mast602

Lecture 1

Properties of the ocean and seawater

The scale of the ocean
Temperature, salinity, density, stability
Sound in the ocean
Light in the ocean

MAST 602

Lecture 1
Properties of the ocean and seawater

Overview of physical oceanography

Reference: Chapter 1 of (Knauss 1997)

Physical oceanography concerns the distribution of temperature, salinity, density, and velocity in the ocean.

We begin by looking at the scale of the oceans then will in turn examine the characteristics of temperature, salinity, density, & stability and other characteristics of seawater

We leave the question of velocity until we have looked at the equations of continuity and motion

The scale of the oceans

71% of earth's surface is water
The ratio of water to land is 4:1 in the Southern Hemisphere

The size of the Pacific approximates the Atlantic & Indian together
These three oceans together make up about 90% of the ocean,

The average depth of the ocean is about 3,800m.
(Average land elevation is about 840m.)

Fig. 1- 1 The ocean is deeper than the land is high! **Fig 1.7, Knauss**

[The hand-out notes will generally include copies of those figures that are not in Knauss]

Maximum ocean depth = 11,524m. (Mindanao Deep)
Maximum land height = 8,840m. (Mt. Everest)

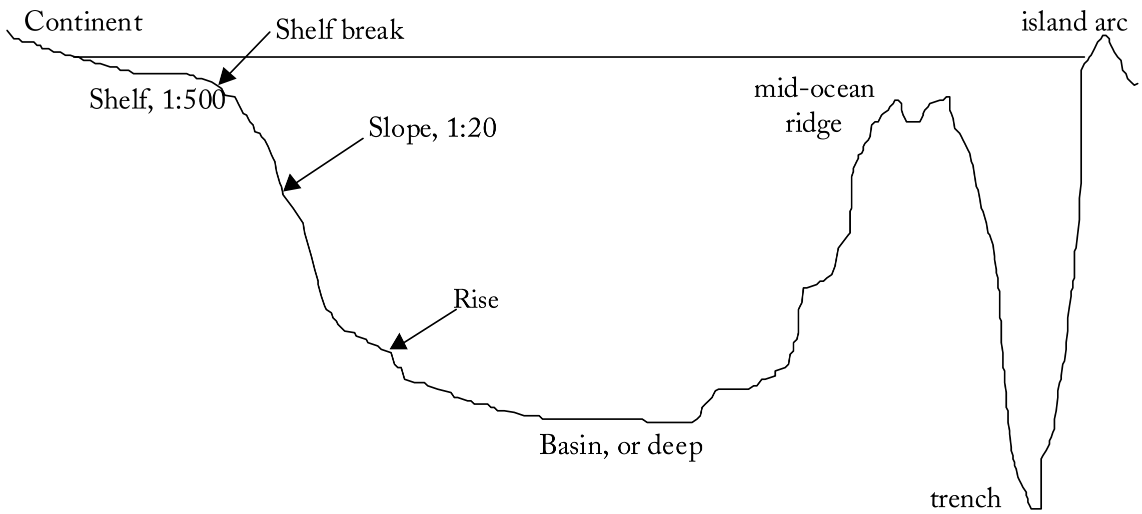

Fig. 1- 2 Typical ocean bottom profile

Beware of vertical exaggeration!
The ocean is really proportionately as thin as a sheet of paper.

Figures with a vertical exaggeration of 1000× can sometimes lead to an erroneous intuition about oceanic processes.

If the above figure were drawn to scale, how high would it be?

Location on planet Earth is specified using latitude and longitude:

1 nautical mile = 1 min. of latitude
60 min. of latitude = 1 deg. of latitude

How many nautical miles from equator to pole? How many kilometers from equator to pole? So what ought to be the conversion factor from nautical miles to kilometers?

e.g., Cannon Building in Lewes is:
39^{o} 47'.3 N
79^{o} 9'.7 W

Temperature, *T*, θ

Temperature is easy to measure
It generally decreases with depth
It generally decreases poleward
The decrease with depth is more rapid
at the surface than at depth

Warm water is found at shallow depths
because the ocean is an effective absorber
of the sun's radiation (more on this in
the next two lectures.)

