MTI MET Gr-II

Advanced Training

Physical oceanography &

Air sea interaction.

2016 May-June.

↳ Oceanography

↳ Air-sea interaction

↳ Turbulence &

boundary Layer.

Physical Oceanography ⊃ Ocean atmos. interaction.

22nd Batch Sem.II Met-Gr.II training.

May-Jun 2016

Topics to be covered.

* The significance of air-sea interaction
* Concept of Boundary Layer
  * Ocean
  * Atmosphere.
* SST, air-temp, wind profiles
* Boundary layer fluxes.
* Turbulence

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* Vertical transport in Frictional B.L.
  * Direct Method
  * Aerodynamic method
  * Bulk method
  * Budget method
  * Bowen's Ratio.
  * Surface stress, drag.

Books / References.

① Ch-2, Gill. A.E., 1982, Atmosphere-Ocean Dynamics, Int. Geoph. Series., Vol. 30

② Ch-1,2,3 & 4, Kantha L.H, C.A. Clayson, Small scale processes in Geophysical Fluid flows., Int. Geoph. Series, Vol. 67.

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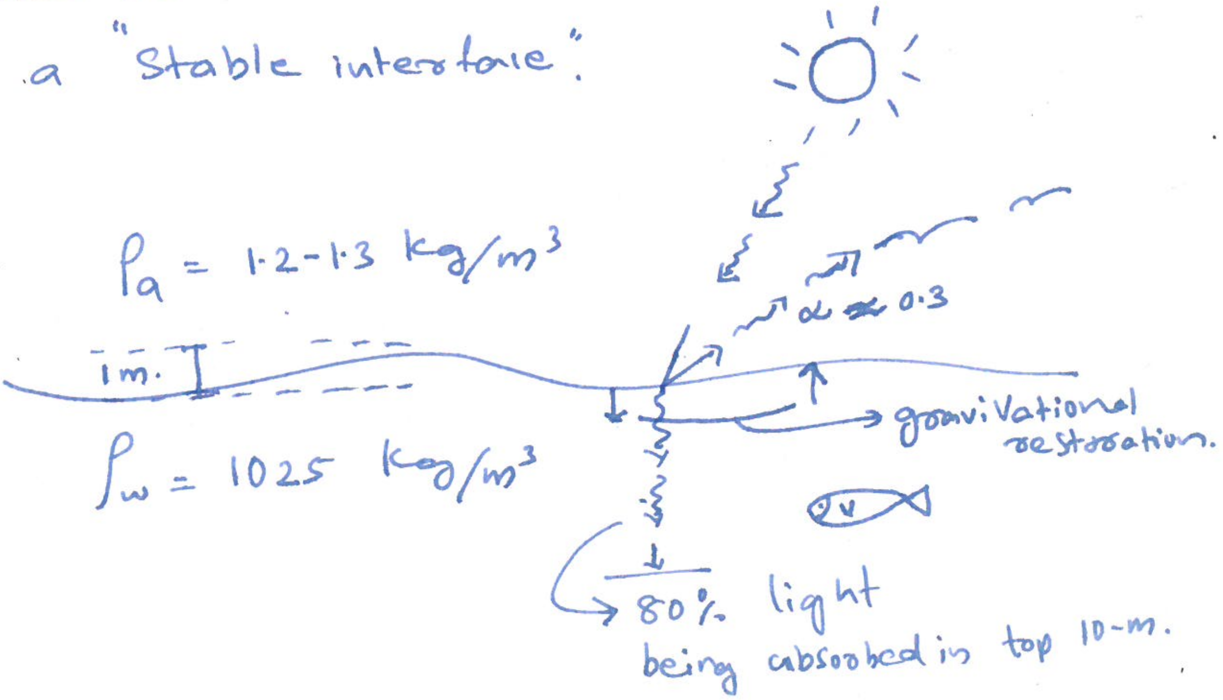
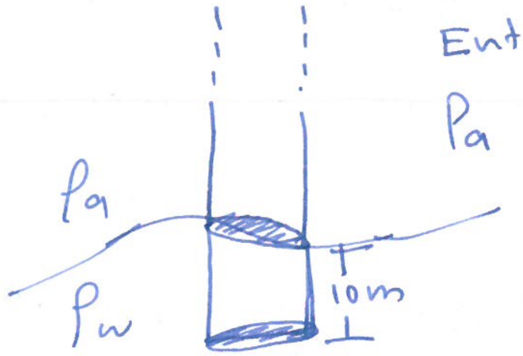
* Significance of air-sea interaction.

The Interface between air and water is a "Stable interface".



Atmospheric Pressure and Oceanic Pressure.



Entire weight of atm/unit area

$P_a = 10^5 \, N/m^2 = 1$ bar.

* A 10-m depth of ocean has the same weight/unit area, So the pressure increases by 1 bar every 10 m.

* For this reason Oceanographers express Ocean pressure in "deci-bars".

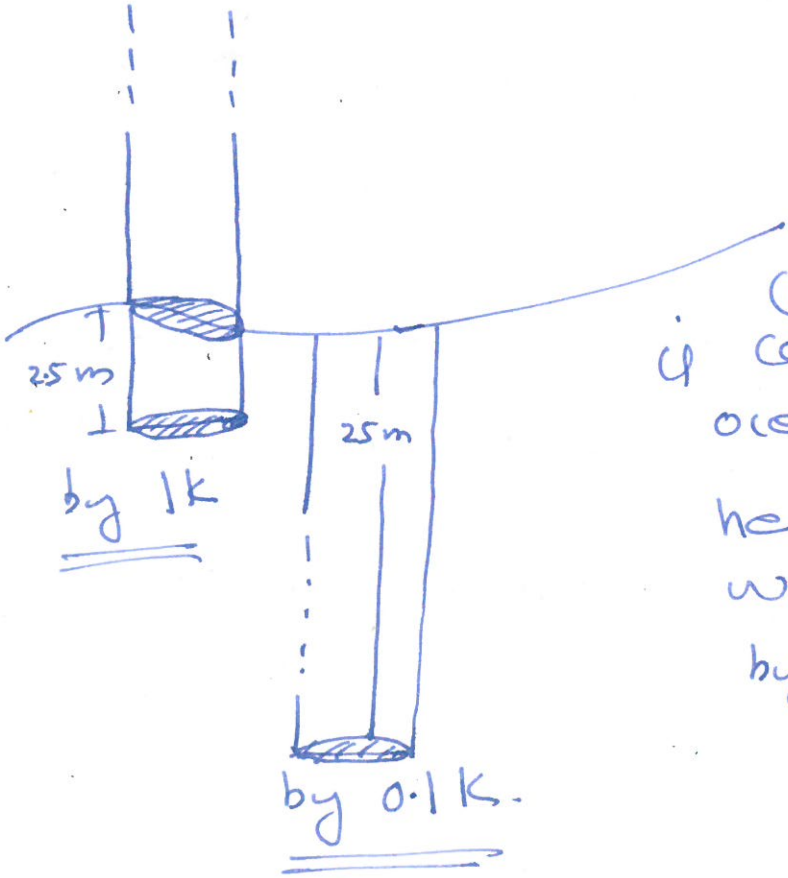
1 db = 1 m water weight.

10 db = 1 bar.

* Heat capacity of water is nearly 4 times as that of air.

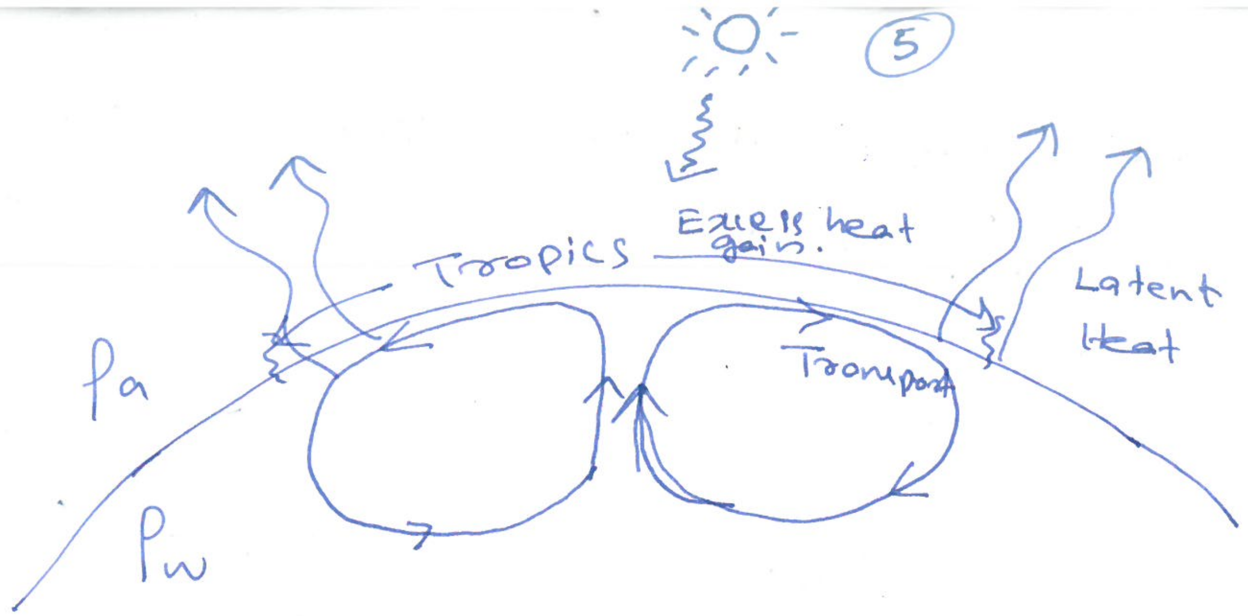
* A 2.5 m depth of water ~~over earth Surface~~ has the same heat capacity per unit area as the whole depth of the atmosphere.

