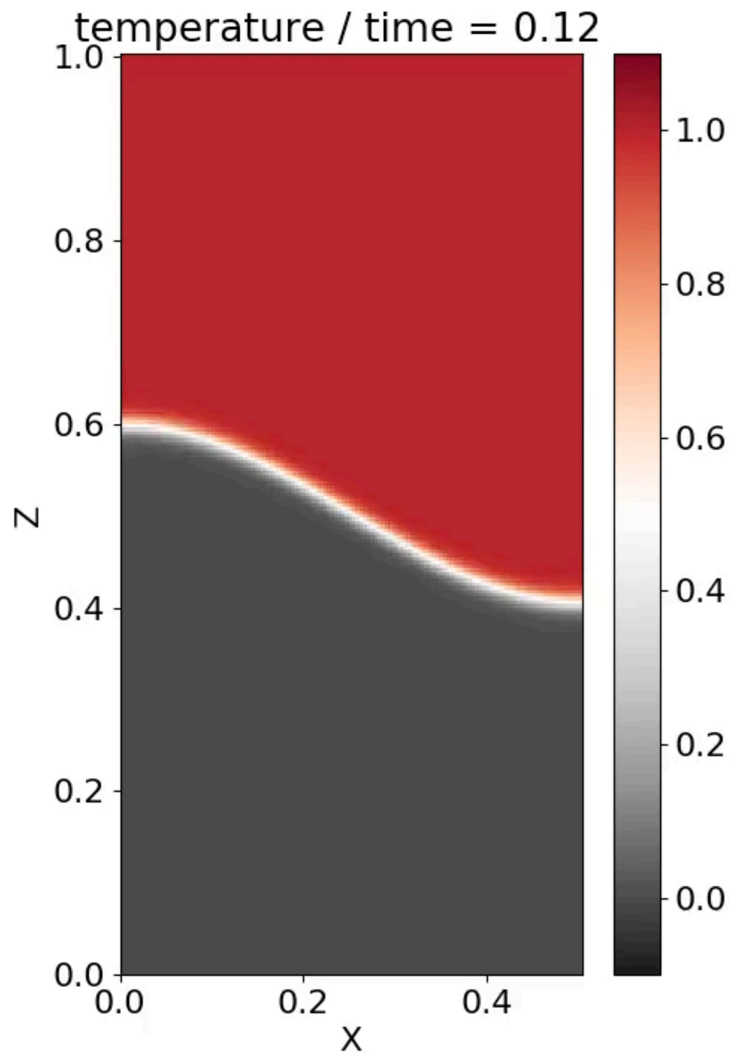
G. Wave breaking due to convective instability



From [B. Arthur]

1.4. Propagation and dissipation of waves

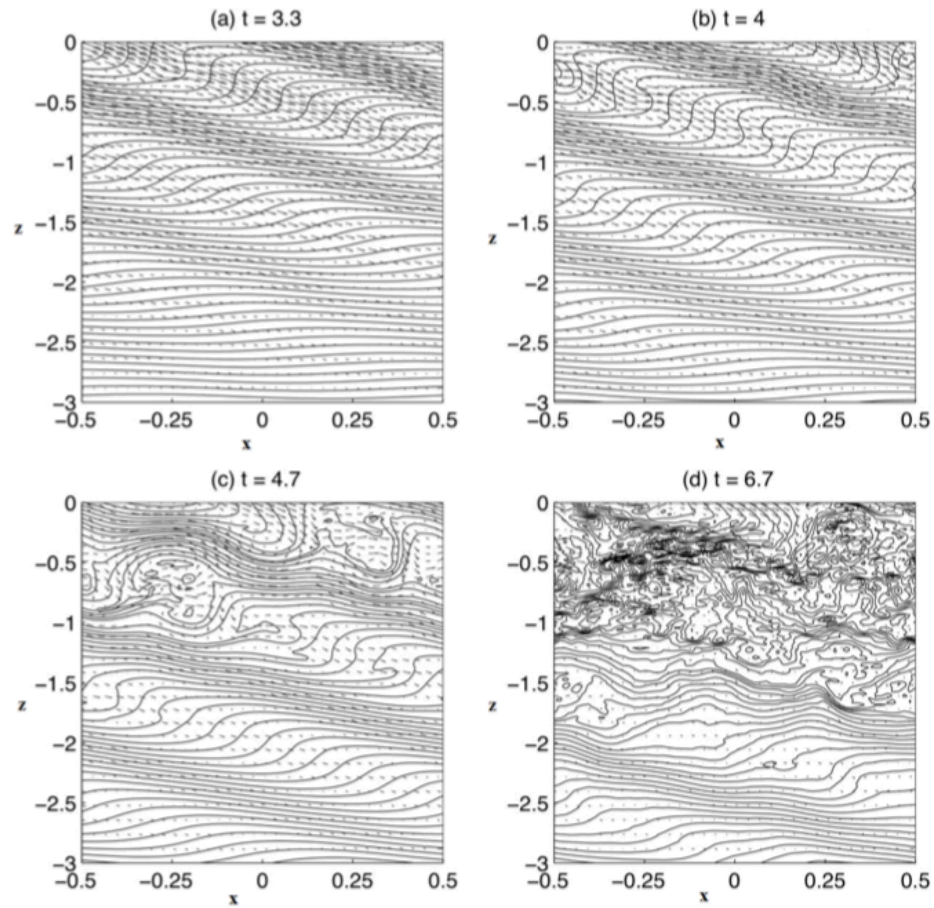
G. Wave breaking due to convective instability



1.4. Propagation and dissipation of waves

G. Wave breaking due to convective instability

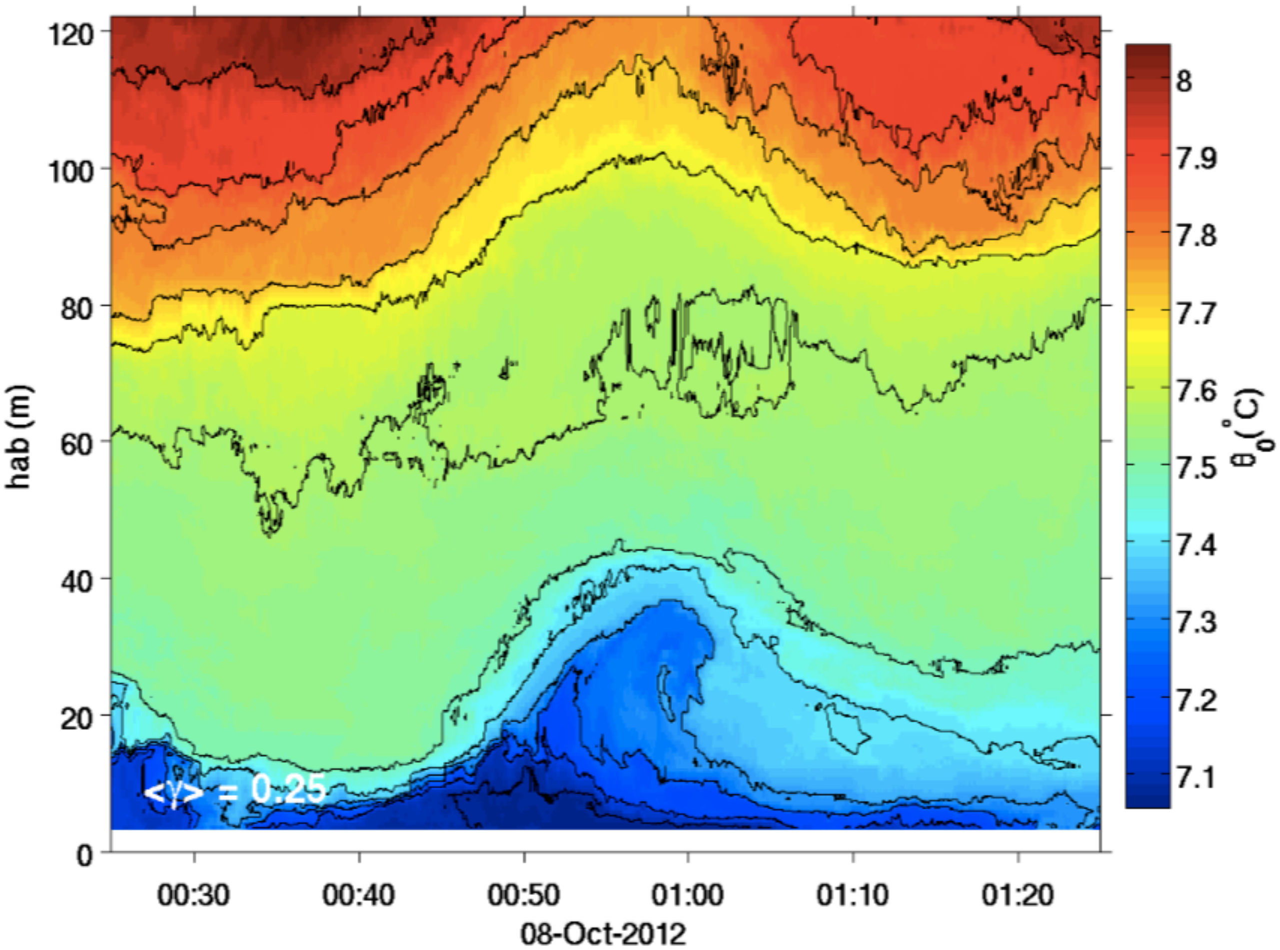
time-series cross-sectional view of a set of isopycnals from relative stability, Figure (a), to relative levels of overturning and breaking in Figures (b-d).
[Liu ,2010]



1.4. Propagation and dissipation of waves

G. Wave breaking due to convective instability

1-hour snapshot of the temperature time/depth series as measured from moored temperature sensors on October-8 2012 on the slopes of Rockall Bank, NE Atlantic (925 m total depth).
[Cyr & Van Haren, 2016]