The surface layers are heated in summer
and cool in winter
Since the surface water is warmer
than the water below, it is less dense
and does not mix downward easily.

This is unlike the atmosphere, which,
since it is heated by the land below,
is subject to convection and active mixing.

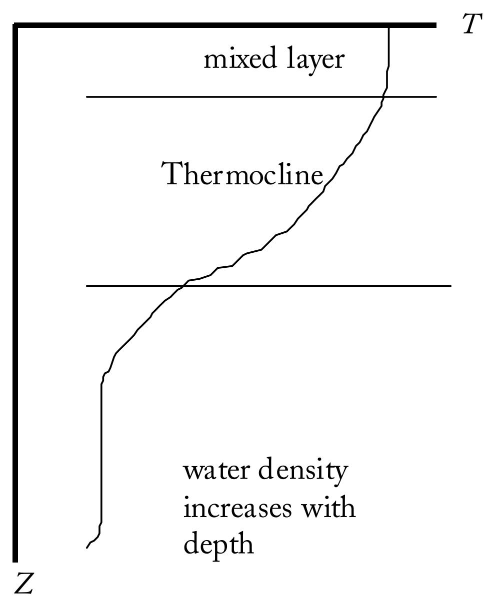

Fig. 1- 3 **Vertical temperature profile**

50% of water in the ocean is colder than 2.3°C [36 °F]

Temperature units are degrees Celsius [°C]
or degrees Kelvin [K]
We describe temperature two ways:

In-situ temperature, T, is the temperature
as measured in place

Potential temperature, θ, is the temperature
a parcel of water would have if it were
raised *adiabatically* to the surface

Potential temperature takes account
of *adiabatic* effects:
Consider an insulated water parcel:
heat added + work done
= change in internal energy

Since the parcel is insulated,
we set heat added to zero.

Thus, as the parcel rises, pressure decreases,
work is done *by* the parcel in expanding,
so the temperature decreases.

As a parcel descends, pressure increases,
work is done *on* the parcel in compressing,
and the temperature increases.

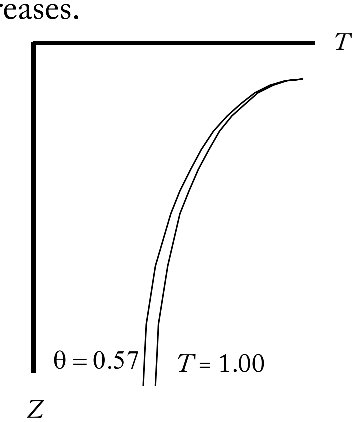

Fig. 1- 4 In-situ and potential temperature profiles

examples:
water at 4000m at 5 °C
raised adiabatically to the surface
decreases to 4.583 °C. Table XVIII

water at the surface at 5 °C
lowered adiabatically to 4000m
increases to 5.429 °C. Table XXI

Salinity, *S*

The fraction, by mass, of dissolved solids
multiplied by 1000

The concentration of salt per unit mass is ρS
where ρ = density, in Kg m^{-3}

The old salinity definition
(‰ = parts per thousand)
was based on titration:
salinity = 1.80655 × chlorinity
= *absolute salinity* $\boxed{35‰}$

The current salinity definition,
practical salinity, is based on the
electrical conductivity of seawater.

The unit is often called psu, the $\boxed{\text{35 psu}}$
practical salinity unit. However,
practical salinity as defined is a
unitless quantity $\boxed{35}$

To measure salinity,
temperature must be known well:
±0.01°C —> ±0.01 in salinity

Using conductivity, salinity can be measured
to about ±0.003
or even ± 0.001
(That's about one part per million.)
Titration can give only about ±0.02.