* Therefore Heat required to warm the entire column of atmosphere per unit area by 1K can be obtained by changing the temperature of 2.5 m of water by the same amount.



(0.1K)
if cooling of 25 m of ocean by. air-sea heat exchange can warm the atmosphere by (1K).

* Large heat capacity of the ocean is important for seasonal change.



~~* Momentum~~

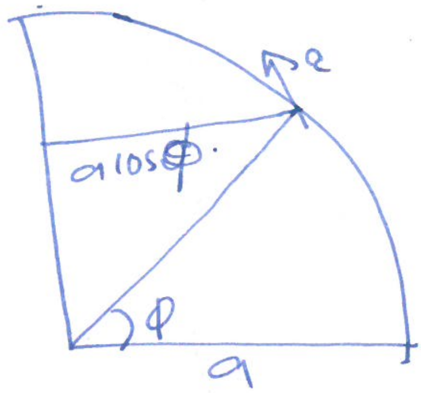
* Therefore, water & air has distinct features intrinsic to the system such as density, optical properties, heat capacity etc

* Momentum Transfer between Air and sea.

Suppose the average eastward force per unit area (wind stress) acting earth's surface at latitude "ϕ" is

$\tau^x(\phi)$.

Then the rate of transfer of angular momentum per unit area about earth's axis



$$= r \times F$$

$$= a\cos\phi \, \tau^{\eta}(\phi)$$

where "a" is the radius of earth.

The area of zonal strip between ϕ and $\phi + d\phi$ is

$$2\pi \cdot a^2 \cos\phi \cdot d\phi.$$

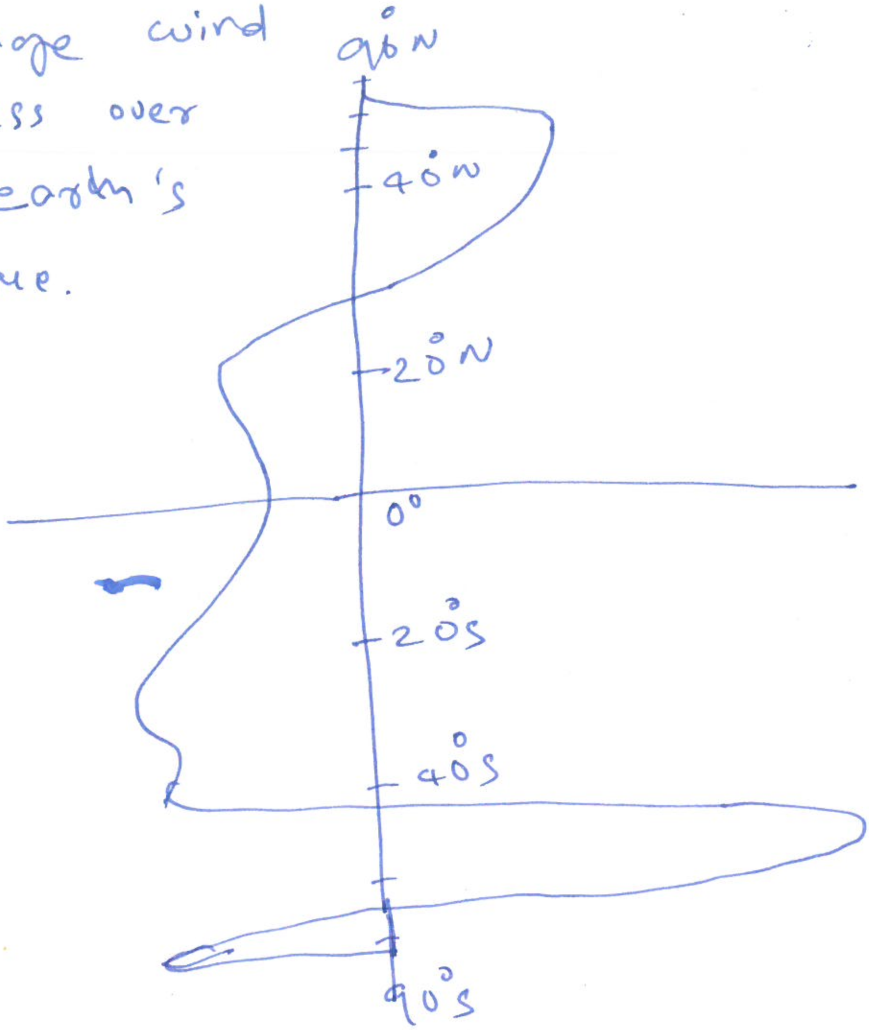
Therefore the total ~~stress~~ angular momentum on the strip

$$\int_{-\pi/2}^{\pi/2} = 2\pi a^3 \tau^{\eta}(\phi) \cos^2(\phi) \, d\phi.$$

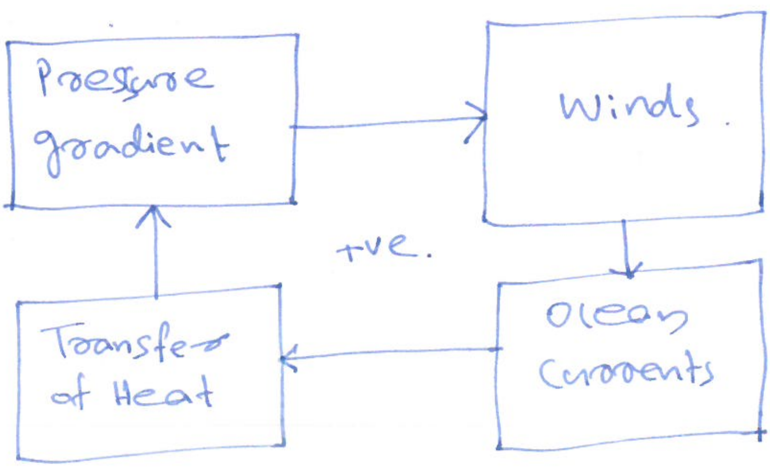
The net torque on earth's surface must vanish

$$\varphi \int_{-\pi/2}^{\pi/2} 2\pi a^3 \tau(\phi) \cos^2(\phi) \, d\phi = 0.$$

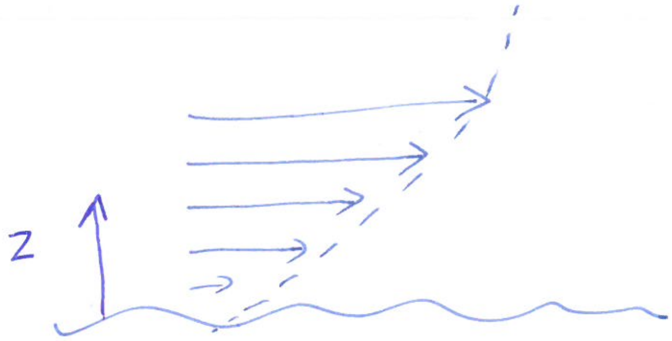
Observed zonal average wind stress over the earth's surface.



* Dependency of exchange rate on air velocity, temperature and humidity.



Exchange of momentum across the atmospheric boundary layer

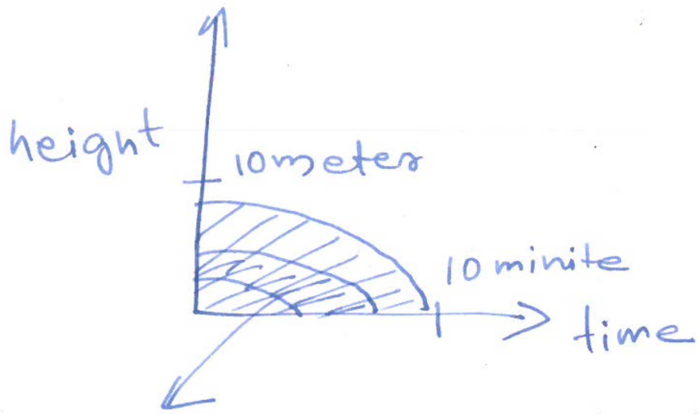


↳ The shear nature of the wind produces Turbulence

↳ The turbulence is an unstable condition

↳ The turbulent eddies (chunk of air/water) modifies the shear.

↳ But over a significantly long time, a well defined mean velocity can be determined for each value of z; the distance above the ground.



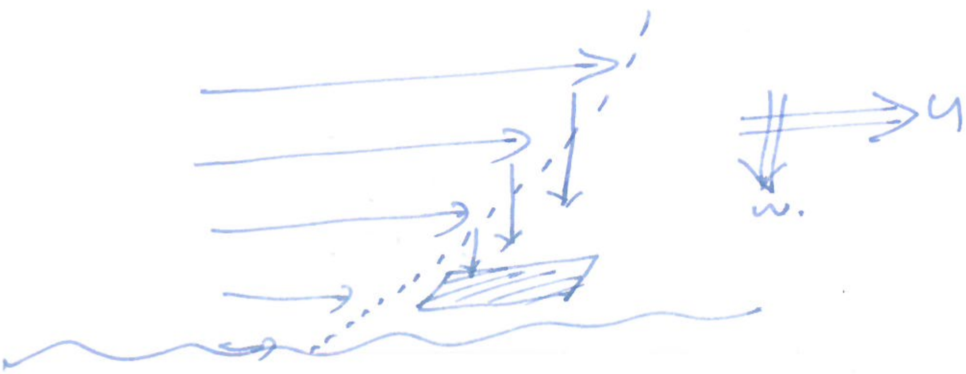
↳ Averaging of wind be at 10meter in space and 10 min. in time.

↳ $$\bar{u} = \sum_{z=0}^{10m.} \sum_{t=0}^{10min} \frac{u.}{N}$$

Then turbulent velocity

$$u' = \bar{u} - u$$

↳ The vertical rate of transport of zonal momentum gives the measure of exchange of momentum between the air - ocean.



* If "ρu" is the mass flux,

i.e. vertical transfer of zonal momentum $= \rho u w =$ which is equal to the zonal stress.