1.4. Propagation and dissipation of waves

- Review of mechanisms *(see Thorpe75.pdf):*

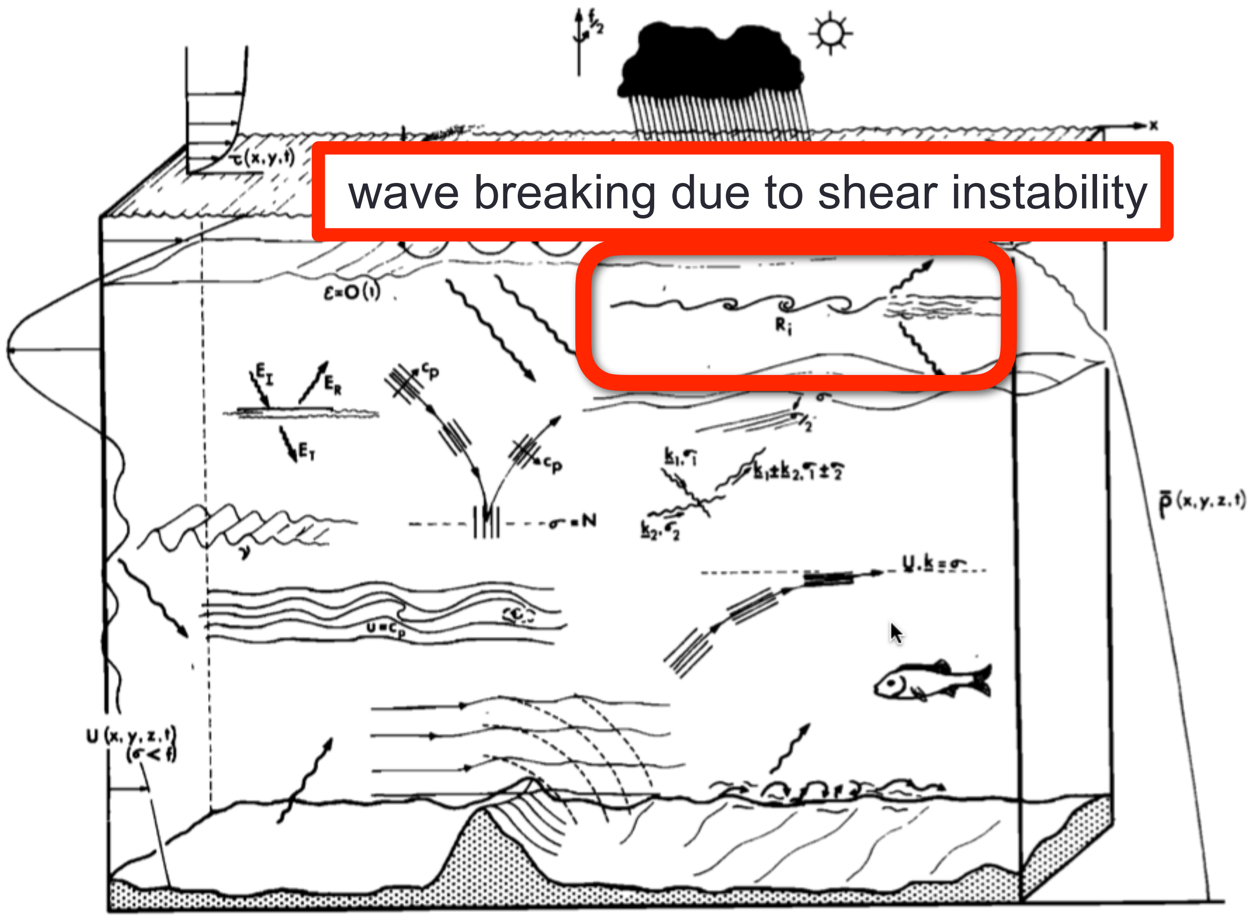
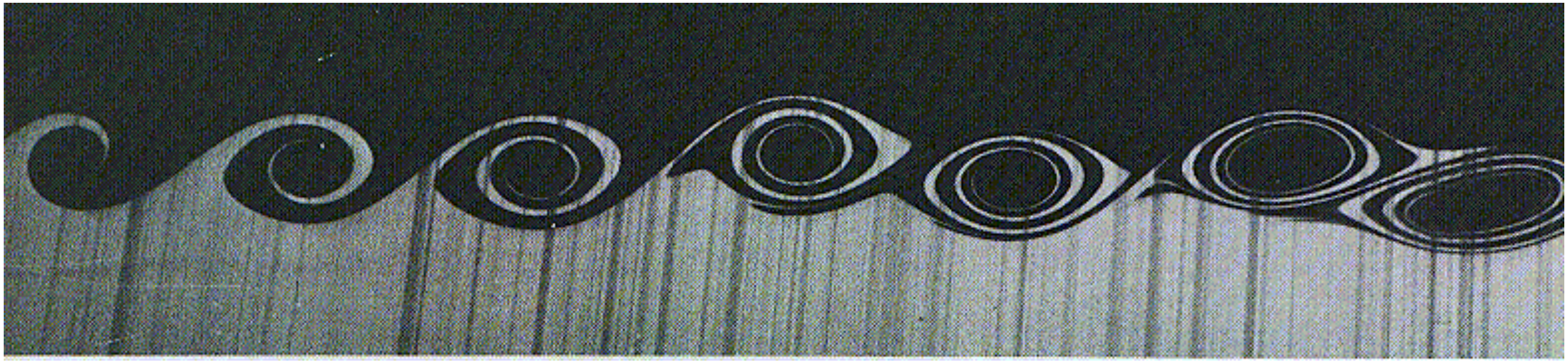


Fig. 5. Physical processes affecting internal waves.

1.4. Propagation and dissipation of waves

H. Wave breaking due to shear instability

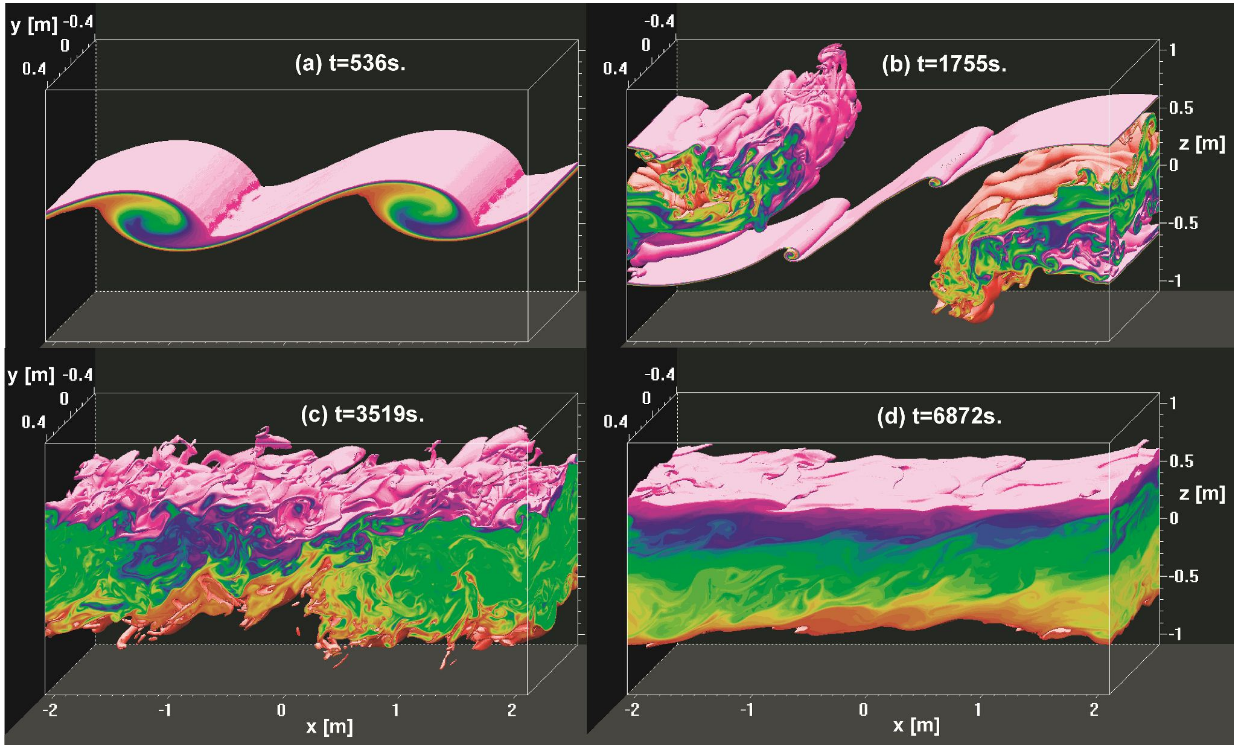
If the shear generated by the wave, added to any preexisting shear, becomes so large that the local Richardson number $Ri = \dfrac{N^2}{|\partial u_h / \partial z|}$ falls below a critical value (usually 0.25) long enough for a shear instability to grow and produce density gradients or billows.



1.4. Propagation and dissipation of waves

H. Wave breaking due to shear instability

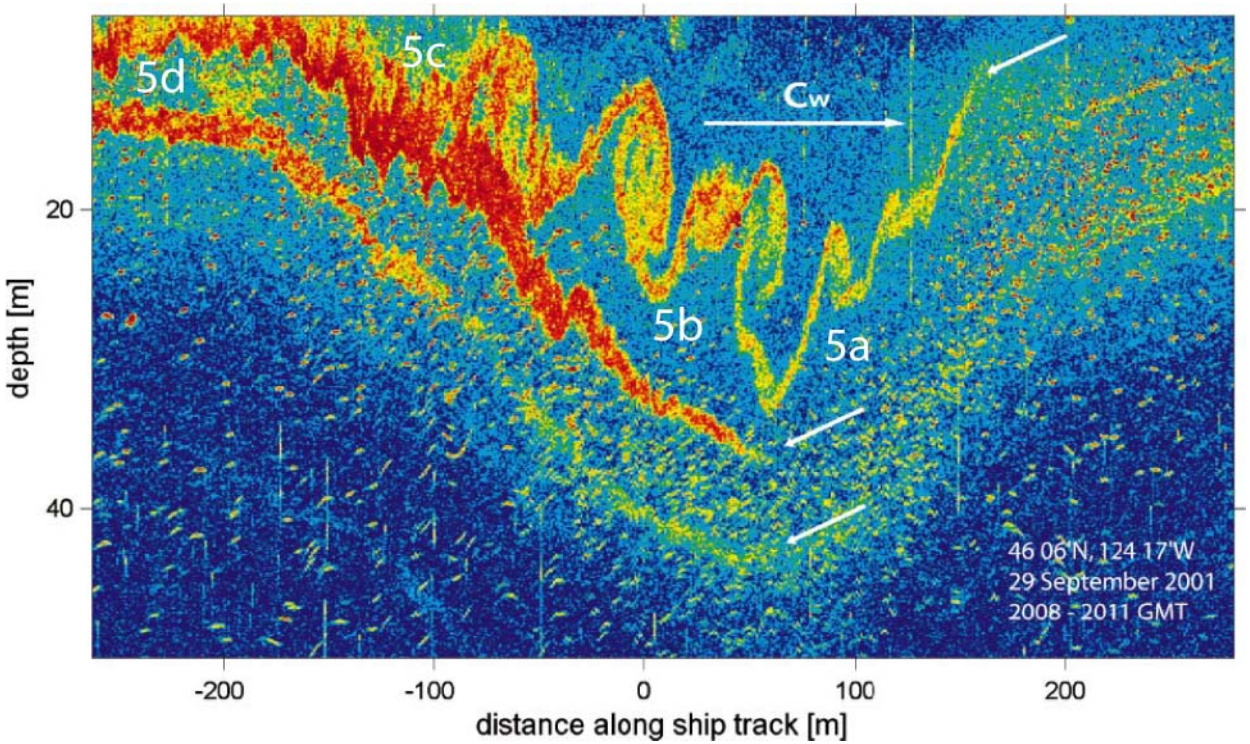
Most of the energy is transferred to turbulent kinetic energy, a fraction is transferred to potential energy (raising the center of gravity of the mixed fluid), and a fraction can be radiated as secondary internal waves (with horizontal scales related to billows scale)



Direct numerical simulation mixing across an interface in the ocean. Colors indicate intermediate values of salinity found in the interfacial layer ***[W. Smyth]***