Salinity is often used as a tracer,
where small differences can
distinguish the movement of
water masses in the deep ocean,
and hence it must be well measured.

75% of salinity values lie in the range
34.50 → 35.00

Half the water in the Pacific
is between 34.6 and 34.7.

Density, ρ

Measured in kilograms/cubic meter [Kg m^{-3}] Table I

Range:

1020 —> 1070	Kg m^{-3}
1020 —> 1030	(ignoring the effects of pressure)
1027.7 —> 1027.9	for 50% of the ocean (ignoring pressure)

Density is a function of
pressure, temperature, salinity;
pressure has the greatest effect

Because of the small range in density,
(the first two digits never change)
some kind of *density anomaly* is used.

That is, σ = density -1000 Kg m^{-3}

Recently, because of an improved equation of state for seawater, the symbol γ is used to denote the *density excess*

> Though σ and γ are often used interchangeably, the equation for density excess, γ, gives different results than that for the density anomaly, σ.
> For example, on page 29 of Knauss, the σ-values are from the old tables.
> The new tables (part of MAST602 handouts) give better values and should be used.

Let $\sigma_{s,t,p} = \gamma_{s,t,p}$ = density − 1000 Kg m^{-3}
$= \rho - \rho_{00}$
where ρ_{00} = 1000 Kg m^{-3} → fresh water @ 4°C

$\gamma_{s,t,p} = \sigma_{s,t,p} = \rho_{s,t,p} - 1000$ Kg m^{-3} *in-situ density anomaly*

$\gamma_t = \sigma_t = \rho_{s,t,0} - 1000$ Kg m^{-3} *atmospheric-pressure density anomaly*

$\gamma_\theta = \sigma_\theta = \rho_{s,\theta,0} - 1000$ Kg m^{-3} "*potential density anomaly*"

Fig. 1- 5 S, θ, T, σ_t, σ_θ, $\sigma_{s,t,p}$ for a North Pacific station **Fig 2.4, Knauss**

Note: Though σ_t is used extensively by physical oceanographers, there is considerable confusion in the oceanographic literature concerning its definition.

[Strictly speaking, this term should really be called γ instead of σ and should be referred to as the *density excess*. In fact, nearly all oceanographers persist in simply calling this term "sigma" and refer to it, when they do use any other term, as the *density anomaly*.]

Thus γ_t is defined as:

$\gamma_t = \rho(s,t,0)$ - 1000.00 kg m^{-3}.

The older definition of σ_t was essentially:

$$\sigma_t = \left[\frac{\rho}{\rho_{max}} - 1\right] \times 10^3$$

$$= \frac{\rho}{\text{kg m}^{-3}} - 999.972$$

where ρ_{max} = 999.972 kg m^{-3} , an early measure of the maximum density of distilled water.

The two definitions are not the same; the differences between them are significant.

Fig. 1- 6 Variation of *T*, *S*, and σ_t with latitude Fig 4.3, (Pickard and Emery 1990)

Why are there two maxima in the salinity vs. latitude curve?
Why is σ_t a minimum near the equator?
Why does salinity decrease at high latitudes?

Fig. 1- 7 *T*-*S* curve, with isopycnals in the North Atlantic Fig 2.8, Knauss

Why are the isopycnals useful in this diagram?

Specific volume, α

The *specific volume*, α is the reciprocal of density:

$\alpha = 1/\rho$ [m^3 kg^{-1}]

Specific volume anomaly, δ

$$\delta = \alpha_{s,t,p} - \alpha_{35,0,p}$$

Stability (Static stability), *E*

Static stability might be considered as the "unwillingness of water to be moved vertically"

Consider the work involved in moving a particle of volume V vertically across an interface between two density layers:

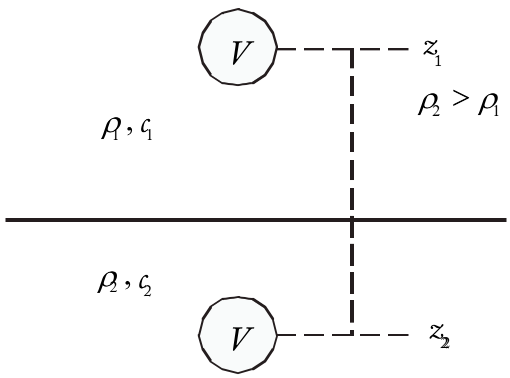

Fig. 1- 8 **Moving a particle between two layers**

Work required to move the particle up is

$(\rho_2 - \rho_1)\, Vg\, Z_1$ g = 9.81 ms^{-1}

(work is only required to move it through the layer of density ρ_1)

or, ΔPE = work/V = $(\rho_2 - \rho_1)\, g\, Z_1$

as $\rho_2 - \rho_1$ —> 0,

ΔPE —> 0

Thus, if $\rho_2 = \rho_1$,no work is required to move a particle of water vertically

And there is no change in potential energy of the water column

In reality, the density of the ocean changes continuously with depth

Stability is defined in terms of the rate of change of density with depth and is a measure of the amount of work required to move a particle vertically in the water column:

$$E = -\frac{1}{\rho}\frac{\partial \rho}{\partial z}$$

… where the minus sign is necessary since z increases upward

or, more simply:

$$E = -\frac{1}{\rho}\frac{\partial \sigma_\theta}{\partial z}$$

For depths less than about 1000 m, σ_t may be used instead of σ_θ

Here's another example of why σ_θ (or σ_t) is a useful quantity

Static stability:
density increases with depth
(E is positive)

Neutral stability:
no change of density with depth
(E is zero)

Static instability:
lighter water below
(E is negative; i.e., convection occurs)

Stability is generally estimated using σ_t
(at depths less than ~ 1,000 m)

or by using σ_θ
(at depths greater than ~ 1000 m or at any depth)

A more exact equation for stability, to account for compressibility, is

$$E = -\frac{1}{\rho}\frac{\partial \rho}{\partial z} - \frac{g}{c^2}$$

where c is the velocity of sound.

Another way of expressing stability is with the *Brunt-Väisälä oscillation* (sometimes called *buoyancy oscillation*)

This is the oscillation of a water particle about its equilibrium depth.

A particle of water forced upward will be heavier than the surrounding water and buoyant forces will push it downward

The Brunt-Väisälä oscillation has:

frequency (sometimes called the *buoyancy frequency*):

$$N = \sqrt{gE} = \left(-\frac{g}{\rho}\frac{\partial \rho}{\partial z} \right)^{\frac{1}{2}} \approx \left[-\frac{g}{\rho}\frac{\partial \sigma_T}{\partial z} \right]^{\frac{1}{2}} \qquad [\mathrm{s}^{-1}]$$

period:

$$T = \frac{2\pi}{N} \qquad [\mathrm{s}]$$

Typical values in the ocean: 1 min. < T < a few hours.

Other Properties of seawater

Compressibility

Compressibility decreases with increasing salinity, temperature, and pressure

Electrical Conductivity

Electrical conductivity increases with temperature and salinity. It is used to estimate seawater salinity.

Coefficient of thermal expansion

The coefficient of thermal expansion increases with temperature and pressure

Fig. 1- 9 Density maximum and freezing point **Fig 2.2, Knauss**

Increasing salinity lowers both temperature of the density maximum and the temperature of the freezing point. The temperature of the density maximum drops off more rapidly; they cross at $S = 24.695$.
For freshwater, what is the temperature of maximum density?
What will happen as freshwater cools below the temperature of maximum density?
The same effect occurs in waters of low salinity (like parts of Delaware Bay).

Sound and Light

The ocean is opaque
to electromagnetic (EM) radiation
e.g., long radio waves, short ultraviolet light

There's a small window
in the visible spectrum,
but in general only about 1% of light
penetrates to 100m depth.