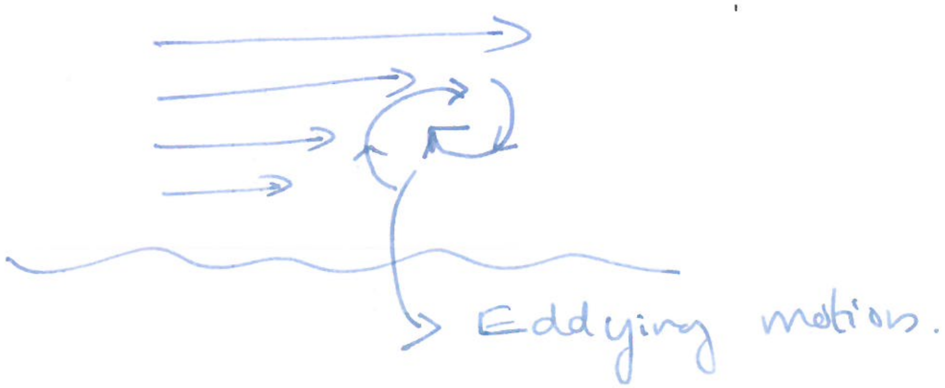
* By diamensional argument, the viscocity of the fluid (air/water)

$$\rho u w \propto \frac{\partial \bar{U}}{\partial z}.$$

wher $\bar{U}$ is the mean background flow.

$$\text{i.e. } \tau_x = \rho u w = \left[K_z \frac{\partial \bar{U}}{\partial z} \right]$$

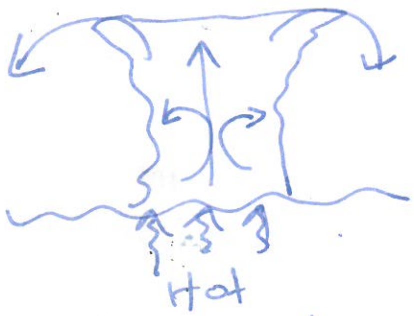
where K_z is "analogous to kinematic viscocity of the fluid but in a large quantity it is called as eddy diffusion/viscocity.



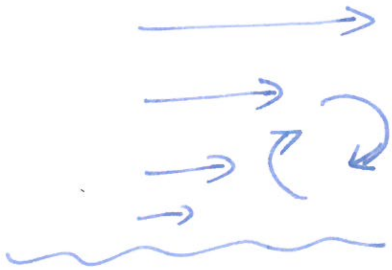
* Because of the eddying nature of the flow, the temperature and mass are transferred by air movement and that depend on wind speed

* The "molecular" anology is with the "eddies".

Turbulence due to convection
Turbulence due to shear



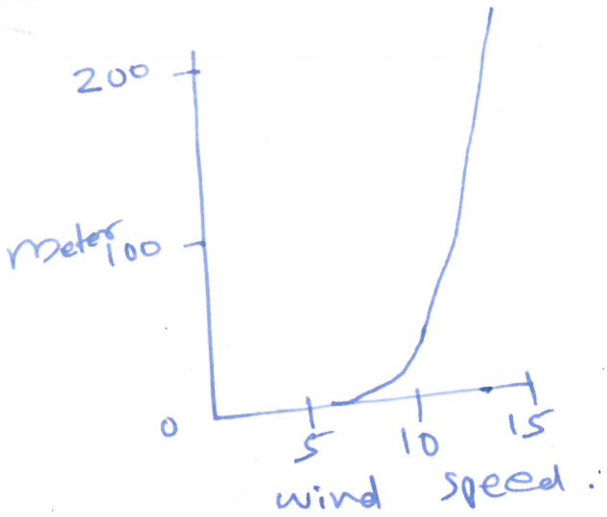
Turbulence due to convection



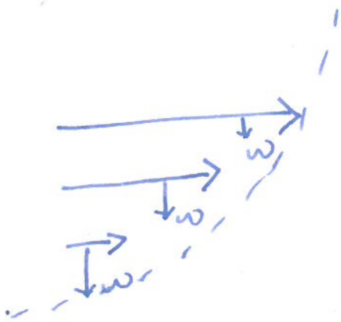
Turbulence due to shear.

* The length at which the turbulence due to convection (buoyancy) is more or equal by those due to wind shear is an important parameter → Monin-Obukhov length.

Proflile of wind close to surface



* As the ground approached, the shear increases in inverse proportion to the distance from the ground.



$w \approx C_D \cdot u$

* The relationship between τ and u can be approximated as

$$\tau = \rho u w$$

$$= \rho u \cdot (u C_D)$$

$$\tau = C_D \cdot \rho \cdot u^2$$

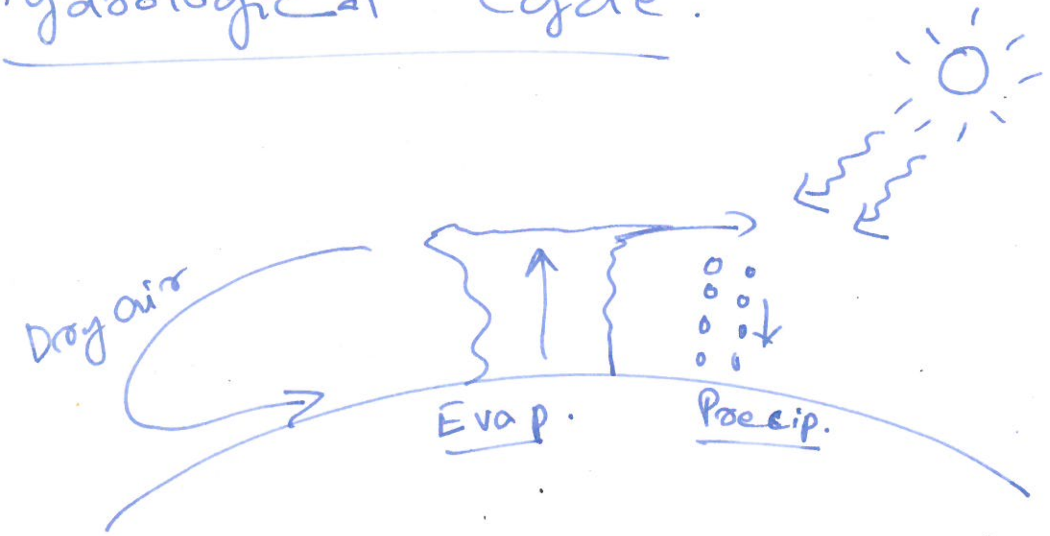
For $u < 6\,m/s$,

$C_D \approx 1.1 \times 10^{-3}$

For u ove $6\,m/s$

$10^3 C_D = 0.61 + 0.063u$

Hydrological Cycle.



↳ At tropic water evaporates.

↳ $\approx 1\,m/year$ (i.e. $3\,mm/day$)

↳ If the entire water in the atmosphere ~~to~~ precipitate, it would cover earth surface by a depth of 2.3 cm.

Water reservoirs of earth surface

Ocean-atmosphere Boundary Layer

* This is composed of Atmospheric boundary layer (ABL) and Ocean mixed Layer (OML)

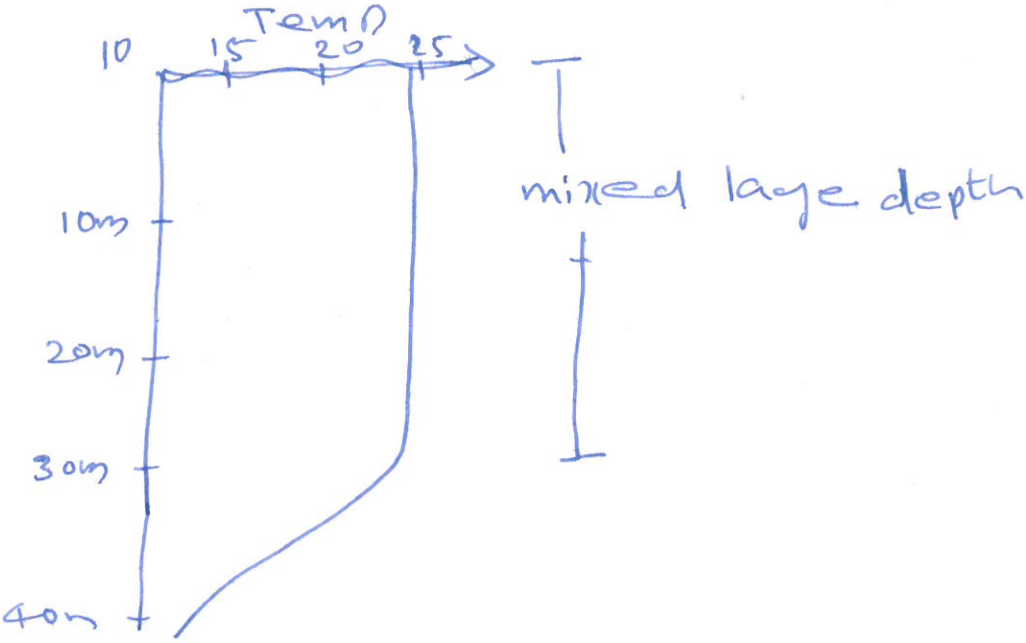
* The mediate the exchange of mass, momentum, energy, and heat. between Atmosphere and Ocean

* This is the principal reason why the OML/ABL is central to air-sea exchange, a problem of importance to long-term whether D climate.

*

Ocean Mixed Layer (OML)

Definition: OML is the well mixed Layer in the surface ocean of few tens of meters thick where water properties such as temperature, salinity and other micro and macro nutrients are well mixed and at uniform concentration or values.



* Various Criteria of "OML" definition.

↳ Temperature Criteria

↳ Density Criteria

* [Lecture Notes of Physical Oceanography be placed/added here].