1.4. Propagation and dissipation of waves

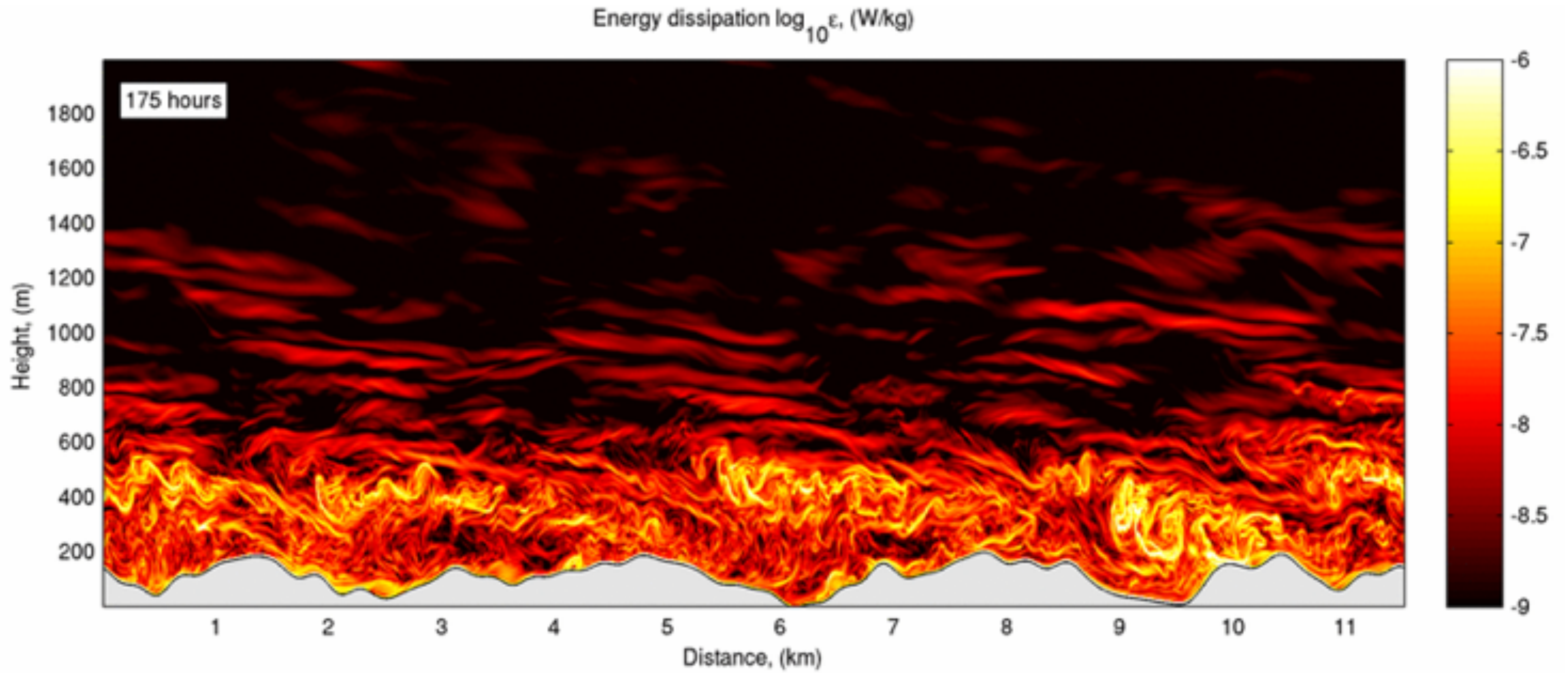
H. Wave breaking due to shear instability



Acoustical snapshot of a nonlinear internal gravity wave approaching the Oregon coast. Near the surface the current flows from left to right, the direction of wave propagation, while the underlying fluid flows right to left to compensate. The resulting current shear generates the Kelvin-Helmlholtz billows and turbulence. ***[Moum et al, 2003]***

1.4. Propagation and dissipation of waves

H. Wave breaking due to shear instability



Breaking Lee waves: Energy dissipation in log10(W/kg) from a high-resolution numerical simulation of lee waves forced by the prescribed mean flow with parameters typical for the Drake Passage region of the Southern Ocean **[From N. Nikurashin]**

1.4. Propagation and dissipation of waves

Theory of linear internal waves is applicable only if their amplitude is small, but internal-wave amplitudes can be quite large. When do internal-wave dynamics become **nonlinear** ?

→ **If particles displacements are comparable to the wavelength**

For a monochromatic wave : $u = U_0 \cos(k_x x + k_z z - \omega t)$

→ The particle displacement is U_0/ω

→ The horizontal wavelength is $2\pi/k_x$

So we need, using the dispersion relation and neglecting rotation,

$$U \ll \frac{2\pi N}{k_z}.$$

This is equivalent to having a Froude number $Fr = \dfrac{U}{NH} \ll 1$

1.5. Wave energy spectra

In nature, It is impossible to isolate triplets or quartets of interacting waves. There is a continuous spectra of waves spanning several order of magnitudes of wavenumbers and frequencies. The interactions take place in a continuous fashion and resonant and forced energy transfers act continuously over wavenumber and frequency domain.

Redistribution of wave action between all possible interacting multiplets. Wave action is not conserved for individual multiplets, but becomes a conserved quantity of the whole spectrum.

1.5. Wave energy spectra

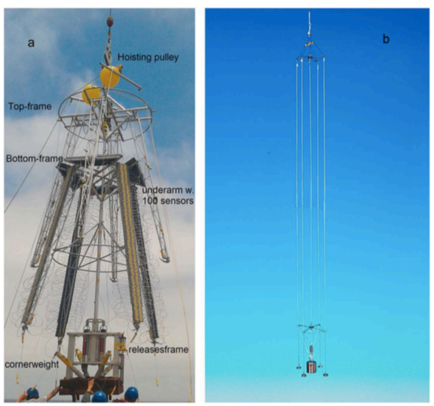
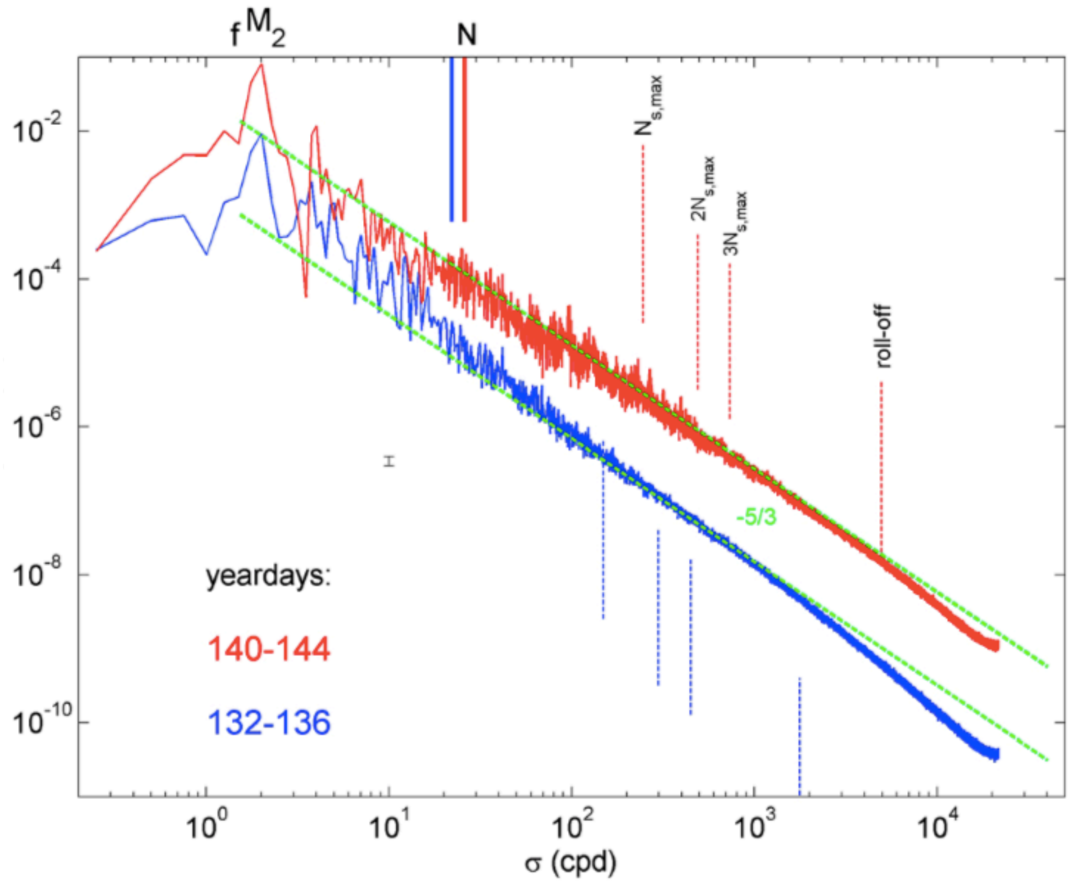


Figure 1. (a) Compacted 3-D mooring array of temperature sensors just before deployment in the ocean. (b) Model of the unfolded array to scale.

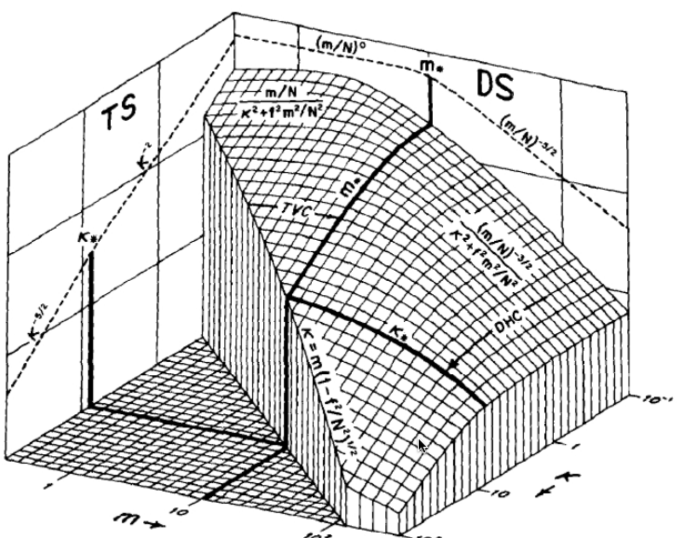


Observed Temperature power spectra. Besides inertial (f) and semidiurnal lunar tidal (M2) frequencies several buoyancy frequencies are indicated including 4 day large-scale mean N and the maximum small-scale Ns,max. **[Van Haren et al., 2016]**

1.5. Wave energy spectra

- The ***Garrett–Munk spectrum*** *is* an empirical spectrum for internal-wave energy based on observations, simple dimensional considerations, and elementary physics. It has been shown to conform to a large number of observations

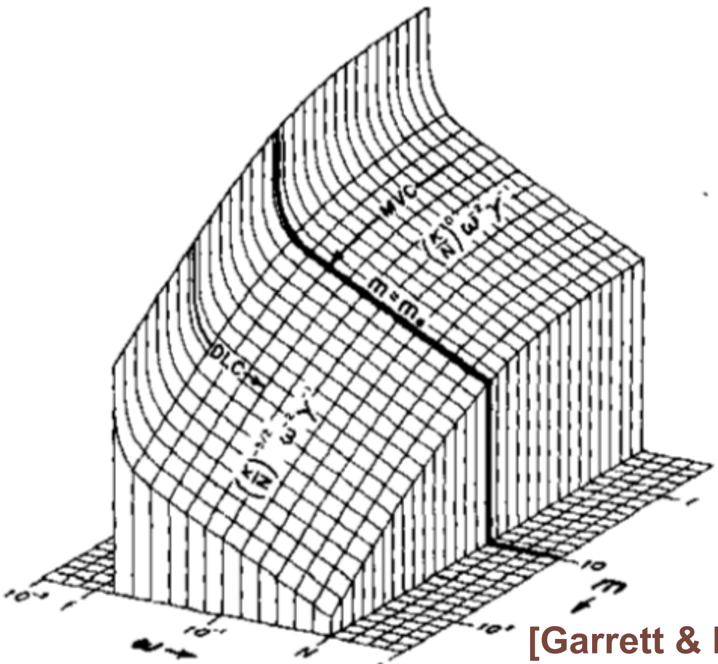
- It gives the spectral energy density anywhere in the ocean *(dimensionless constant E setting the overall energy level)*

