ENVIRONMENTAL FLUID MECHANICS

Lecture 1 – Introduction

Why do we need to worry about
motion of air and water?

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Air is pure. Air is blue.

The wind brings the oxygen we need.

It lifts birds' wings .

There is no life without water.

Water is healthy.

Water is strong. It lifts boats.

http://www.kathiyavad.com/wallpaper/

But, air can be stormy and menacing.

http://www.sodahead.com/

And, water can be threatening.

http://www.hydrolance.net/

Even deadly!

http://news.nationalgeographic.com/news/2011/03/pictures

11 March 2011 tsunami hitting Miyako City, Japan

Importance of fluids in the environment

Think about it. All living creatures are immersed in a fluid or another, in the *air* of the atmosphere or *water* of rivers, lakes and sea.

Air and water are vital, in the literal sense that life would not be possible without them:

- Air of the atmosphere holds the oxygen that our breathing needs and should not contains certain other things like toxic fumes;

- Water is essential to life on Earth, both inside of our bodies as well as outdoors to grow our foods and much more.

At further thought, it is not the only presence of these fluids that is vital. It is also the fact that they flow:

- Without atmospheric motion around us, we would quickly suffocate in our own cloud of carbon dioxide;

- Thanks to the hydrological cycle, rain and snow provide us with (initially) distilled water; water is also cleansed by passing through forests, wetlands, and underground aquifers.

Flow transport has a number of benefits:

✓ It takes things away, and usually to a larger place,
river to lake, or river to estuary, and estuary to sea,
local airshed to broader atmosphere,
troposphere to stratosphere.

✓ It creates time delay to allow modifications to take place along the way (*ex*: biodegredation of organic wastes down a river, chemical reactions in air).

✓ Fluid motion at human and larger scales is almost always unstable, and the ensuing turbulence generates mixing, which

- dilutes and lowers concentrations, and

- promotes encounters (fluid-fluid, interfaces with other fluid or soil), which in turn promote chemical reactions and transformation.

The study of these flows has received considerable attention over the years, and several distinct disciplines have emerged:

meteorology, climatology, hydrology, hydraulics, limnology, and oceanography.

Whereas the particular objectives of each of these disciplines, such as *weather forecasting* in meteorology and design of *water-resource projects* in hydraulics, encourage disciplinary segregation, environmental concerns compel experts in those disciplines to consider problems that are essentially similar:

- the flow over irregular topography,
- the rise of a buoyant plume,
- the effect of turbulence on the dispersion of a dilute contaminant,
- the transfer of a substance between the fluid and a boundary.

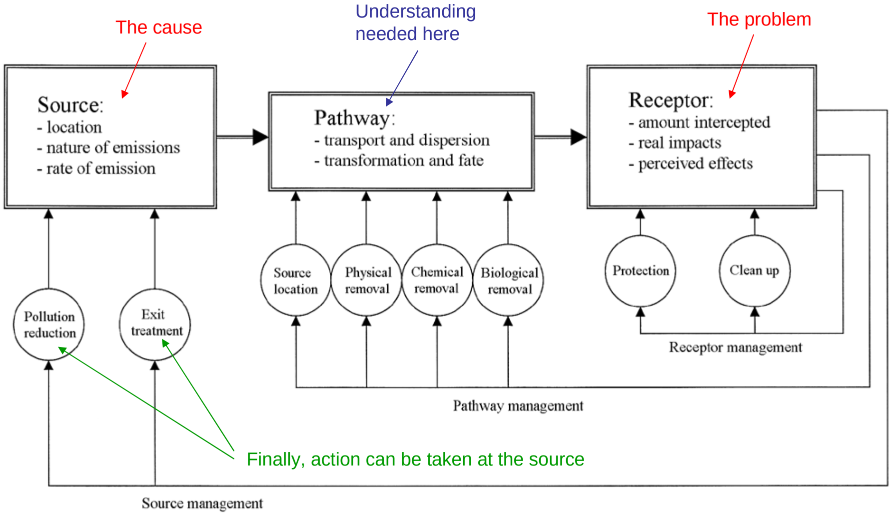
Such common points encourage interdisciplinarity to a degree that is increasing in proportion to the acuity of our environmental problems.