* Although the "OML" defined by simple criteria above and "OML" is considered as a "uniform" property/characteristic region of the upper ocean, they are still composed of or "OML" is characterized as follows

↳ OML can be ~~de~~ divided into four pars

↳ A very thin but important molecular sub layer of a few millimeter thick at the surface

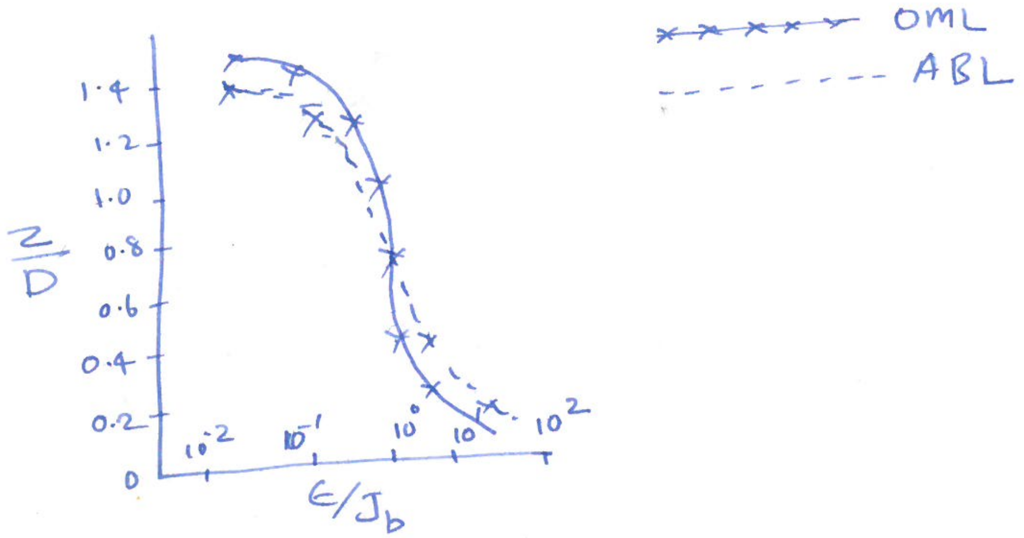
↳ A wave- sub layer of 2-6m thick

↳ The main or bulk of OML 10-40m thick

↳ The interfacial layer or entrainment sub-layer of 5-10m thick at the bottom of OML.

Similarities and differences between Atmospheric boundary Layer and OML.

Striking Similarity between ABL and OML



Where D = ABL depth or OML depth

z = vertical height co-ordinate

ϵ = dissipation rate of Turbulent kinetic Energy

J_b = Buoyancy flux (sum of all heat and mass fluxes).

Differemes between ABL and OML

OML	ABL
① Typically tens of meters deep (20- 100 m)	Hundreds of meters deep (1-2 km)
② Shear mixed during day and night albiet weak convective mixing during night.	Convectively mixed during day and shear mixed during night
③ Heated from above	Heated primarily from below
④ cooling is by evaporative loses.	Entire atmosphere cools during absence of sun-light
⑤ Seasonal modulation is very significant and Large	seasonal modulation is insignificant except at poles where length of the day is largely different between ~~&~~ winter & sumer

* Before Proceeding to ABL, a few excerpts of Turbulance theory is given below.

* Scales of turbulance

Integral length scale.

↳ Turbulance is highly a chaotic phenomenon and operates over a variety of length or dimensions. The size and time involved in turbulence characterizes the features of typical turbulent phenomenon.

↳ Therefore various length scales are associated with Turbulence for their descriptions.

↳ There are three Integral length scales of most importance, ① Macroscale "Λ", ② Microscale or commonly known as Kolmogoroff's microscale and ③ Taylor's microscale "λ".

↳ All these scales are for ~~[illegible]~~ Nuetrally stratified stream line flows.

↳ Therefore certain analogy has to be made for Atmospheric and oceanic boundary layer.

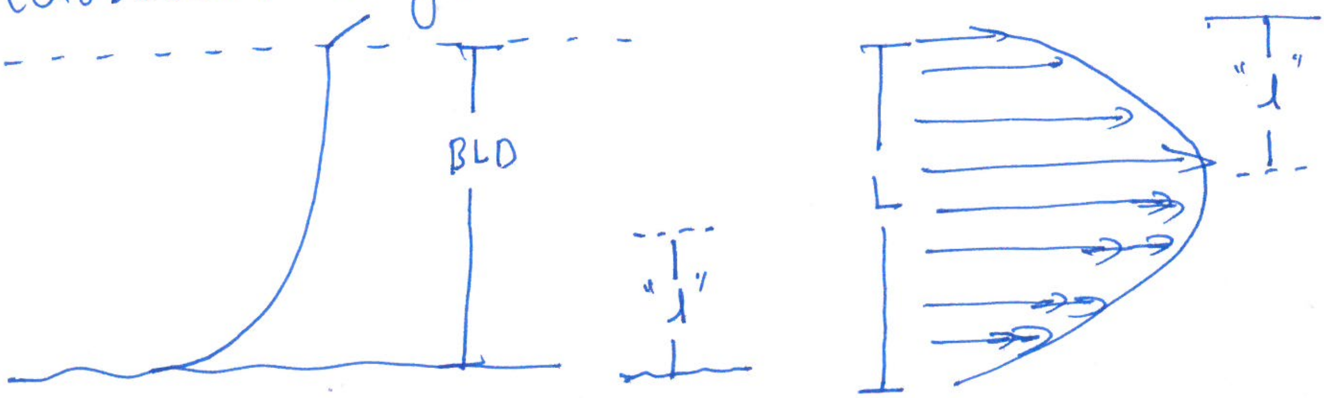
① Macroscale "l".

↳ "l" denotes the length scale over which the turbulent quantities are self-correlated.

↳ It also signifies the scale of most energetic eddies

↳ The microscale "l" depends on the physical size of the turbulent region.

①

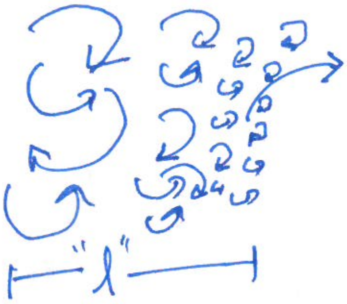


↳ It is approximately half of the ~~the~~ dimension of the boundary layer.

↳ The importance of "l" is that it essentially determine the dissipation rate of the flow.

↳ Within the macroscale "l", the large eddy loses energy to smaller scales.

Large eddy region



small eddy region where most of the dissipation is concentrated,

↳ The "Raynold's Number" indicate the macroscale intensity of mixing $\left[R_t = \frac{q\,l}{\nu} \right]$.

where q = velocity scale
l = macroscale "l"
ν = kinematic viscosity.

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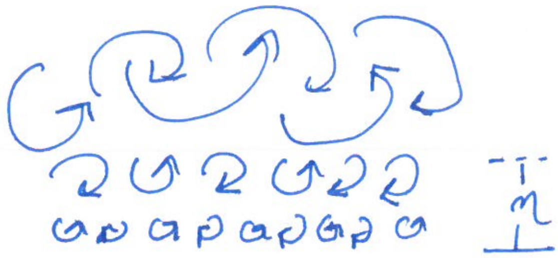
② Kolmogoroff's micro scale

↳ This scale determines the size of the "smallest possible eddies" in the flow.



$$\eta = \left[\frac{\nu^3}{\epsilon}\right]^{1/4}$$

where ν = kimematic viscosity

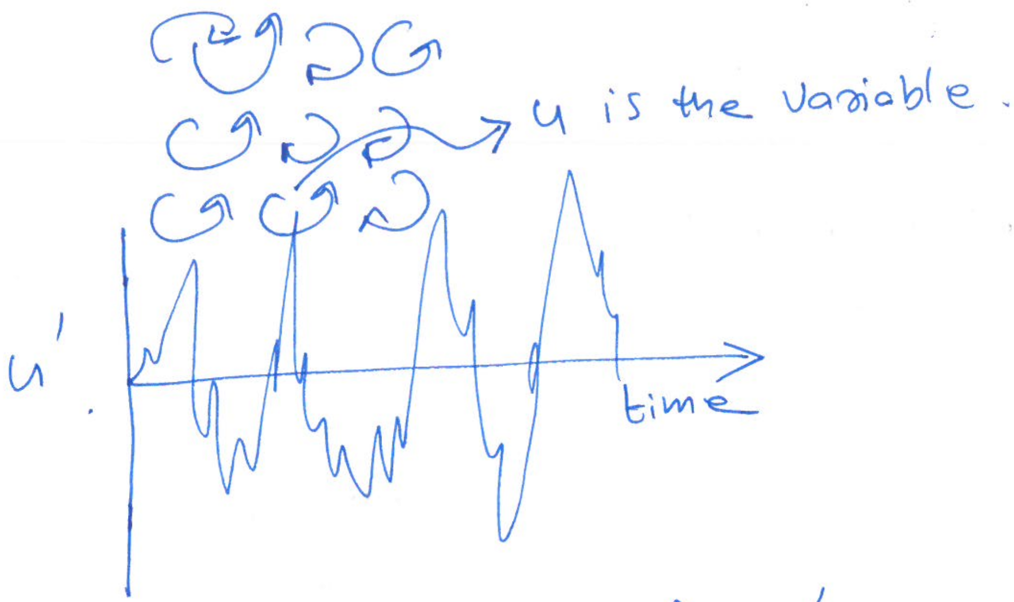
ϵ = dissipation of TKE.