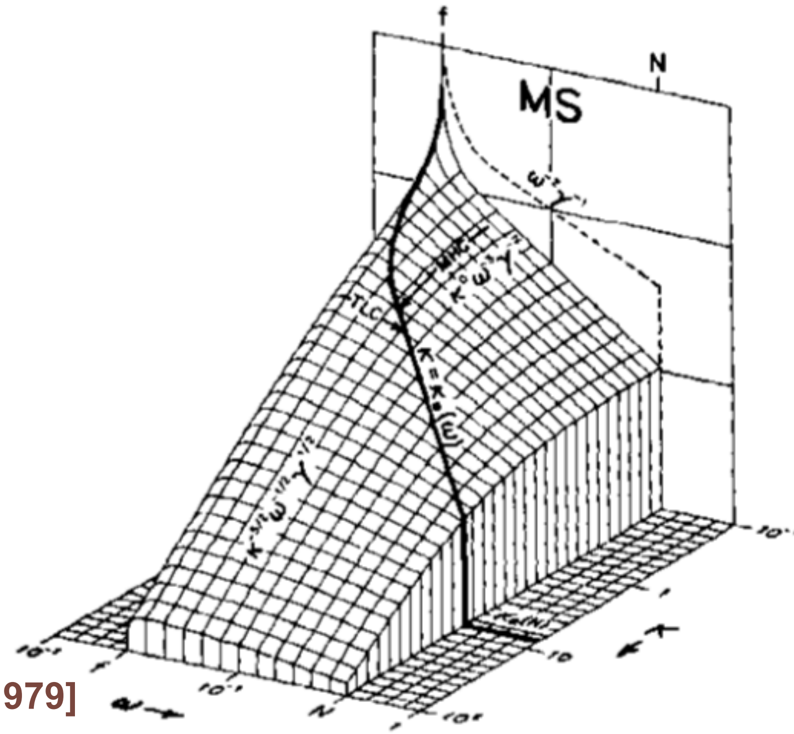


$$E(k,m) = \frac{3fNE\,m\,m_*^{3/2}}{\pi(m+m_*)^{5/2}\,(N^2k^2+f^2m^2)},$$

[Garrett & Munk, 1979]

1.5. Wave energy spectra

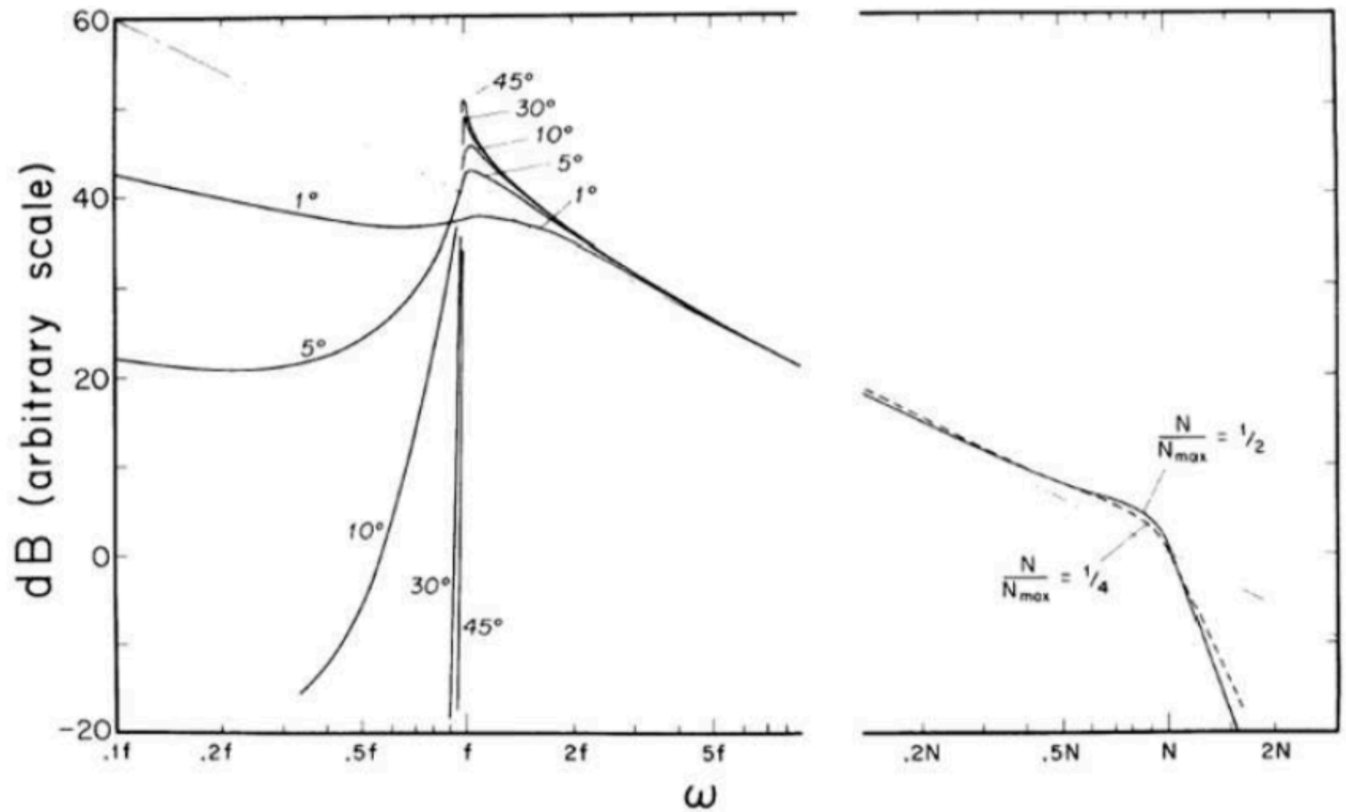




[Garrett & Munk, 1979]

As a function of frequency and/or wavenumbers.

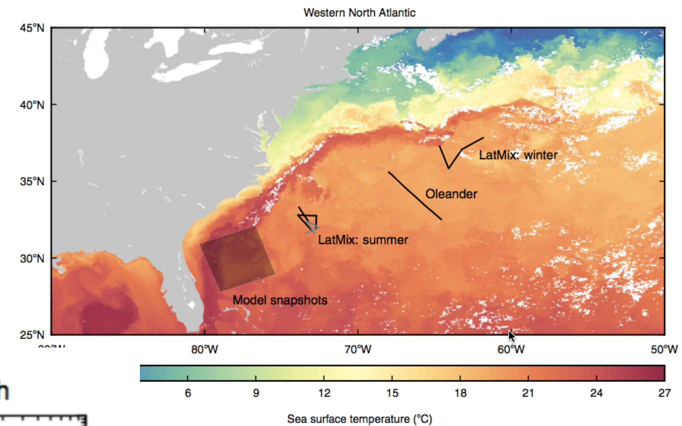
1.5. Wave energy spectra

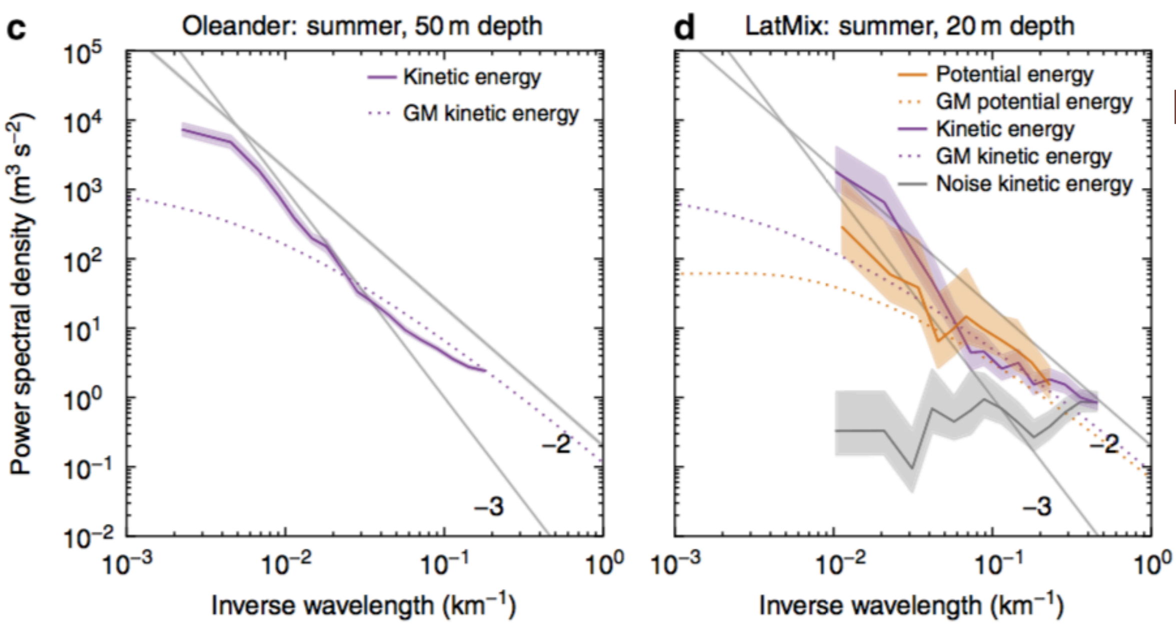


The Garrett-Munk spectrum, at different latitudes, showing a peak at the inertial frequency (f), and a steady fall-off for higher frequencies. **[Munk, 1981].**

1.5. Wave energy spectra

- Ex: Interpretation of observation

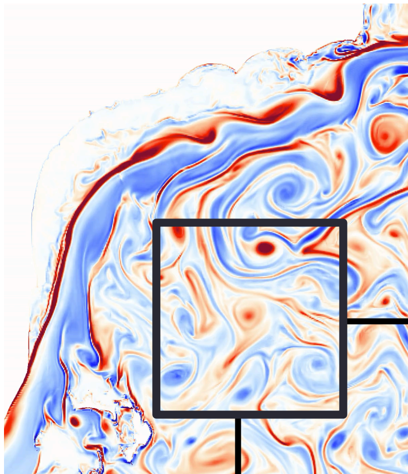




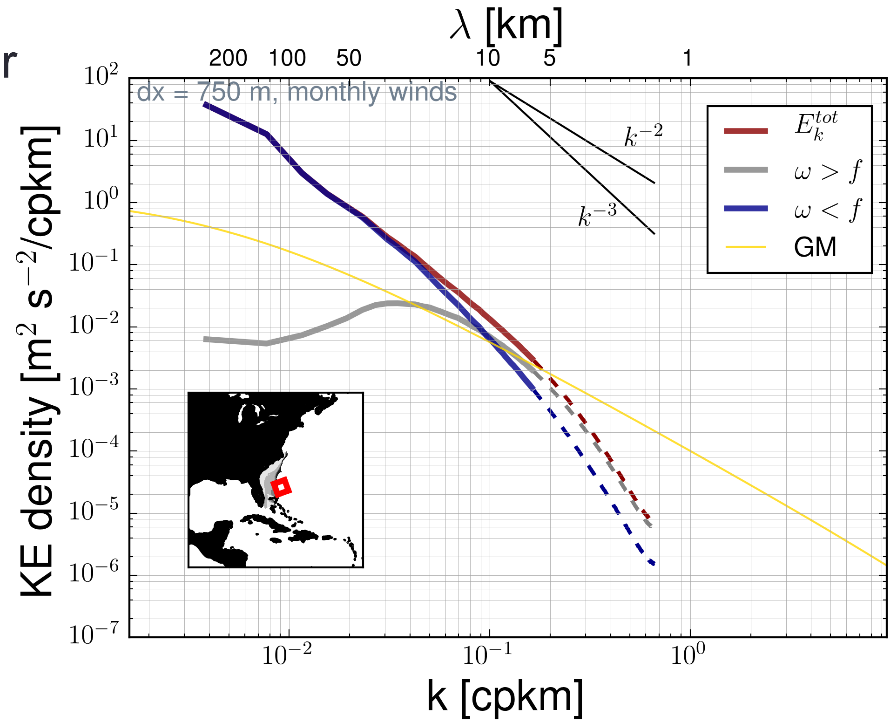
[Callies et al, 2015]