EM waves penetrate a fraction of a wavelength
e.g., submarine communications
at very long wavelengths

On the other hand,
the ocean is transparent
to acoustic transmission
e.g., an acoustic source in Australia
can be heard by a receiver in Bermuda.

(Interestingly, the atmosphere is more
transparent to light, and less
transparent to sound)

Because sound propagates so well in the
ocean, it has been extensively studied
and used for a variety of marine purposes

Underwater sound

Ref: Chapter 12 of Knauss

Ocean acoustics were heavily studied in the past for tracking enemy submarines.

Sound velocity is about
1500 ms^{-1}
or 5000 ft/sec,
or about 1 mile/sec

Sound speed is a non-linear function of temperature and salinity and a linear function of pressure

The pressure (i.e., depth) effect is:

$$\frac{\Delta c}{\Delta z} = \frac{1.7\mathrm{ms}^{-1}}{100\mathrm{m}} \approx 0.017\mathrm{s}^{-1}$$

To look at the temperature and salinity effects, consider Wilson's formula for sound velocity (simplified):

$$c = 1449 + 4.6T - 0.055T^2 + 0.0003T^3 + (1.39 - 0.012T)(S - 35) + 0.017Z$$

where c = sound velocity (ms^{-1})
T = Temperature (°C)
S = Salinity
Z = Depth (dbars) [*Not* measured up in this case!]

Note that salinity has little effect on sound velocity except in regions where T is small.

The *frequency* of sound, assuming a sound velocity of 1500 m/s is:
(1 hertz [Hz] = 1 cycle/sec.)

Frequency (Hz)	10^2	10^3	10^4	10^5
Wavelength	15 m	1.5 m	15 cm	1.5 cm

Sound level

The units of *sound level* are
decibels, dB.

Don't confuse units of sound level, decibels, dB, with units of pressure, decibars, dbars!

Since sound level is non-dimensional, a reference level must be defined:

$$dB = 10 \ \log_{10}\left(\frac{I}{I_{\mathrm{ref}}}\right)$$

where

I = *sound intensity*, is a measure of the flux of sound energy

I is proportional to the square of the pressure fluctuation, Δp, which varies about normal hydrostatic pressure:

$$I = \frac{(\Delta p)^2}{\rho c}$$

c = sound velocity
p = pressure
ρ = density

For ocean acoustics, I_{ref} corresponds to

$\Delta p = 1\times 10^{-6}\ \mathrm{Nm}^{-2}$, or 1 micro Pascal

Absorption and Scattering

Sound energy is *absorbed* and *scattered* in the ocean, together called *attenuation*

Absorption of sound is due to viscosity and increases with the square of frequency

Scattering is due to inhomogeneities in the ocean: temperature microstructure, bubbles, plankton, particulate matter

Sound attenuation follows *Beer's law*, where the loss of energy is proportional to the energy present:

If I is the sound intensity,

$$\frac{dI}{dR} = -jI$$

where j is the attenuation coefficient.

If R is the distance from the source, Beer's law gives

$$I = I_0 e^{-jR}$$

The sound intensity, I, is a function of frequency

For example, over the frequency range 5 →10 kHz,
the attenuation coefficient is about

For seawater:
$j \sim 1.5 \times 10^{-8} f^2$ dB/km

For freshwater:
$j \sim 2.08 \times 10^{-10} f^2$ dB/km

So, for sound with a frequency of
10 KHz in the ocean,
attenuation ~ 1.5 dB/km.

In contrast, for radio signals, in seawater
where $f \sim 30$ MHz,
absorption is about 1 dB/30cm.