This overlap between the various disciplines concerned with the environmental aspects of natural fluid flows has given rise to a body of knowledge that has become known as *Environmental Fluid Mechanics*.

Environmental Fluid Mechanics in comparison to related disciplines

	Environmental Fluid Mechanics	Fluid Mechanics	Geophysical Fluid Dynamics	Hydraulics	Hydrology
Air example	Sea breeze	Airfoil	Storm	---	---
Water example	Danube River	Pump	Gulf Stream	Dam	Watershed
Turbulence	Beneficial (dilution)	Detrimental (drag)	Secondary importance	Secondary importance	Unimportant
Human control	Limited	Dominant	Nil	Dominant	Limited
Purpose	Prediction & Decision	Design & Operation	Prediction & Warnings	Design & Operation	Prediction & Decision

The generic story behind environmental problems



The cause

Understanding needed here

The problem

Finally, action can be taken at the source

A definition of *Environmental Fluid Mechanics*:

Environmental Fluid Mechanics (EFM) is the scientific study of naturally occurring flows of air and water on our planet Earth, especially of those that affect the environmental quality of those fluids.

In sum, only two fluids, moreover within relatively narrow ranges of pressures and temperatures. What could possibly be so challenging in such a discipline?

Serious difficulties usually arise from

- the turbulent character of the flow (no good physical theory of turbulence),
- the large size the domain (containing a variety of length and time scales),
- the complexity of the domain geometry (irregular topography, bending vegetation, *etc*.),
- complex processes at interfaces (air-water, air-vegetation, water-sediments, *etc*.).

Processes & Systems

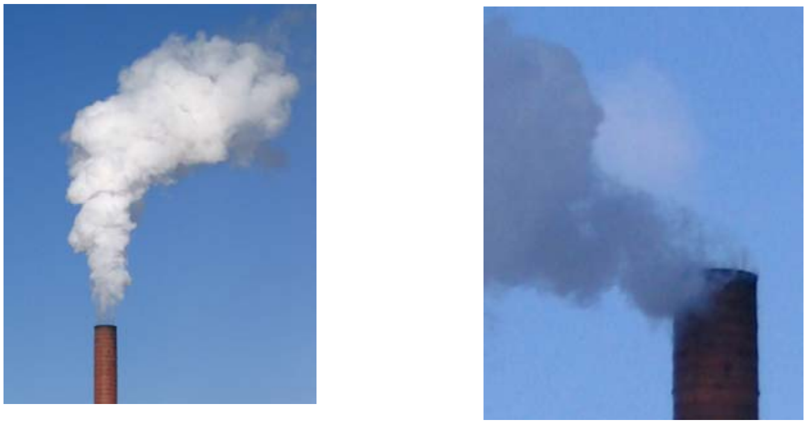
In EFM, it is useful to distinguish between processes and systems.

Processes allow us to consider one type of physics at a time, and the objective is understanding.

Systems consist of multiple processes occurring in one place, and the objective is modeling, prediction and/or decision making.

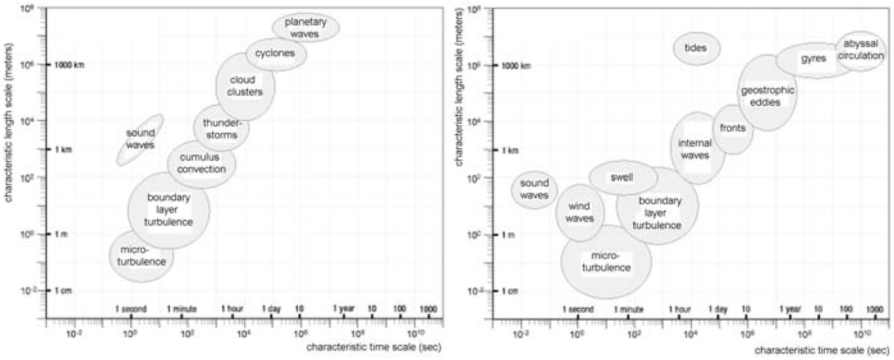
Example of processes	*Examples of systems*
Turbulence	River
Convection	Lake
Waves	Estuary
Instability	Urban airshed
Sea breeze	Atmospheric boundary layer
Baroclinic instability	Troposphere (weather)

An example: The smokestack plume



Five processes can be identified in these photographs:
- vertical rise due to the buoyancy of the hot fumes (left),
- downwash by fast wind creating low pressure downstream of the stack (right),
- horizontal transport by the ambient wind,
- turbulent billowing,
- condensation of water vapor.