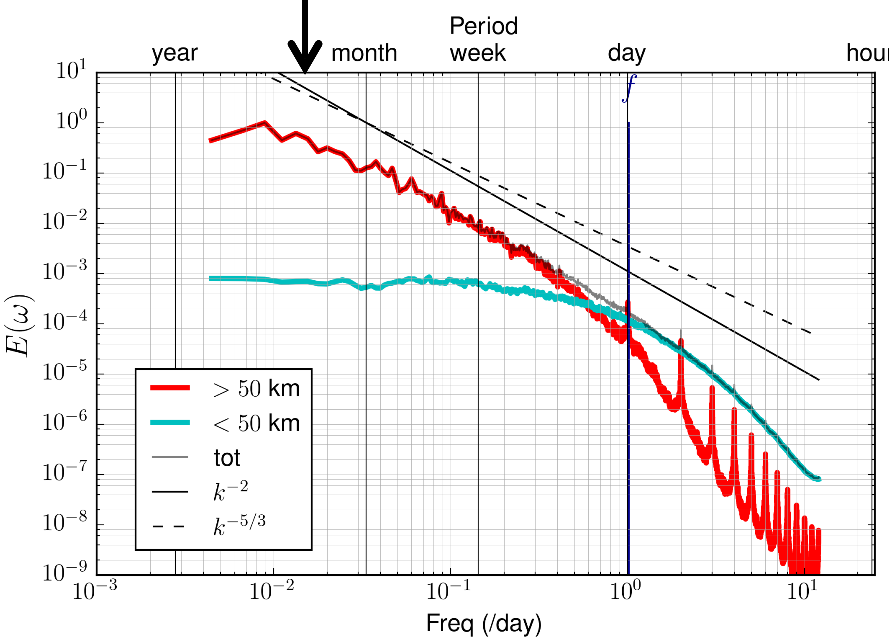
(c) Kinetic energy spectrum at 50 m depth for the Oleander summer data. (d) Potential and kinetic energy spectra at 20 m depth for the LatMix summer experiment. Also shown are the GM model spectra for internal waves in the seasonal thermocline and reference lines with slopes 2 and 3.