As an example of sound attenuation,
consider the extinction distance, X_e,
over which the sound intensity
falls to $1/e$ of its value.:

frequency f s^{-1}	wavelength Λ m	extinction distance X_e Km
10^2	15	4.37×10^6
10^3	1.5	4.37×10^4
10^4	.15	437
10^5	.015	4.37

Refraction of sound

Sound does not travel in straight lines.
It can be *refracted* (bent) or *reflected* by density and sound velocity changes in the interior of the ocean.

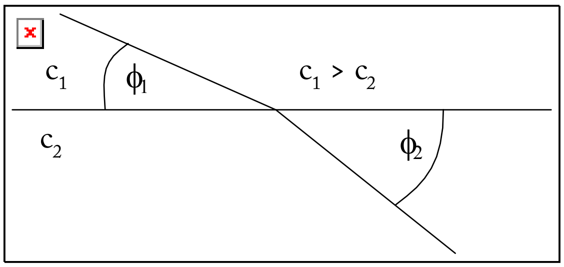

Fig. 1- 10 Sound ray refraction

Snell's law:

$$\frac{c_1}{c_2} = \frac{\cos\phi_1}{\cos\phi_2}$$

for multiple layers,

$$\frac{c_1}{\cos\phi_1} = \frac{c_2}{\cos\phi_2} = \cdots = \frac{c_n}{\cos\phi_n}$$

The critical angle, ϕ_c, is defined as

$$\cos\phi_c = \frac{c_1}{c_2}$$

if $\phi < \phi_c$, then no refraction can occur.

Sound refraction can be traced with rays

Fig. 1- 11 Sound ray paths **Fig 12.7, Knauss**

Sound rays bend to lower velocity

A source of sound (like a submarine) can be in a shadow zone.

Normally, the ocean has a mid-depth minimum in sound velocity.

Fig. 1- 12 Sound speed profile in the Pacific Fig 12.1, KNAUSS

The mid-depth velocity minimum is called the *sound channel.* Why?

Since sound tends to bend into it, there is excellent sound propagation in the sound channel.

Fig. 1- 13 Typical ray diagram Fig 12.2, KNAUSS

How does this mixture of rays correspond to a sound channel?

The sound channel can be used
for tracking acoustic floats
(and hence for tracking ocean currents).

Reflection of sound

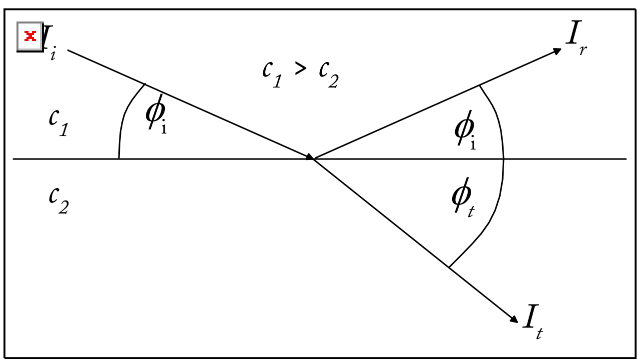

Fig. 1- 14 Sound ray reflection

If sound velocity changes between two layers there can be good reflection:

$$\frac{I_r}{I_i} = \left(\frac{\rho_2 c_2 \sin\phi_i - \rho_1 c_1 \sin\phi_t}{\rho_2 c_2 \sin\phi_i + \rho_1 c_1 \sin\phi_t} \right)^2$$

and $I_i = I_r + I_t$

where

ϕ_i = incident angle	I_i = incident intensity
ϕ_r = reflected angle	I_r = reflected intensity
ϕ_t = transmitted angle	I_t = transmitted intensity

Underwater sound applications

- Communications
- Finding submarines
- Finding fish
- Measuring ocean depth
- Probing ocean sediments

Fig. 1- 15 Sonoprobe section **Fig 12.9, KNAUSS**

- Tracking oceanographic instruments
  SOFAR = **SO**und **F**ixing **A**nd **R**anging
- Acoustic tomography

We can choose an appropriate frequency in order to find the best tradeoff between signal attenuation and target definition:

Frequency	Attenuation	Definition
high	high	better
low	low	low

Underwater optics

The ocean is nearly opaque to light.