Length and time scales
of typical environmental fluid processes and systems



in air

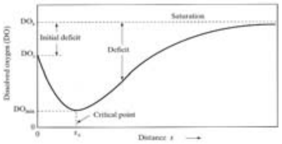
in water

	Horizontal length scale	*Vertical length scale*	*Velocity scale*	*Time scale*
Processes				
Micro-turbulence	1-10 cm	1-10 cm	1-10 cm/s	few seconds
Shear turbulence	0.1-10 m	0.1-10 m	0.1-1 m/s	few minutes
Water gravity waves	1-100 m	1-100 m	1-10 m/s	seconds to minutes
Convection	1-1000 m	1-1000 m	0.1-1 m/s	hours, days or seasons
Atmospheric systems				
Urban airshed	1-10 km	100-1000 m	1-10 m/s	hours
Sea breeze	1-10 km	100-1000 m	1-10 m/s	hours
Thunderstorms	1-10 km	100-5000 m	1-10 m/s	hours
Mountain waves	10-100 km	10-1000 m	1-10 m/s	days
Tornado	10-100 m	100-1000 m	100 m/s	minutes to hours
Hurricane / Typhoon	100-1000 km	10 km	100 m/s	days to weeks
Weather patterns	100-5000 km	10 km	1-10 m/s	days to weeks
Climatic variations	global	50 km	1-10 m/s	decades and beyond
Water systems				
Wetlands	10-1000 m	10 m	0 to 1 cm/s	days to seasons
Small stream	1-10 m	0.1-1 m	1-100 cm/s	seconds to minutes
Major river	10-1000 m	1-10 m	1-10 cm/s	minutes to hours
Lakes	1-100 km	10-1000 m	1-10 m/s	hours (winds) to seasons (heat)
Ocean tides	basin scale	basin depth	0.1-10 m/s	hours
Estuaries	1-10 km	1-10 m	0.1-1 m/s	hours to days
Coastal Ocean	1-100 km	10-1000 m	0.1-1 m/s	few days
Deep ocean	basin scale	basin depth	0.01-1 m/s	weeks to decades

Examples of questions being asked in EFM:

Smokestack plume:
Where should it be located?
How high should it be?

Wastewater discharge jet:
Where may it be placed?
How narrow should be the nozzle?
How fast should the exit velocity be?



Biochemical oxygen demand (BOD) and dissolved oxygen in a river:
Where can a discharge be permitted?
How much can be permitted?

Urban airshed:
How high does the daily atmospheric boundary layer reach?
How much emission can be allowed?
At what point do we get smog?

The two basic ingredients of EFM

1. Turbulence:
   always there
   but of intensity varying with time and place
2. Stratification:
   fluid parcels of different densities, usually due to temperature differences
   can be stable or unstable (heavier fluid below or on top)

Turbulence and stratification interact with each other:

- Turbulence churns the fluid and can diminish or erase stable stratification;
- Heating from above (summer heating of a lake) or cooling from below (warm air passing over cold sea) create stratification, which quenches turbulence.
- Heating from below (atmospheric boundary layer) or cooling from above (lakes in winter) generate reverse stratification, which amplifies turbulence.

Turbulence

Sketch of turbulence by Leornado da Vinci

Turbulence in fluids is measured by the Reynolds number:

$$\mathrm{Re} = \frac{\rho U\, L}{\mu}$$

In environmental fluids, the Reynolds number is always very large ($\mathrm{Re} >> 1$) because of the large scales involved, and turbulence is ubiquitous. Plainly said, there is lots of room and many things are possible.