1.5. Wave energy spectra



Wavenumber spectra



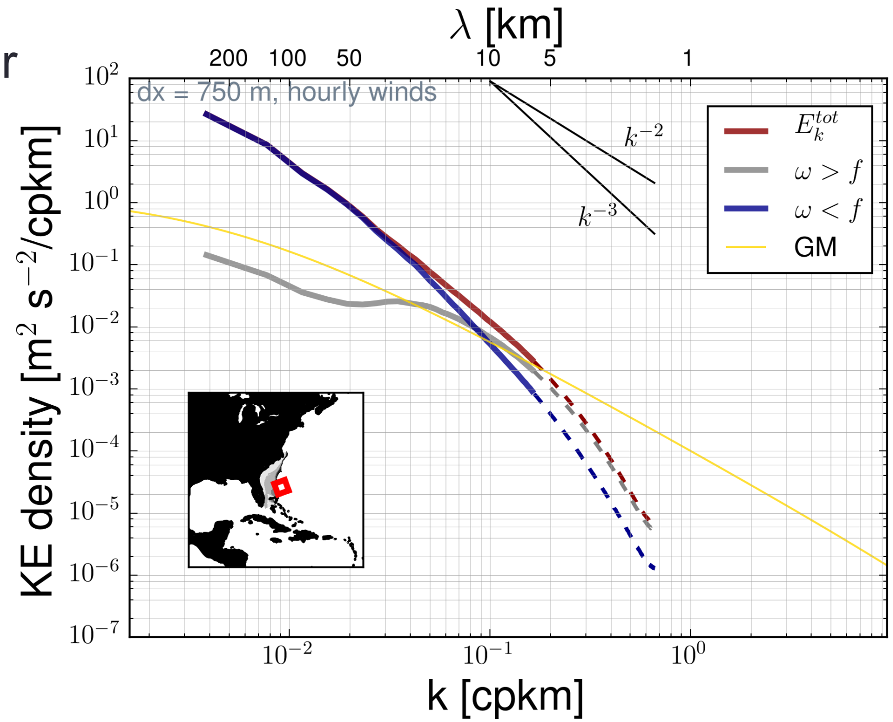


Monthly wind forcings / No tides

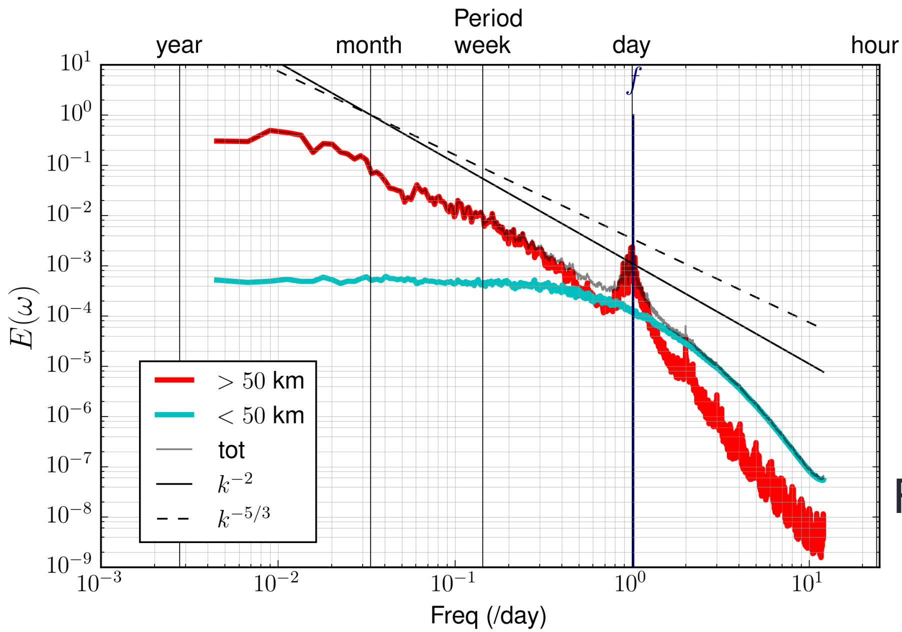
Frequency spectra

1.5. Wave energy spectra

Wavenumber spectra



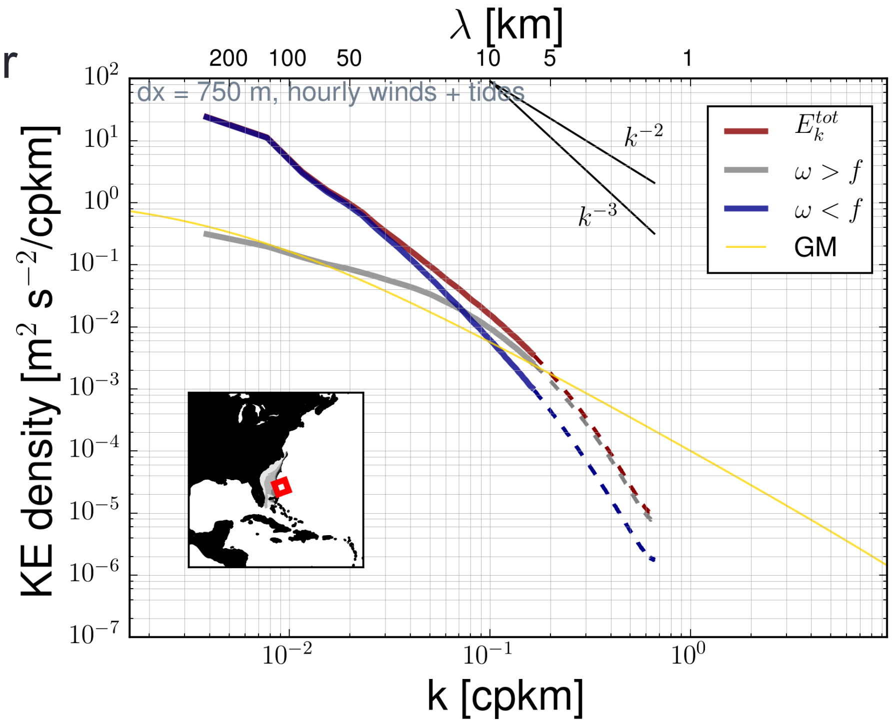
Hourly wind forcings / No tides

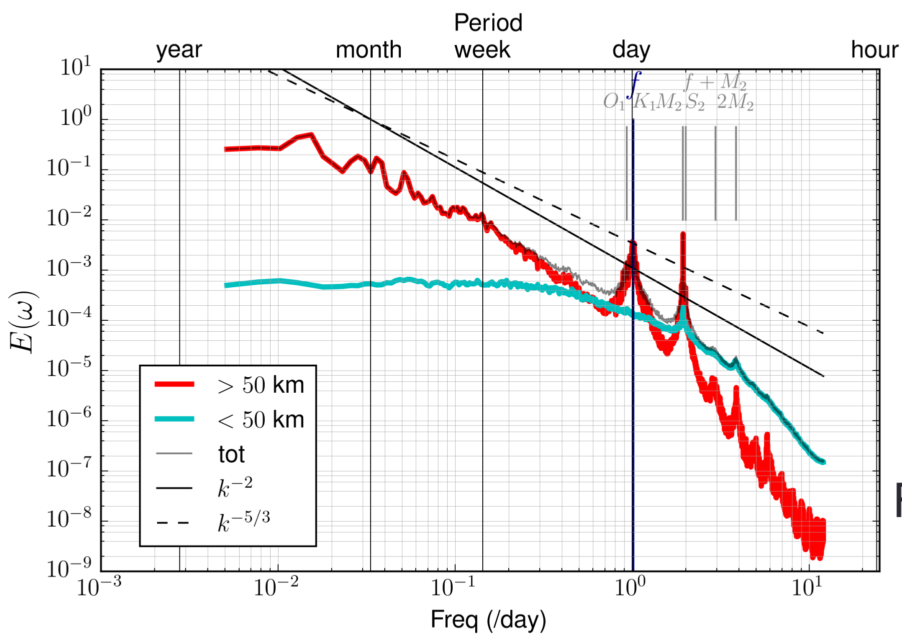


Frequency spectra

1.5. Wave energy spectra

Wavenumber spectra



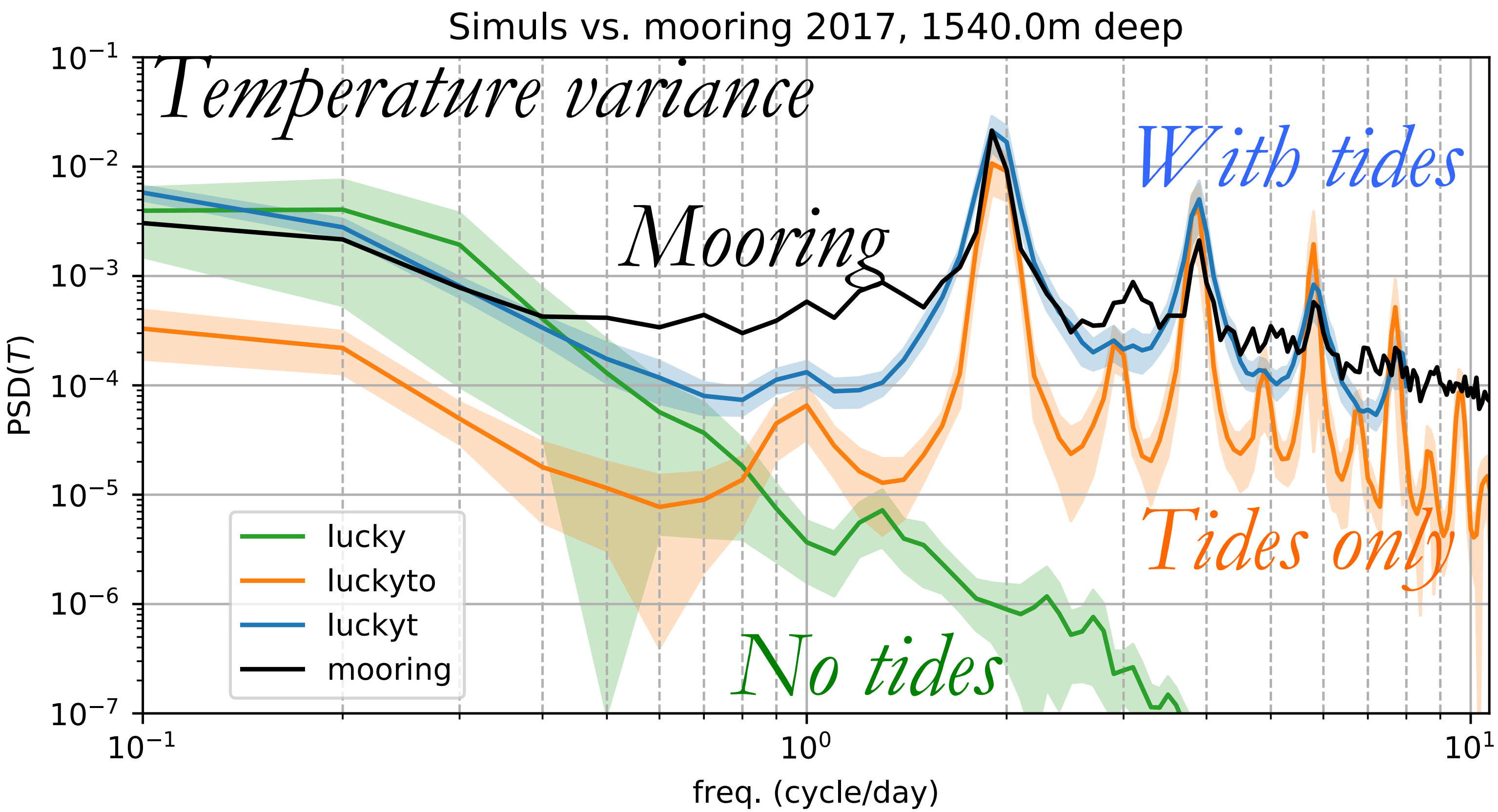


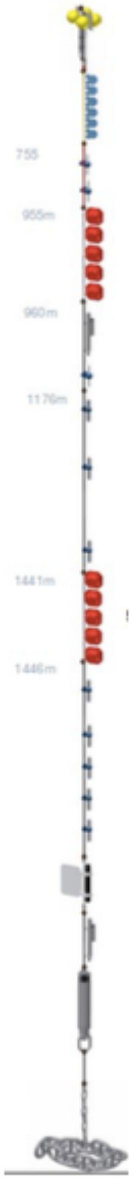
Hourly wind forcings / Tides

Frequency spectra

1.5. Wave energy spectra

Turbulence near the Mid-Atlantic Ridge





1.5. Wave energy spectra

Do deep-ocean kinetic energy spectra represent deterministic or stochastic signals?

Hans van Haren[1]

[1]Royal Netherlands Institute for Sea Research (NIOZ), Den Burg, The Netherlands

Abstract In analogy with historic analyses of shallow-water tide-gauge records, in which tides and their higher harmonics are modified by sea level changes induced by atmospheric disturbances, it is shown that deep-sea currents can be interpreted as motions at predominantly inertial-tidal harmonic frequencies modified by slowly varying background conditions. In this interpretation, their kinetic energy spectra may not be smoothed into a quasi-stochastic continuum for (random-)statistic confidence. Instead, they are considered as quasi-deterministic line-spectra. Thus, the climatology of the internal wave field and its slowly varying background can be inferred from line spectra filling the cusps around nonlinear tidal-inertial harmonics, as suggested previously.

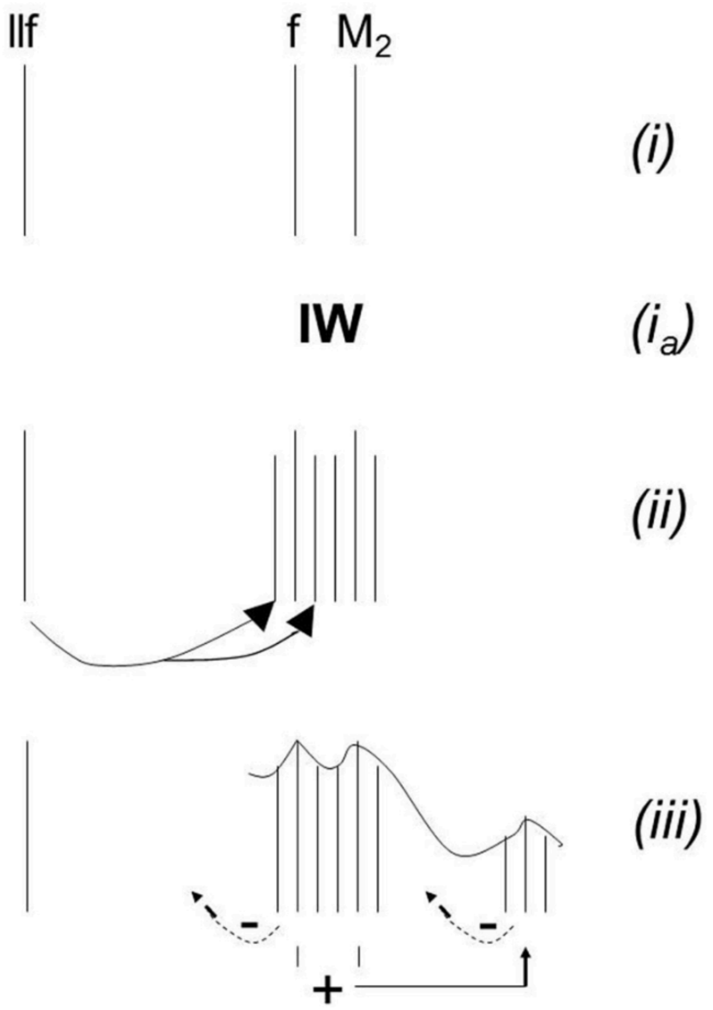
1.5. Wave energy spectra

Impression of the filling of an ocean kinetic energy spectrum, following the steps below.

(i) Motions are generated (externally) at only three distinguished (deterministic) frequencies: a tidal (M2), inertial (f) and a low-frequency sub-inertial (llf).

(ii) After energy transfer to internal waves (IW) at stage (ia), nonlinear interactions with the low-frequency current result in a split of the tidal-inertial lines. Further splits following interaction between the same low-frequency current with the initial split lines create a quasi-continuum tidal-inertial band.

(iii) Finally, the motions in these bands interact and fill the rest of the spectrum. This (re)generates some of the low-frequency motions, through (tidal) rectification, indicated by the -signs.



A few open questions concerning internal waves (among many others)

a) How to compute surface velocities from SSH in presence of internal waves?
b) Extraction of energy from the mesoscale by internal waves
c) How to observe internal waves?

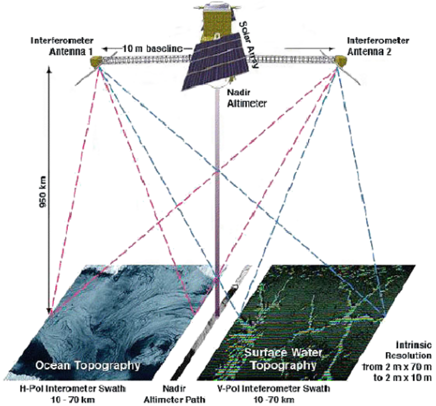
a. Diagnosing surface currents from satellite SSH measurements

- SSH observations allow to retrieve surface currents assuming the geostrophic balance:

$$\mathbf{u}_g = \frac{g}{f}\hat{\mathbf{z}} \times \nabla_h \eta$$

- This framework has served us very well for present altimetry measurements of large-mesoscale eddies with wavelength > 100-200 km.

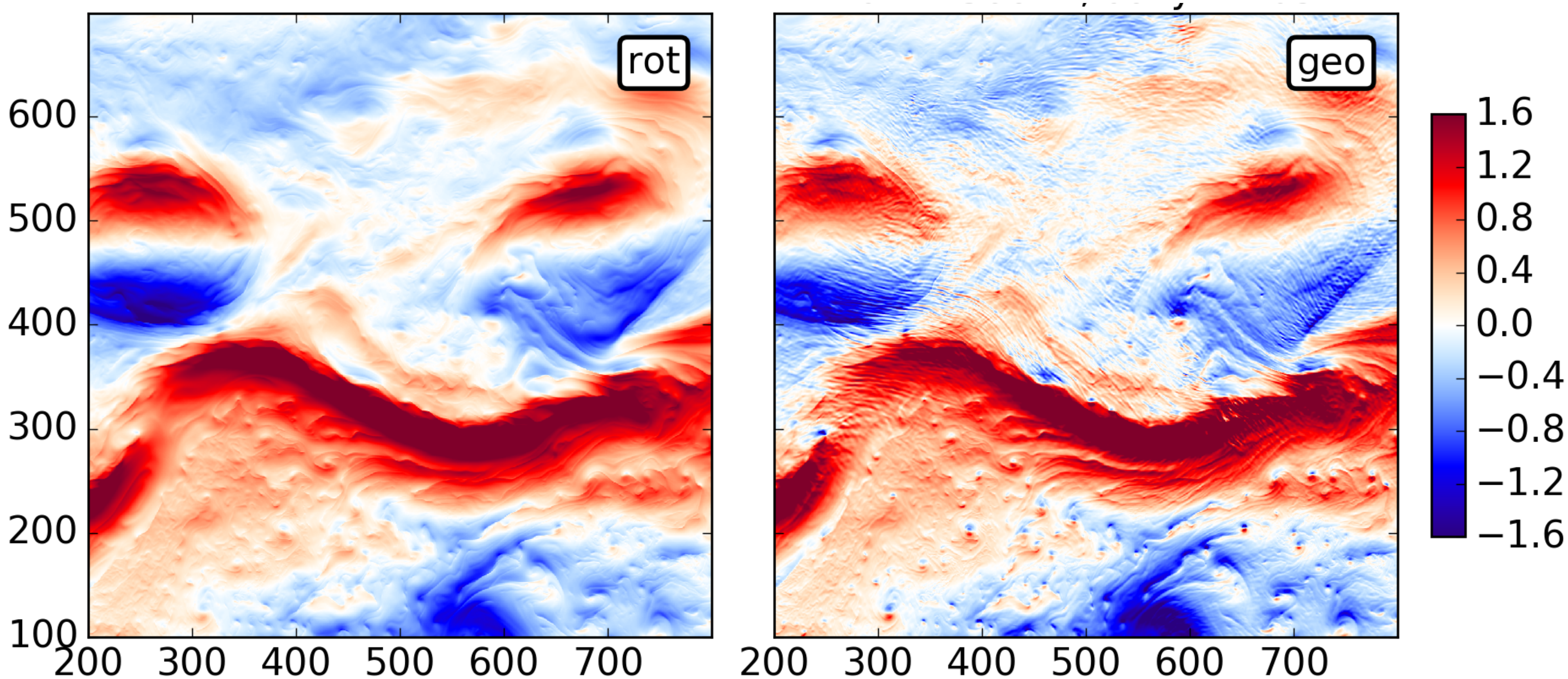
- With **SWOT** (launch 2021) we will see much smaller scales in SSH (~10 km) but we get into submesoscale and internal waves territory and we may need more general surface layer balances.



a. Diagnosing surface currents from satellite SSH measurements

Unbalanced signal in the SSH interferes with the geostrophic velocity computation.