③ Taylor's micro scale "λ"

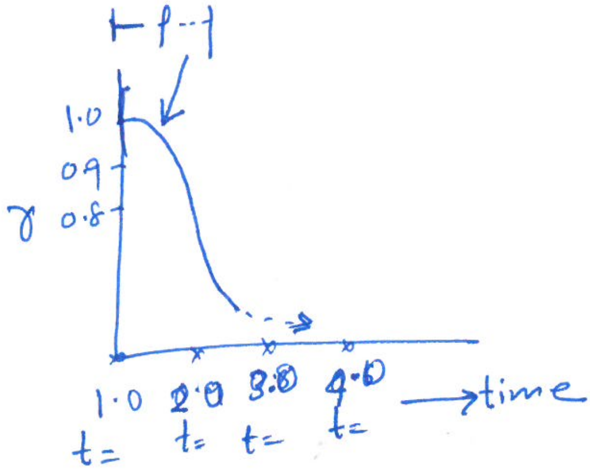
↳ It is the characteristic length scale of energy dissipation.

↳ It is measured as the size of auto-correlation of turbulent variables.

↳ It is related to "curvature" of the auto-correlation function at origine.



Find auto correlation of u'.



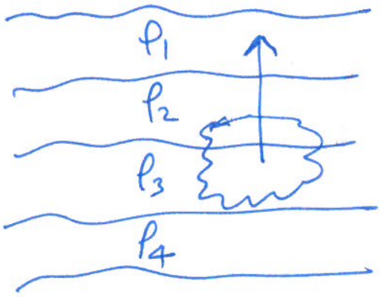
The Taylor's microscale "λ" is related to the auto-correlation ~~[illegible]~~ "r"

$$f_{(r)} = 1 - \left[\frac{r}{\lambda}\right]^2$$

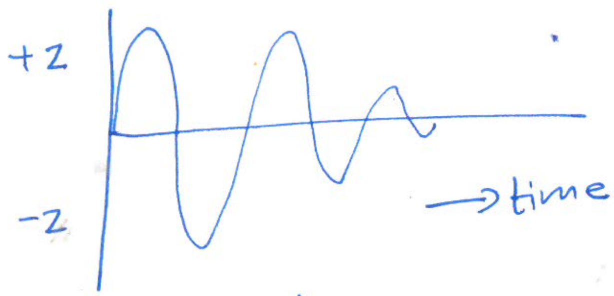
where $f_{(r)}$ is the curvature $\left(\frac{1}{R^2}\right)$ of the auto-correlation at origine.

Integral time-scales

Buoyancy time scale



$$N^2 = -\frac{g}{\rho}\frac{d\rho}{\partial z}$$

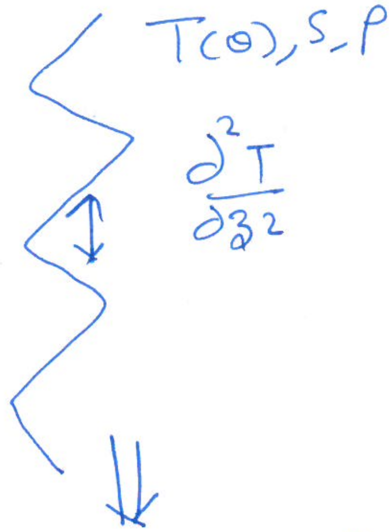


$$N = \frac{1}{S}$$

where S = stability

The timescale of oscillation is called Buoyancy timescale

Dissipation time scale



After time "t_0", the turbulent diffusion "smoothes" the gradient. This time scale is called dissipation timescale.

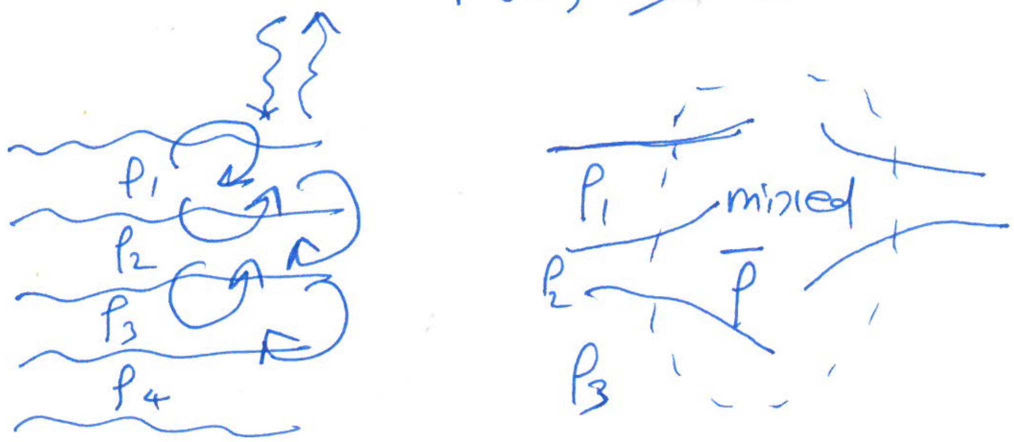
Froude Number

$$F(z) = \frac{\text{Dissipation time scale}}{\text{Buoyancy time scale}}$$

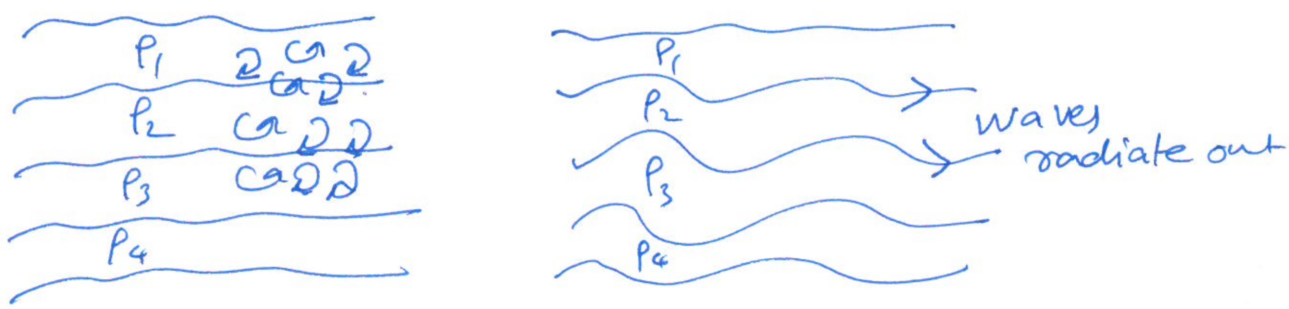
↳ Typically if $F_{(2)} < 0.3$ the turbulence appears to collapse

↳ Also, interestingly, since the stratified fluid permits internal waves at frequencies lower than "N", the eddies with $F_{(2)} < 0.3$ loses its energy on radiating internal waves.

$F_{(2)} > 0.3$

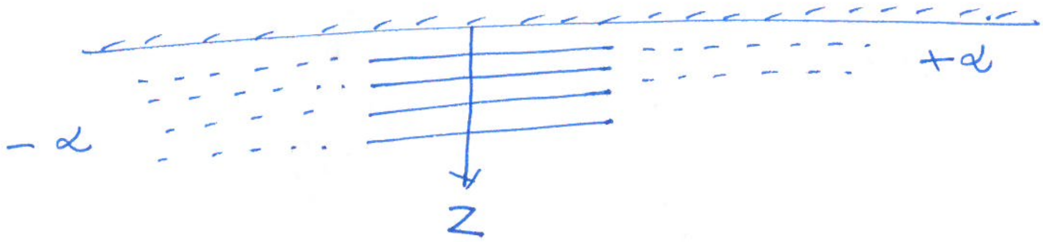


$F_{(2)} < 0.3$



Logarithmic Law of the wall

* This is an universal Law of the wall for nuetrally stratified fluid close to the wall.



* The law applies adjascent to an unbounded wall in horizontal.

* In a region close to the wall various fluxes (momentum, heat etc.) remain approximately constant.

* It is then possible to define a velocity scale $U_* = \left(\frac{\tau}{\rho}\right)^{1/2}$

* More over the mean shear $\frac{dU}{dz}$ is scaled by $\frac{U_*}{Z}$, where "Z" is the distance from the wall (for close to the wall).

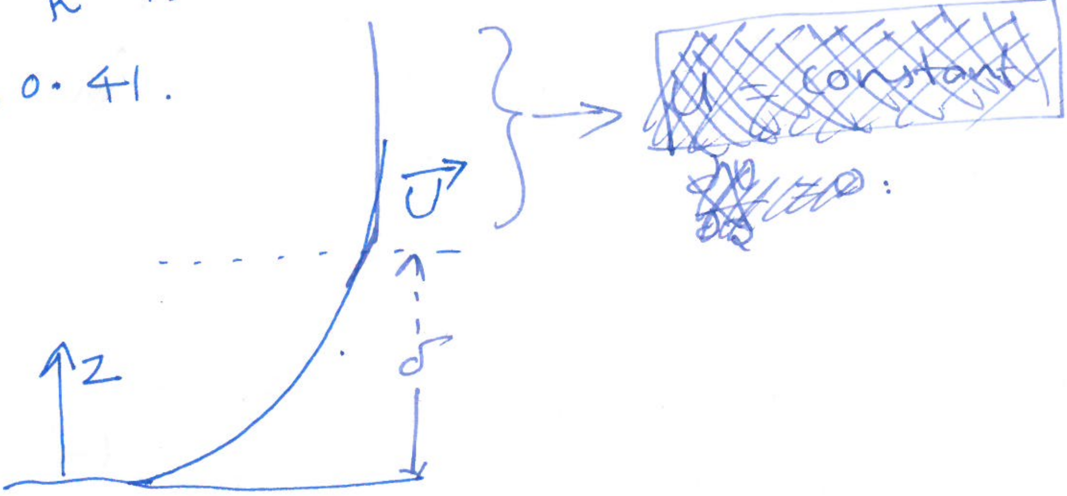
$$\frac{dU}{dz} \propto \frac{u_*}{z}$$

$$\frac{dU}{dz} = \frac{1}{k}\,\frac{u_*}{z}$$

$$\varphi \quad \frac{U}{U_*} = \frac{1}{k}\ln(z) + C.$$

where "k" is the von Karman constant

≈ 0.41.



$$u_* = \left(\frac{\tau}{\rho}\right)^{1/2}.$$

Monin - Obukhov Similarity theory

↳ The universel law of the wall or von Karman law of wall in boundary layer theory is for non-stratified or nuetrally stratified fluids.