Because the molecular viscosity μ is of negligible effect, the Reynolds becomes unimportant. Yet, the intensity of turbulence may vary from a situation to another, and its intensity matters because it affects the amount of mixing.

Question: How do we then measure the turbulence level?

Answer:

We need to compare the size of the turbulent velocities to the mean flow velocity

$$\frac{\text{Turbulent velocity}}{\text{Mean flow velocity}} = \frac{u_*}{\bar{u}}$$

What is the turbulent velocity u_*?

Answer: It depends on the nature of the turbulence.

In environmental fluid flows, there are two basic forms of turbulence,

- shear turbulence, created by flow scrubbing against a wall
  (*ex*: wind along ground surface)
- convective turbulence, created by rising/sinking thermals
  (heating from below or cooling from above generating vertical motion).

In **shear turbulence**, the key ingredient is the stress τ along the boundary, and u_* is the defined as

$$u_* = \sqrt{\frac{\tau}{\rho}} \quad .$$

In **convective turbulence**, the key ingredient is the vertical velocity of the rising / sinking thermals, denoted w_* and obtained from the conversion of potential energy into kinetic energy production

$$w_* = \left(\frac{\alpha\, g\, h Q}{\rho C_p} \right)^{1/3}$$

where ρ = fluid density, C_p = heat capacity (at constant pressure), α = thermal expansion coefficient, g = gravitational acceleration, h = height, and Q = heat flux per horizontal area (in W/m^2).

Stratification

Two cases: Either the stratification is unstable (heavier fluid on top)
or it is stable (heavier fluid at bottom).

In the case of unstable stratification, turbulence is enhanced. The question becomes which type dominates. Is it the u_* from mechanical shear or w_* from thermal convection? This reduces to comparing

$$u_* = \sqrt{\frac{\tau}{\rho}} \quad \overset{?}{\leftrightarrow} \quad w_* = \left(\frac{\alpha\, g\, h\, Q}{\rho\, C_p} \right)^{1/3}$$

In the atmospheric boundary layer (first 500 to 1000 m of atmosphere above ground), mechanical turbulence dominates at high winds and during cloudy days (wind-induced shear turbulence generated at ground level helps mix the atmosphere vertically), while convective turbulence dominates at low winds and on sunny days (rising thermals create more turbulence than the wind scrubbing against the ground surface).

The vertical height h over which u_* and w_* are equal is called the Monin-Obukhov length, usually denoted by L :

$$\begin{aligned} u_* &= w_* \\ &= \left(\frac{\alpha\, g\, L\, Q}{\rho\, C_p} \right)^{1/3} \quad \rightarrow \quad u_*^3 = \frac{\alpha\, g\, L\, Q}{\rho\, C_p} \end{aligned}$$

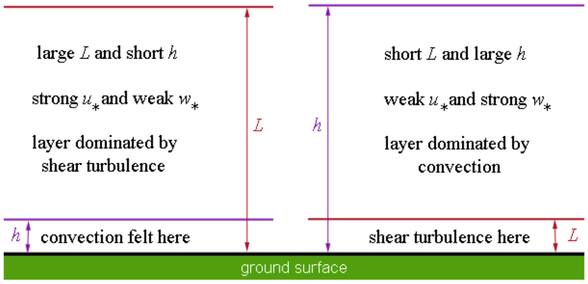
$$\downarrow$$

h replaced by L here

$$L = \frac{\rho\, C_p\, u_*^3}{\alpha\, g\, Q} = \frac{u_*^3}{\alpha\, g < w'T' >}$$

The actual height h of the system generally differs from the Monin-Obukhov length L .
The question then arises as to which is the greater and which is the shorter.