Refraction

Snell's law at the surface is

$$\frac{\sin i}{\sin j} = 1.33$$

assuming no waves!

Where i is the incident angle, and j is the transmitted angle *measured from the vertical*

(You'd think this would be the same equation as the refraction of sound as given above, but optics people use a different convention —watch out for it!)

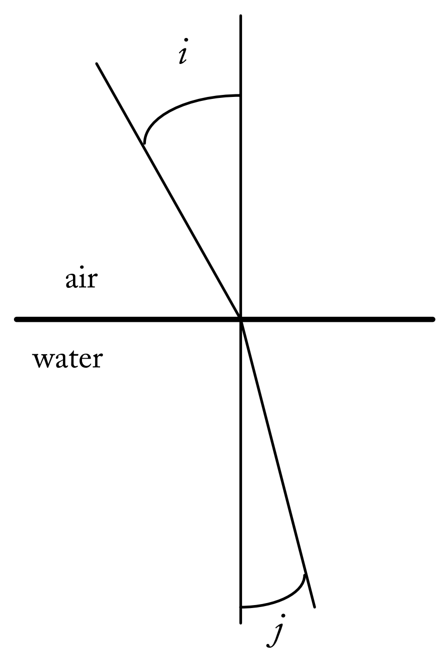

Fig. 1- 16 Light ray refraction

Scattering (ν) and *absorption* (ξ) are
wavelength (Λ) dependent,
$\sim \Lambda^{-4}$

About 6% of the energy is back-scattered.

Beer's law (the same law as above, but again, with a slightly different formulation) is:

$$\Gamma_2 = \Gamma_1 e^{-\varepsilon(z_2 - z_1)}$$

or

$$\ln \Gamma_2 = \ln \Gamma_1 - \varepsilon(z_2 - z_1)$$

where Γ_1 and Γ_2 are the *irradiances*
(the flux of radiant energy)
at depths z_1 and z_2.

Warning: In this case, Knauss's text (and these notes) have z measured *downwards*

The *attenuation* coefficient,

$$\varepsilon = \xi + \nu$$

(ξ is generally the largest of the two terms)

Fig. 1- 17 Absorption coefficient for phytoplankton **Fig 12.15, Knauss**

What conclusions can one draw from the low absorption in the "visible" part of the spectrum?

Absorption dominates attenuation
over most of the spectrum
(except in the visible).

Transmittance, Π, is the
percentage of irradiance
that passes a distance.

If z_2 - z_1 is one meter,

$\Pi = \Gamma_2 / \Gamma_1$
= transmittance per meter

Fig. 1- 18 Downwelling solar spectral irradiance **Fig 12.13, Knauss**

> Note the peak frequency shifts to shorter wavelengths with depth. How might this be manifested?

Note that 50% of radiant energy
is absorbed in the first meter of the ocean.

Particulate matter absorbs and scatters
shorter wavelengths the most

When *turbidity* increases,
longer wavelengths
dominate the transmittance;
shorter wavelengths
are absorbed by turbidity.

Ocean color:

Determined mostly by biological activity

Remote sensing is used to observe
and measure ocean color

Fig. 1- 19 Upwelling light leaving the surface **Fig 12.16, Knauss**

> What do the various *c*-values correspond to in the ocean?
> What does the shift to longer wavelengths do to the observed ocean color?

Fig. 1- 20 Illumination and depth **Fig 5.1, (Brown, Colling et al. 1995)**

> Note the potential effect on many biological processes.

e.g., phytoplankton can only grow
during daylight in a narrow depth range
near the surface (say the upper 200 m)

Lecture 1 Figures

References

Brown, Evelyn, Angela Colling, et al. (1995). *Seawater: its composition, properties and behaviour*. 2nd, Pergamon, Milton Keynes, 168 pp.

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