↳ However, the atmospheric and ocean boundary layers are strictly under the stratified condition.

↳ The shear induced mixing in such cases is under the expense of the kinetic energy.

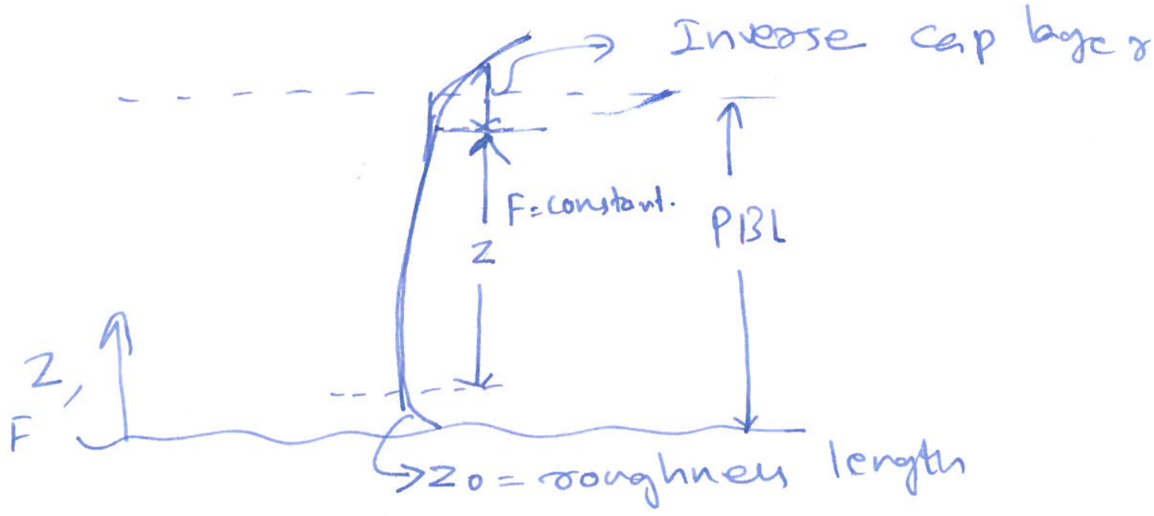
↳ The buoyancy induced mixing in such cases is under the expense of the potential energy.

↳ However the following assumptions can be made use of,

① There exist a region (usually few tens of meters thick) near the ground or ocean surface but sufficiently far from the roughness layer over the ground or skin of the ocean, where the outer & inner scaling laws can be matched to obtain "universal shape profiles" of various quantities.

↳ For the nuetral stratification this yields the von karman logarithmic law of the wall.

↳ For Atmospheric/oceanic stratified case the profiles of various parameter depart from logarithmic behaviour, but still exhibit a similarity. This is called Monin-Obukhoff's similarity law.



↳ Assume the atmospheric condition nearly steady. Let $\frac{\partial F}{\partial z}$ be the gradient of any property (say for example, T, q, U etc...).

↳ Far away from roughness length (z_0) and inversion layer (δ), the gradient of "F" is independent of z_0 and δ, but only a function of z.

↳ What about various fluxes?
Far away from "δ" and "z_0", the fluxes of momentum, heat and density are constant with depth.

Therefore $u_* = \left(\frac{\tau}{\rho}\right)^{1/2}$ or

$$\tau = \rho (u_*)^2$$

$$Q_H = \frac{H}{\rho c_p}$$

$$Q_b = (w_* b).$$

↳ Therefore any property F is a function of, $\frac{dF}{dz} = f(z, u_*, Q_?, Q_b)$.

↳ In such cases a non-diamensional form of relation connecting "F" and the fluxes are,

$$\phi_F = \frac{\kappa z}{F_*}\frac{\partial F}{\partial z} = \phi_F\left(\frac{z}{L}\right) = \phi(\zeta).$$

where $\zeta = \dfrac{z}{L}$.

and F_* frictional velocity scale.

And, $L = \dfrac{u_*^3}{\kappa Q_b}$.

where Q_b = buoyancy forcing.

$u_* = \left(\dfrac{\tau}{\rho}\right)^{1/2}$ = frictional velocity.

Similarity Theory.

Von karman universal law of the wall	Monin-Obukhoff Law of the wall
(1) non-stratified or Nuetrally stratified flow	Stratified flow.

$\boxed{\text{Similar}}$

$$\frac{dU}{dz} = \frac{1}{K}\frac{u_*}{z}$$

Or $\boxed{\dfrac{\kappa z}{U_*}\dfrac{dU}{dz} = 1}$ $\approx$ $\boxed{\dfrac{\kappa z}{F_*}\dfrac{\partial F}{\partial z} = \left[\text{Profile of } \dfrac{z}{L}\right]}$

Therefore the various parameter profiles in the "PBL" is

$$\frac{kz}{U_*}\frac{dU}{dz} = \phi_m(z_l)$$

$$\frac{kz}{\theta_*}\frac{dT}{dz} = \phi_H(z_l)$$

$$\frac{kz}{q_*}\frac{dq}{dz} = \phi_E(z_l)$$

$$\frac{kz}{C_*}\frac{dC}{dz} = \phi_G(z_l).$$

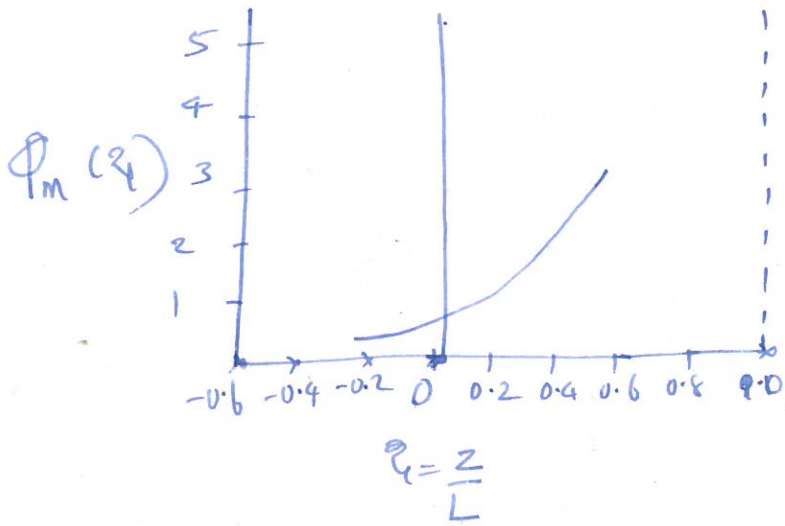
Usual range of "L".

The "L" is called the Monin-Obukhoff length.

$$\boxed{L = \frac{U_*^3}{kQ_b}}$$

This can be few tens of meters.
This can be either (+ve) or (-ve).

$$\boxed{\frac{\kappa z}{u_*}\frac{\partial U}{\partial z} = 1} \approx \boxed{\frac{\kappa z}{u_*}\frac{\partial U}{\partial z} = \phi_m(\zeta)}$$



$\zeta = 1$ the limit of non-stratified fluid.

The Heat Budget

* The net heat flow in the ocean often referred by Q in J/sec/m^2 or W/m^2.

* The net Q-flux can be split into

↳ Q_s = rate of inflow of solar energy through the air sea interface

↳ Q_h = net rate of heat loss by the sea as longwave radiation to the atmosphere and space

↳ Q_h = rate of heat loss/gain through heat conduction (sensible heat flux)

↳ Q_e = rate of heat loss/gain by evaporation/condensation (latent heat flux)

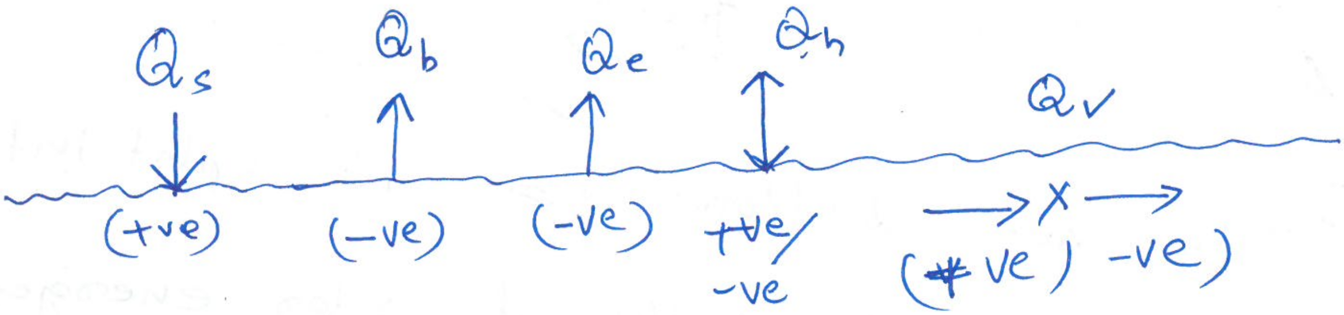
↳ Q_v = rate of heat loss/gain by water body due to currents (advection).

[Figure. 5.5].

$$Q = Q_s + Q_b + Q_h + Q_e + Q_v.$$

$Q > 0$ means water gains heat

$Q < 0$ means water loss heat.



<u>Computation of each heatflux components.</u>

* They are measured by mean of temperature, humidity, wind speed, cloud cover and surface reflectivity..

* The heat fluxes are caclulated from these observations, based on empirical approximations. called "bulk formulae".

↳ Alternatively there are precise observations of the individual heat fluxes.

Such observations are sufficiently complex that they cannot be routinely made.

* Observational Flux estimate errors are as large as 10 W/m^2.

Review of Electromagnetic Radiation Theory.