In the lower atmosphere, in the presence of a wind scrubbing against the ground and warming from below



large L and short h

strong u_* and weak w_*

layer dominated by shear turbulence

L

h

convection felt here

short L and large h

weak u_* and strong w_*

layer dominated by convection

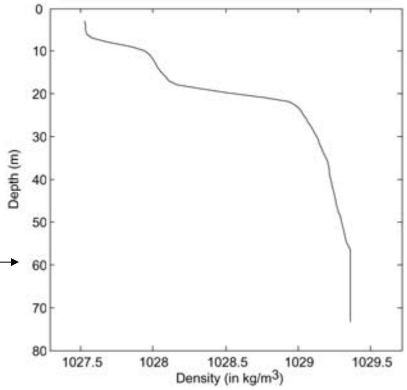
shear turbulence here

ground surface

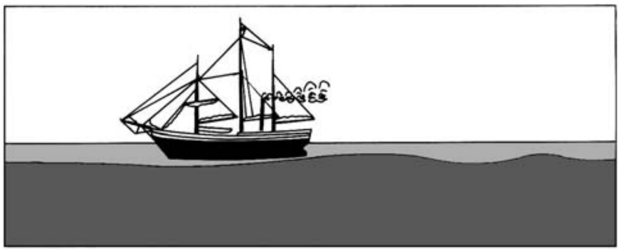
where $$u_* = \sqrt{\frac{\tau}{\rho}} \qquad w_* = \left(\frac{\alpha\, g\, h Q}{\rho C_p}\right)^{1/3} \qquad \frac{L}{h} = \frac{u_*^3}{w_*^3}$$

In considering the mixing of pollutants, it is necessary to know whether this dilution is mostly caused by the wind (*left case*) or by the convection (*right case*).

Oftentimes, the stratification is in the direction of the warmer (hence lighter) fluid on top of the colder (denser) fluid.

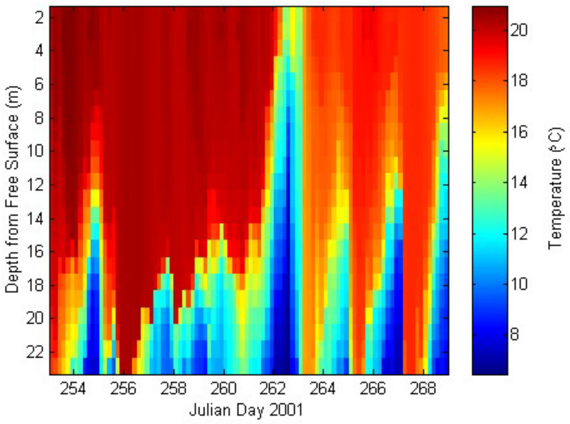


Vertical profile of density in the Northern Adriatic Sea on 27 May 2003. Density increases downward by leaps and bounds, revealing the presence of different water masses stacked on top of one another, with lighter waters floating on top of heavier waters.



A laboratory experiment by Vagn W. Ekman in 1904 to explain the phenomenon of dead waters with internal waves produced by ship motion and causing drag on the ship.

Sometimes, the thermal stratification is subject to large vertical oscillations.



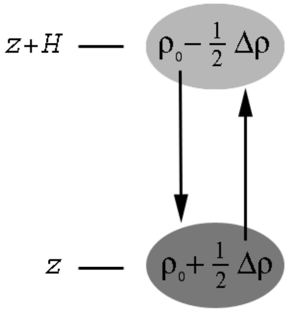
Vertical oscillations in Cayuga Lake during summer stratification (Courtesy of Prof. E. Cowen, Cornell U.)

How can large vertical oscillations take place against buoyancy?

Stable stratification

A stable stratification occurs when the lighter fluid lays on top of the heavier fluid, as gravity wants to have it.

Since this represents a state of lowest potential energy, energy needs to be spent to create vertical motion, because vertical displacement demands that work be done against buoyancy forces.



Take for example the exchange two fluid parcels of same volume V but of different densities and at different levels.

The potential energy increases by

$$\left(\rho_o + \frac{1}{2}\Delta\rho\right) V\, g\, H - \left(\rho_o - \frac{1}{2}\Delta\rho\right) V\, g\, H = \Delta\rho\, V\, g\, H$$

and this needs to occur at the expense of kinetic energy. (There is no other stirring mechanism in the environment than ambient turbulence.)

If the kinetic energy level is $2 \times \frac{1}{2} m U^2 = \rho_o V U^2$ then, the ratio

$$\frac{\Delta\rho V\, g\, H}{\rho_o V U^2} = \frac{g\, H\, \Delta\rho}{\rho_o U^2}$$

expresses the energy need over the energy available.