Snapshot from a 500 m resolution simulation over the Gulf Stream



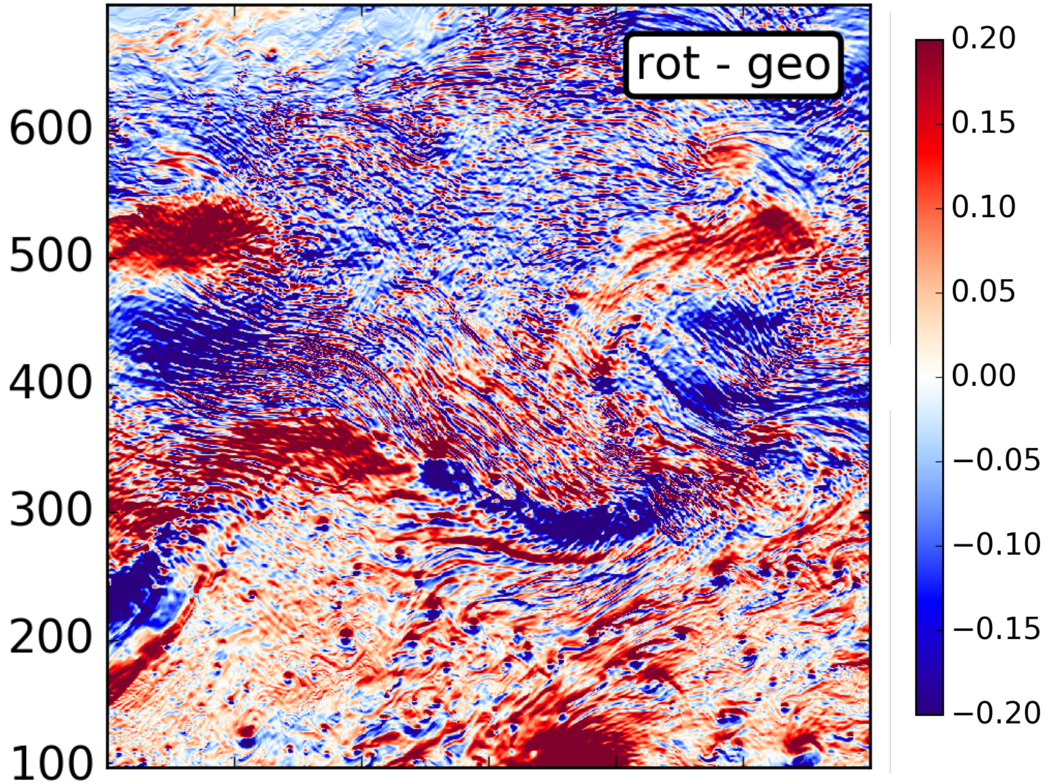
"True" Model velocity (rotational part only)

Geostrophic velocity

$$\vec{u}_g = \frac{g}{f}\vec{k} \times \vec{\nabla}_h \eta$$

a. Diagnosing surface currents from satellite SSH measurements

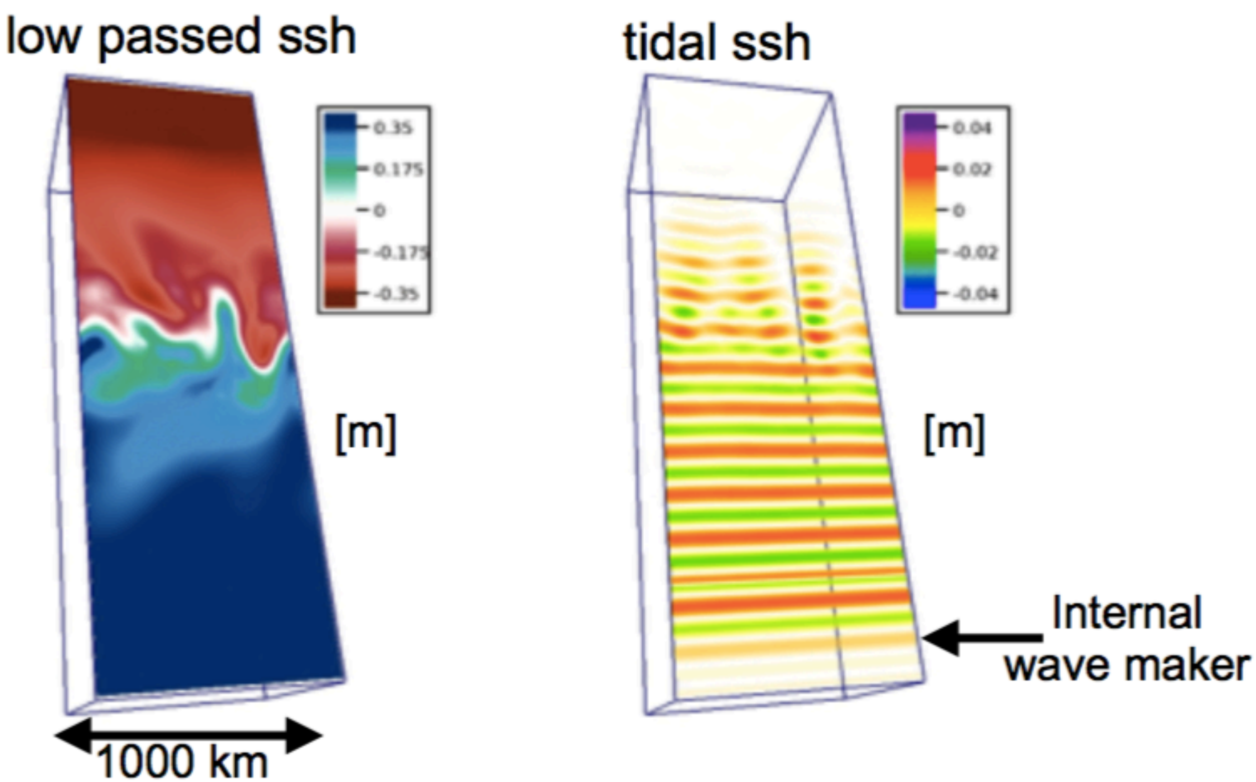
Unbalanced signal in the SSH interferes with the geostrophic velocity computation.



Difference between true and diagnosed geostrophic velocity

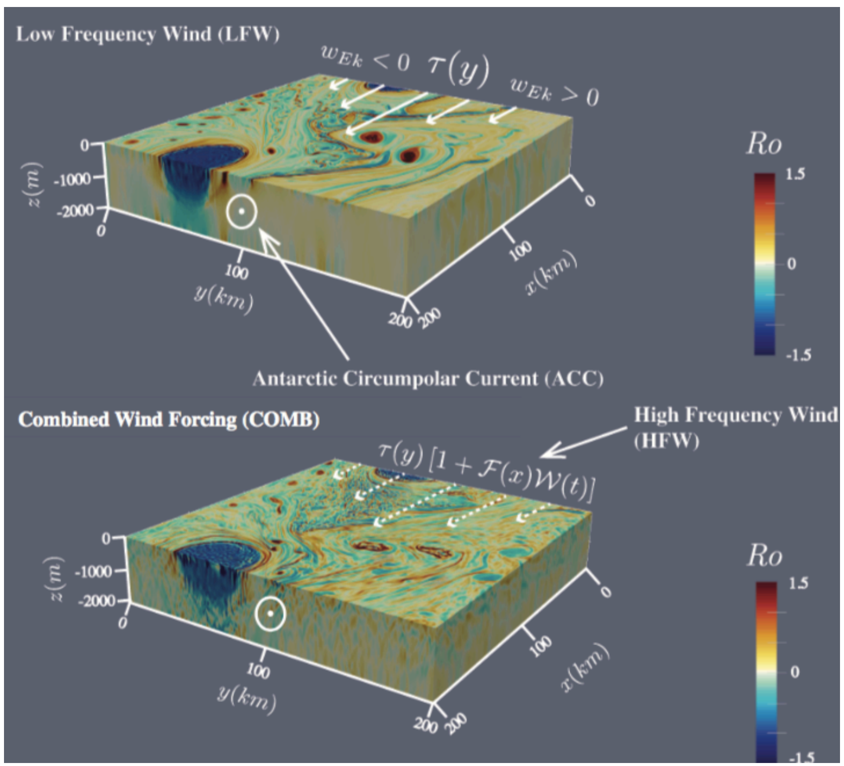
a. Diagnosing surface currents from satellite SSH measurements

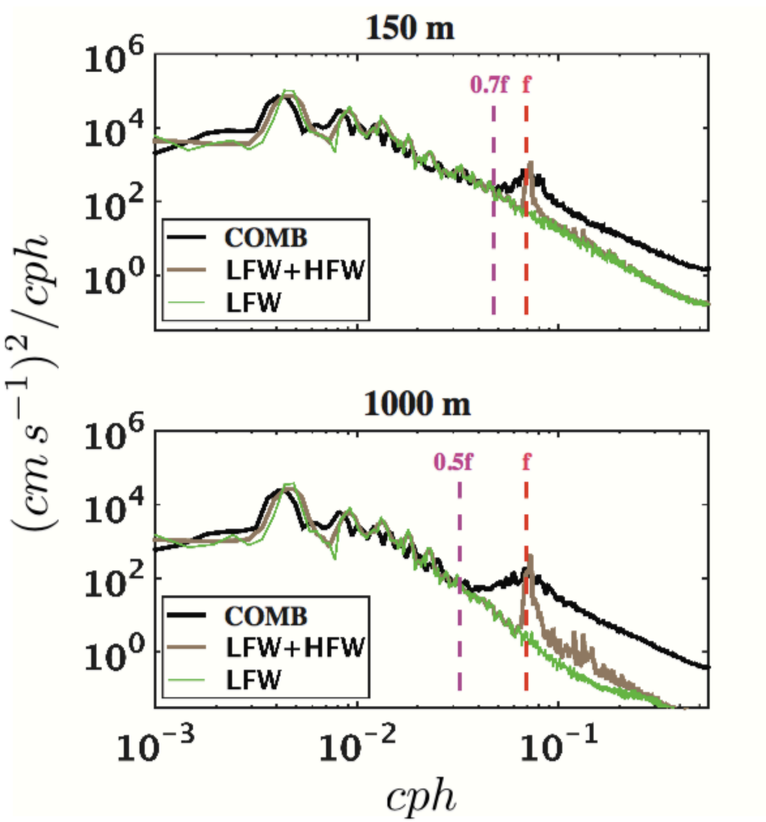
- Emergence of internal tide incoherence in presence of eddy turbulence. In strongly turbulent situations, the internal tide signature on sea level forms complex interference patterns with large amplifications of the initial internal wave **[Ponte & Klein, 2015]**