* Stefan's Law: All bodies radiate energy at a rate proportional to the fourth power of their absolute temperature T.

$$I = \sigma T^4.$$

* Wien's Law: Concentration of radiation energy is not same at all wavelength but has marked peak $\lambda T = 2897\ \mu m K$

Therefor $\lambda \propto \frac{1}{T}$.

↳ body at higher T, emits lower "λ".

↳ body at lower T, emits higher "λ"

* The Sun has surface $T \approx 6000k$.

Therefore maximum energy is concentrated around $0.5\ \mu m$.

* Around 50% the radiation energy from the Sun is in the visible part of the spectrum ($0.35 - 0.7\ \mu m$).

* This energy is referred to as "Short wave radiation" (Q_s).

* The Q_s penetrating atmosphere & cloud reach Ocean Surface where it is absorbed by the water and converted into heat energy.

* The surface ocean heated by the "short wave" emits long wave by

$Q_b = \sigma T^4$.

* Since $\lambda \propto \frac{1}{T}$, the average "lower" temperature of the ocean surface emits long wave radiation around $3 \sim 80\ \mu m$.

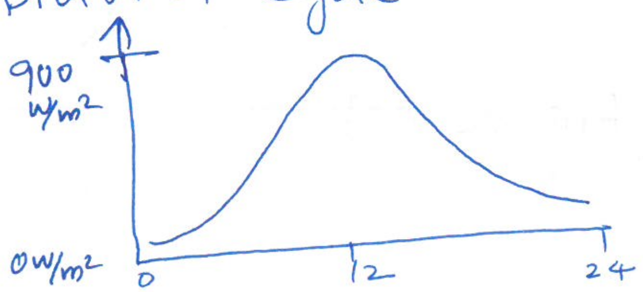
The Bulk (or average) estimation of Q_s is

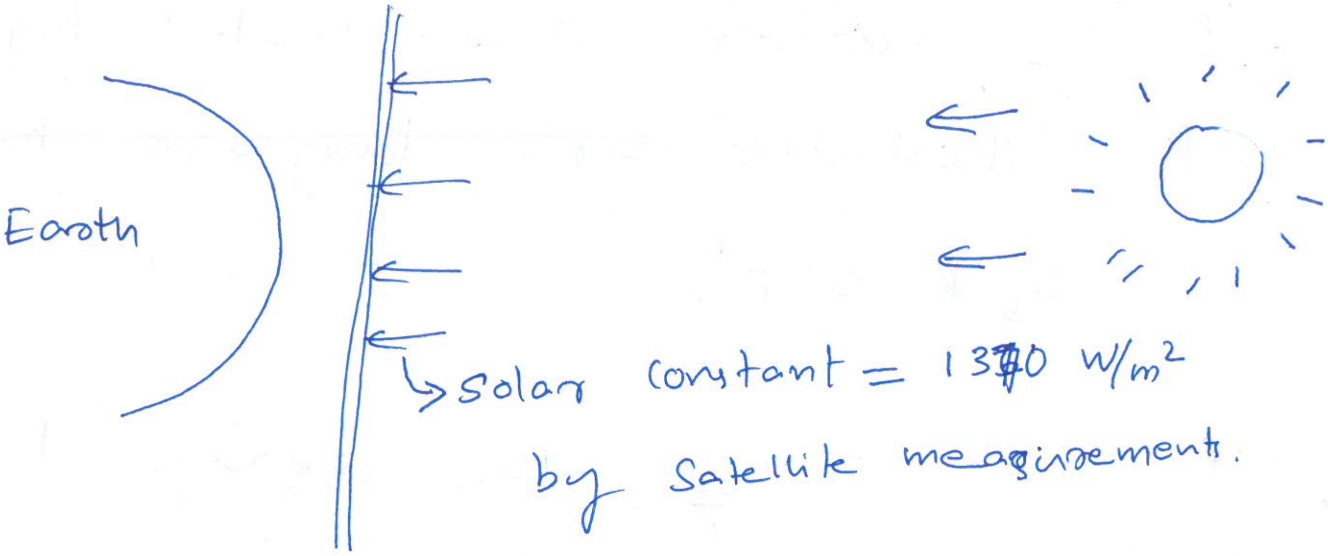
$$Q_s = (1-\alpha)\,Q_c\,(1 - 0.62\,C + 0.0019\,Q_N)$$

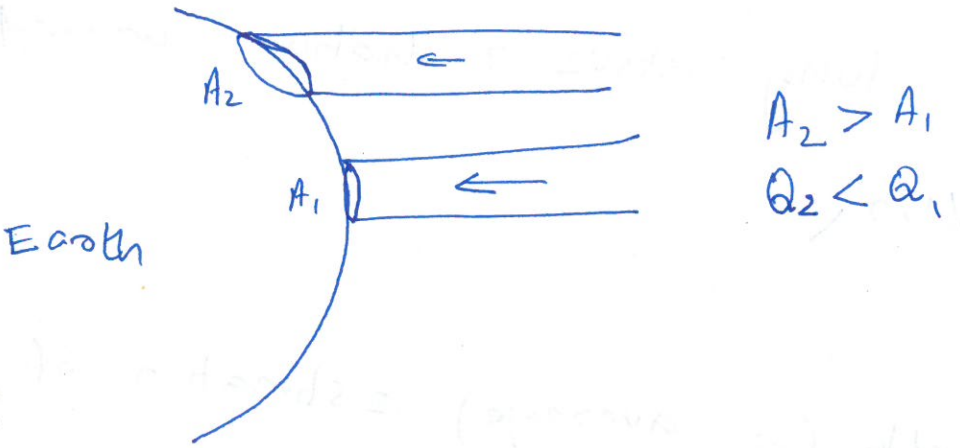
where Q_c = clear sky shortwave solar influx or "Solar constant".

α = albedo, Q_N = noon Solar elevation in degrees.

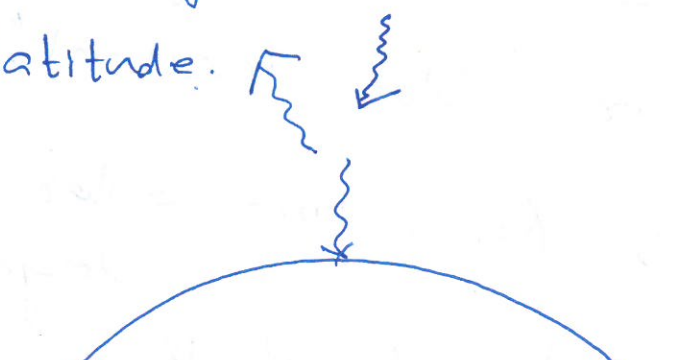
Typical Diurnal Cycle of Q_s in clear sky in Tropic





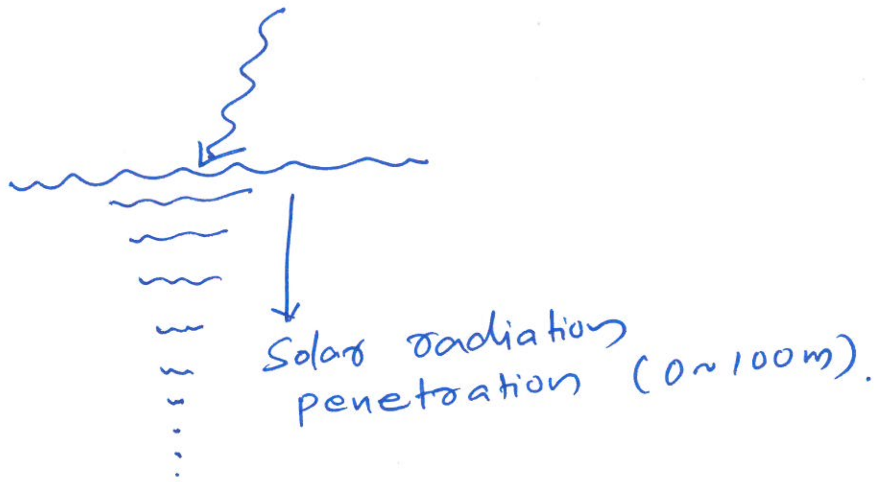


For a given solar energy reaching Tropics & mid-latitude, the $Q = W/m^2$ is larger in tropics than in high latitude.



[Figure 5.5]

Absorption of Q_s in the sea.



* $I_z = I_0 \exp(-k z)$.

where "k" is the absorption coefficint.

↳ $k = f(\lambda)$.

$k = f($ contaminations in the water such as biological growth$)$.

* Light photons are used for photosynthesis in the ocean.

[Read chapter-3 of Talley et al; book]

Long Wave radiation (Q_b)

$$Q_b = \epsilon \sigma_{SB} T_w^4 (0.39 - 0.05 e^{1/2})(1 - k c^2)$$
$$+ 4 \epsilon \sigma_{SB} T_w^3 (T_w - T_A).$$

ϵ is the emittance of the sea surface (0.98).

σ_{SB} = Stefan-Boltzmann constant ($5.67 \times 10^{-8}\ W/m^2/K^4$).

T_W = is water temperature.

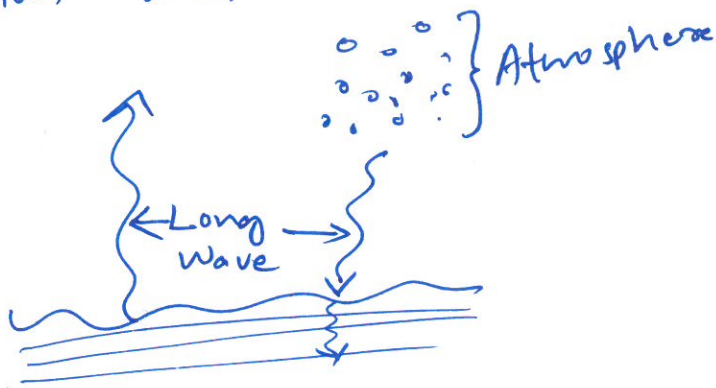
T_A = Air temperature.

k = cloud cover coefficient

C = cloud fraction.