This ratio is called the Richardson number:

$$Ri = \frac{g\,H\,\Delta\rho}{\rho_o\,U^2} = \frac{\text{Potential energy change}}{\text{Kinetic energy available}}$$

It has the following interpretation:

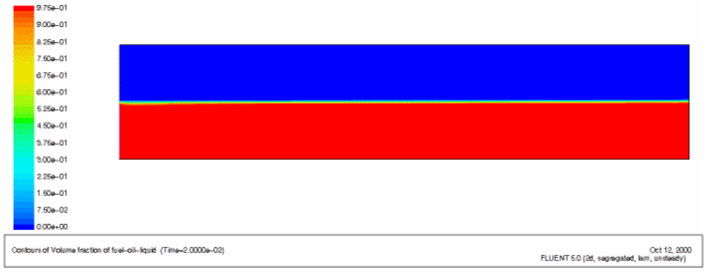
- If Ri is much less than unity,
  kinetic energy supply is abundant, and
  vertical motions are only weakly inhibited by the stable stratification.

- If Ri is on the order of unity,
  kinetic and potential energies are comparable, and
  vertical motions are significantly affected by the stratification.

- If Ri is much greater than unity,
  potential energy demand exceeds kinetic availability, and
  vertical motions are very restricted by the stratification.

A nice example of this is the Kelvin-Helmholtz instability

(Photo by jean-Marie Beckers)



(http://www.enseeiht.fr/hmf/travaux/CD0001/travaux/optmfn/hi/01pa/hyb72/kh/kh_theo.htm)

An example of unwanted stratification

Smog over Los Angeles, California
Note the horizontal layering of the haze

Another example of stable stratification, typical of early morning.

The lower atmosphere was cooled during the night by radiative heat loss to space, creating coldest air nearest the ground.

The smokestack plume has initial buoyancy, rises and overshoots its equilibrium level to ultimately settle at the level where its density matches the ambient air density.

Note that in the previous developments, the following combinations of variables occurred:

$\frac{\alpha\, g\, Q}{\rho\, C_p} = \alpha\, g < w'T' >$ which includes a factor $\alpha\, g\, \Delta T$.

This factor is related to the density difference by virtue of the equation of state, which expresses thermal expansion (decrease of density with increased temperature):

$$\rho = \rho_o\,[1 - \alpha\,(T - T_o)] \quad \rightarrow \quad \Delta\rho = -\rho_o\,\alpha\,\Delta T$$

Thus, the temperature differences are important because they generate density differences.

The factor g in front tells that, in turn, these density differences are important insofar as they create buoyancy forces.

This leads us to defining the buoyancy: $b = \alpha\, g\, T_{\text{anomaly}} = \alpha\, g\, \Delta T = -\frac{g\,\Delta\rho}{\rho_o}$

Equation of state for thermal stratification

$$\rho = \rho_o\,[1 - \alpha\,(T - T_o)] \quad \rightarrow \quad \Delta\rho = -\rho_o\,\alpha\,\Delta T$$

with

AIR at standard temperature (15°C) and pressure (1 atm = 101.33 kPa):

$\rho_o = 1.225$ kg/m^3
$\alpha = 3.47 \times 10^{-3}$ °C^{-1}

FRESHWATER at standard temperature (15°C) and atmospheric pressure:

$\rho_o = 999$ kg/m^3
$\alpha = 2.57 \times 10^{-4}$ °C^{-1}

SEAWATER at standard temperature (10°C) and salinity of 35 ppt:

$\rho_o = 1027$ kg/m^3
$\alpha = 1.7 \times 10^{-4}$ °C^{-1}

Other kinds of stratification in environmental fluids

In water

Salt: Estuaries are places of encounter between salt-heavy waters and relatively lighter freshwater from a river.

↓ Salinity

Sediments: Near the bottom of rivers, the bottom stress often entrains lose particles from the sediments, in a process called erosion.

↓ Turbidity

In air

Water vapor: Moist air is denser than dry air.

↓ Humidity