b. Extraction of energy from the mesoscale by internal waves

Temporal-scale analysis of energy exchanges among low (mesoscale), intermediate (submesoscale), and high (IW) frequency bands shows a corresponding increase in kinetic (Ek) and available potential (APE) energy transfers from mesoscales to submesoscales (stimulated imbalance) and mesoscales to IWs (direct extraction). **[Barkan et al, 2017]**





c. Observation of internal waves using Lagrangian floats

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ARTICLE

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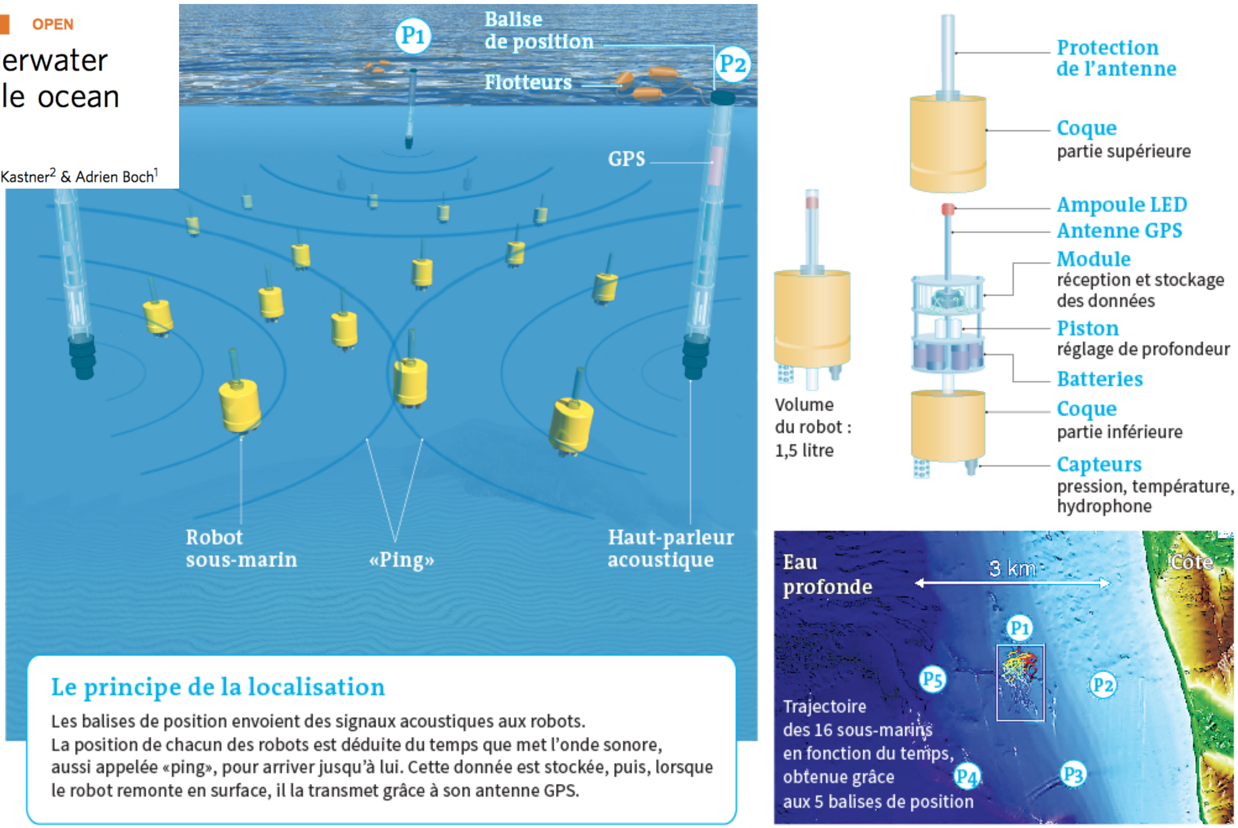
A swarm of autonomous miniature underwater robot drifters for exploring submesoscale ocean dynamics

Jules S. Jaffe[1], Peter J.S. Franks[1], Paul L.D. Roberts[1], Diba Mirza[2], Curt Schurgers[3], Ryan Kastner[2] & Adrien Boch[1]



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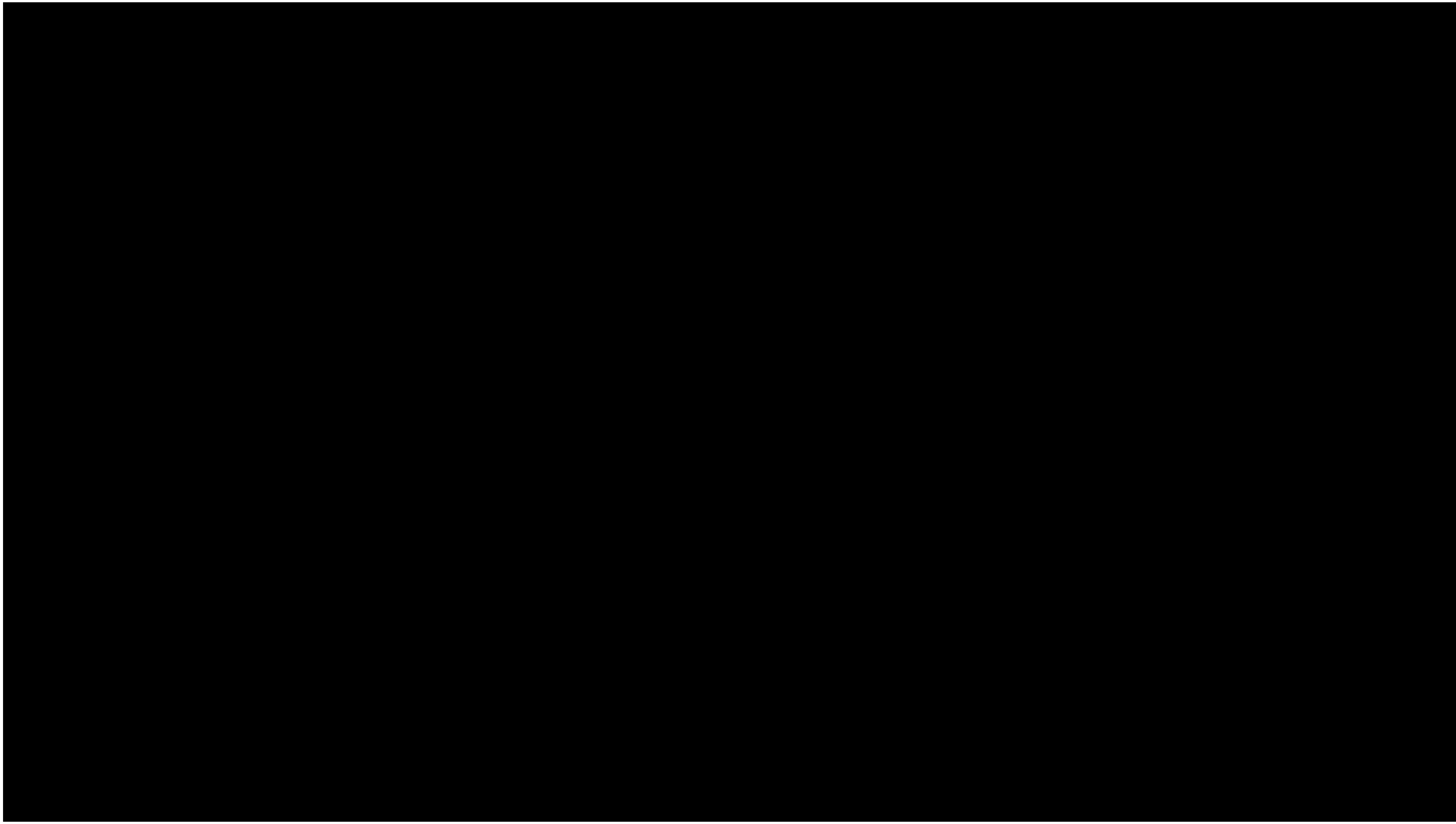
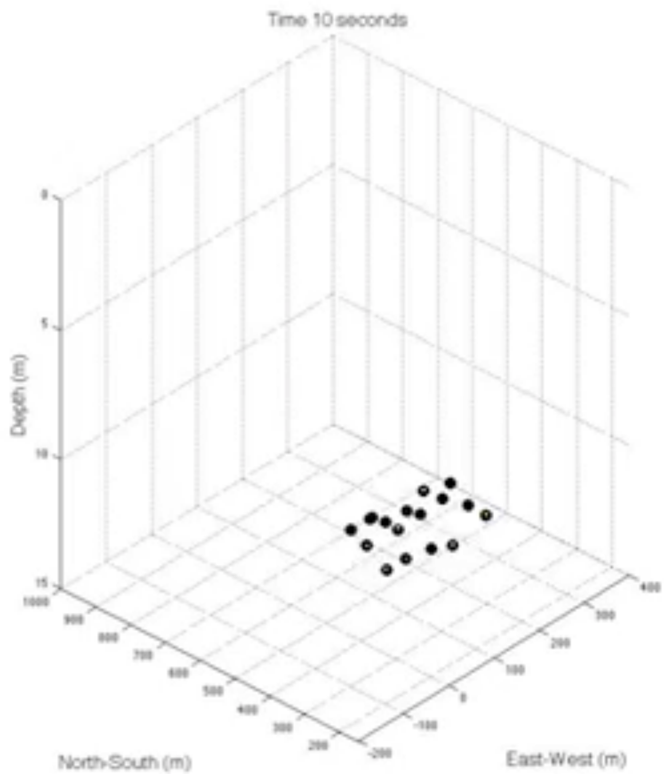
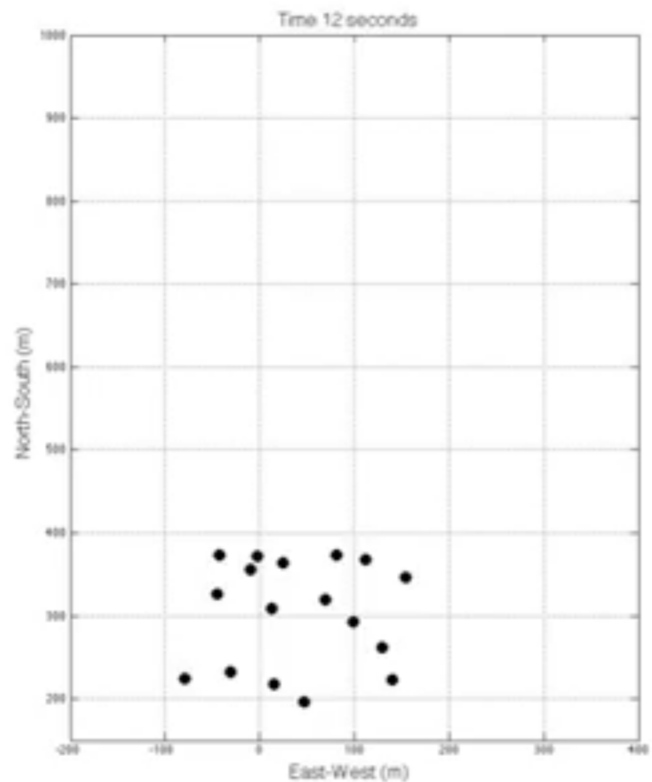
c. Observation of internal waves using Lagrangian floats



c. Observation of internal waves using Lagrangian floats





c. Observation of internal waves using Lagrangian floats

Depth-holding objects in the upper water column should accumulate over the troughs of internal waves (that is, in warm water) and disperse over the wave crests (cold water).

Changes in concentration of the M-AUEs were calculated from changes in the area of the swarm: smaller areas indicate higher concentrations (Fig. 5).

As predicted by theory, the swarm concentrated over wave troughs and dispersed over wave crests.