[Figure. 5.11].

SST and Penetration depth of Long wave.



* Long wave emission is calculated from the Bulk surface ocean temperature (0.5~1m beneath ~~the~~).

* The Skin SST can be larger/smaller than bulk SST.

(See the Talley et al., book for further reading).

Evaporative or Latent heatflux.

* Evaporation requires supply of heat from an outside source or from the remaining liquid

* Besides the loss of water volume, it also implies loss of heat.

The rate of heat loss

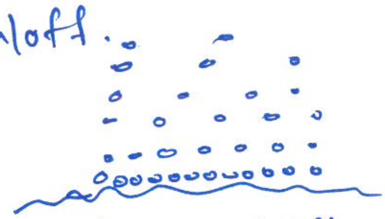
$$Q_e = F_e \cdot L$$

F_e = ratio of evaporation of water in $\frac{kg}{sec\, m^2}$ and L is the latent heat of evaporation. (KJ).

* The average amount of "Fe" from the sea-surface is ≈ 120 cm/yr.

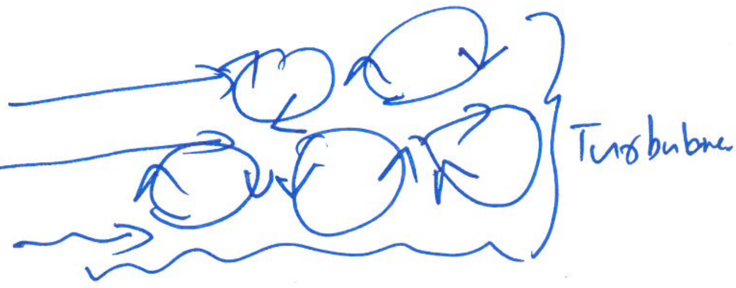
≈ 1 m/yr

* Evaporation is basically a diffusive process that depends on water vapor concentration gradient from sea surface to aloft.



* This is analogous to molecular diffussion but occurring because of "eddy diffusion" in the atmosphere

* "Eddy" is a large chunk of mass which is analogous to "a molecule".



Bulk formula for F_e

$$F_e = \rho C_D |U| (q_s - q_a)$$

$$Q_e = F_e \cdot L$$
$$= \rho C_D L |U| (q_s - q_a)$$

C_D = Transfer coefficient

L = Latent heat of vaporization.

q_s = the saturated specific humidity.

* q_s is the maximum amount of water vapor in "g", per unit mass of air, in kg.

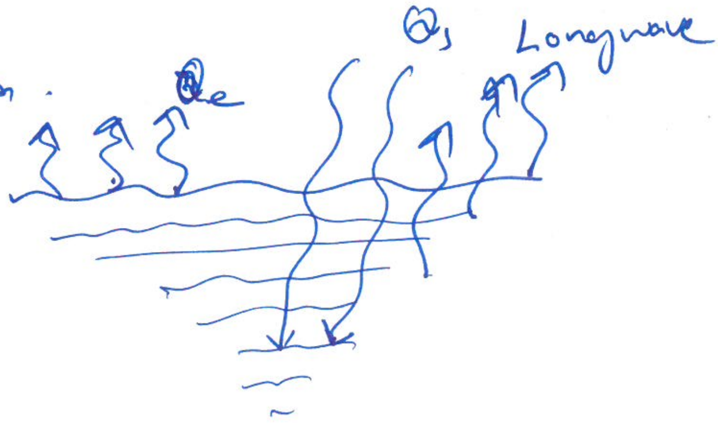
* q_a is the measured specific humidity.

* $q_s > q_a$.

* Therefore in most part of the ocean evaporative heat loss occurs.

* If the ocean temperature is $> 0.3°C$ than air temperature, $q_s > q_a$; there will be loss of heat from the sea due to evaporation,

* Qe occurs from the topmost layer of the sea.



Sensible Heat flux.

(Heat conduction).

$$Q_h = -A_n c_p \frac{\partial T}{\partial z}.$$

* This is the diffusive heat loss by difference in air temperature between sea surface an aloft.

* The bulk formula for ~~Q_h~~

$$Q_h = \rho c_p C_h |U| (T_s - T_a).$$

or $$= \rho c_p C_h |U| \left[T_s - T_a(1+\gamma Z)\right]$$

where T_s = air temp ~~above~~ close to the sea

T_a = air temp above the sea

γ = lapse rate of atmosphere.

C_h = transfer coefficient.

Geographical distribution and temporal variation of the heat budget term.

Figure 5:11

Figure 5:12

Figure 5:13

Oceanic Meridional Heat transport

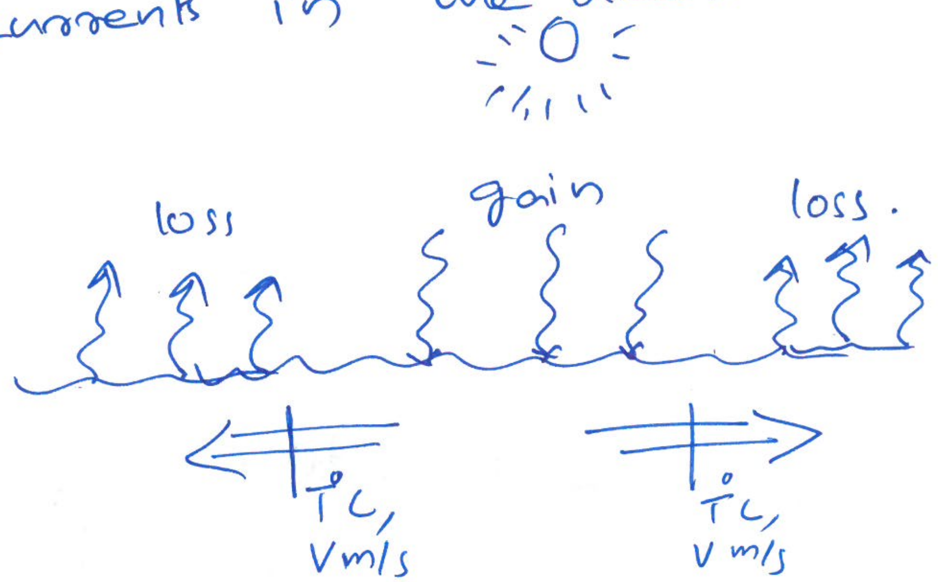
* The ocean gains heat in the tropics between 30°N ~ 30°S, and loses heat at highter latitudes.

* Therefore the net heat gained in the tropics is transported to the high-latitude in the ocean.

*

* The meridional Heat transport within the ocean is calculated in two ways.

(1) In-direct method which uses the information of net heat flux in and out of the ocean and calculate the heat transport without looking at the currents in the ocean.



(2) Direct-method: The heat transport calculated by ocean current and temperature.

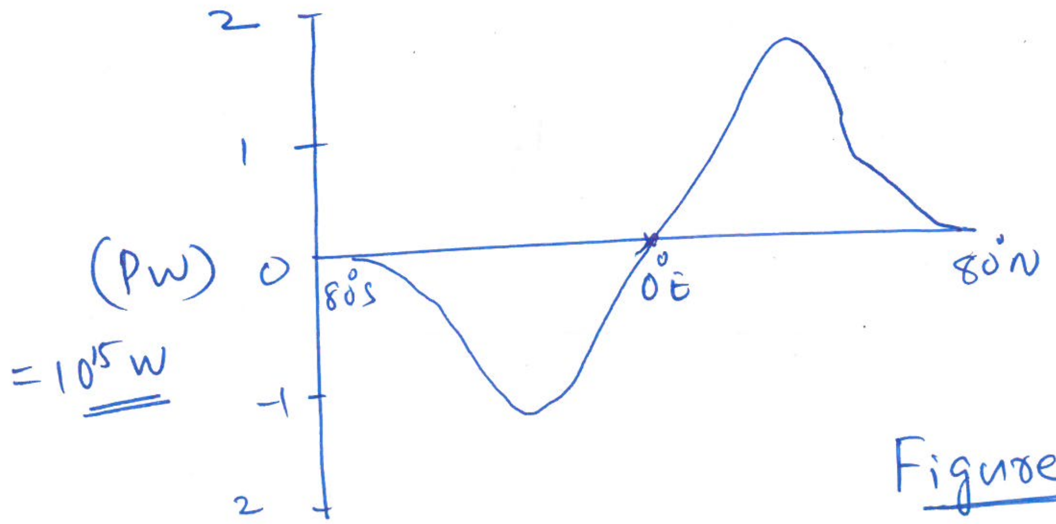


Figure 5:14

Buoyancy Fluxes.

* Buoyancy forcing changes the density of seawater.
* This includes heat fluxes and evaporation – precipitation and river-run-off.

Figure 5:15.

* Difference between buoyancy fluxes and net heat fluxes.
  - ↳ Heat fluxes change the temperature of the ocean.
  - ↳ Buoyancy fluxes change the density of the ocean.
  - ↳ Heat fluxes do not have effect of precipitation, river-run off, ice formation, brine rejection etc.

Wind forcing

Figure 5:16

Wind stress (Bulk formula).

$$\tau_x = \rho C_D |U| U.$$

Wind stress curl.

Figure 5:16.

Wind stress curl and Sverdrup transport

$$-fv = \frac{\tau_x}{\rho} \quad \text{—— (1)}$$

$$fu = \frac{\tau_y}{\rho} \quad \text{—— (2)}$$

$\frac{d}{dy}$ (1) and $\frac{d}{dx}$ (2)

$$-f\frac{dv}{dy} - v\frac{df}{dy} = \frac{1}{\rho}\frac{d\tau_x}{dy}$$

$$f\frac{du}{dx} = \frac{1}{\rho}\frac{d\tau_y}{dx}$$

(−)