Chapter 4 - Measuring T, C, p, and derived quantities

Contents

1. Measuring temperature

There are several means available to measure temperature:

1. Expansion of a liquid or metal

From 1874 (Negretti and Zambra) and again in the 1990s (at that time mainly with the aim of calibrating the measurements of the new instruments that were beginning to spread in the community), oceanographers measured *in situ* temperature at depth using a reversing mercury thermometer.

2. Differential expansion of two metals (bimetal)
3. Vapour pressure of a liquid
4. Thermocouple

A thermocouple is an electrical device consisting of two dissimilar electrical conductors forming an electrical junction. A thermocouple produces a temperature-dependent voltage as a result of the Seebeck effect (1821), and this voltage can be interpreted to measure temperature. Commercial thermocouples are inexpensive, interchangeable, are supplied with standard connectors, and can measure a wide range of temperatures. In contrast to most other methods of temperature measurement, thermocouples are self powered and require no external form of excitation. The main limitation with thermocouples is accuracy; system errors of less than one degree Celsius (°C) can be difficult to achieve.

5. Modification of the resistance of a conductor (the most commonly used today, see below)
6. Infrared radiation emanating from the sea surface

All these quantities are linked to T by approximate/empirical laws: a power expansion is carried out and only lower order terms are retained.

Note: if you would like to know more about temperature measurements, thermometers, and standards please see these sites:
https://en.wikipedia.org/wiki/Thermometer
https://en.wikipedia.org/wiki/Temperature_measurement

Cases 1, 5, and 6 from the above list are the ones that are most commonly used in oceanography. We will look at case 5 in some more detail as this method is typically used to in oceanographic instruments that measure temperature *in situ*.

Changing the resistance of a conductor

The electrical resistance of metals and other materials varies with temperature; hence these materials can be used as a temperature sensor.

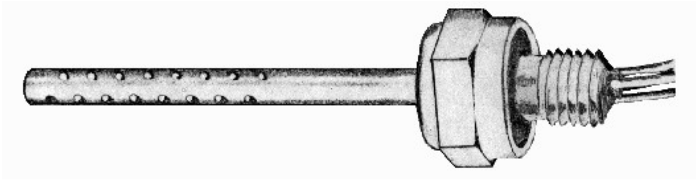
The resistance (R) of most metals depends on temperature and can be expressed as a polynomial:

$R = R\,(1+aT+bT^2+cT^3+\ldots..)$ where a, b and c are constants.

a) Wire temperature sensor

The most commonly used metals are copper, platinum and nickel. See details of their measurement ranges on the web; e.g.,:

https://www.electronics-notes.com/articles/basic_concepts/resistance/electrical-resistivity-table-materials.php



e.g., Rosemont sensor

b) Thermistor temperature sensor

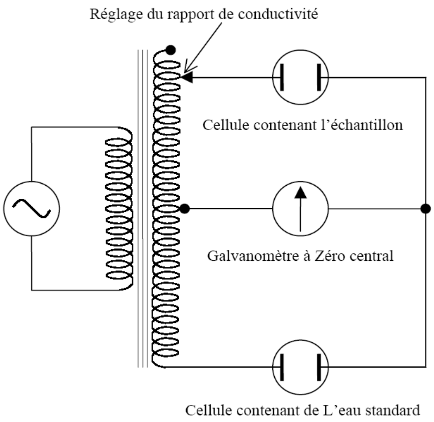
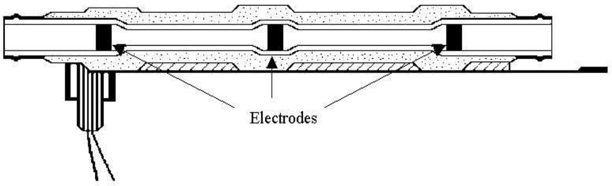
There is another class of materials used for temperature measurement: **semiconductors**.

These are mixtures of metal oxides such as Nickel, Cobalt, Manganese,…
They are obtained by sintering (manufacturing process consisting in heating the powder until the grains weld together, but without fusion) metal oxide powder at high temperatures.
The types of semiconductors used in oceanography are called **thermistors**.

Examples of use:
1) You can combine thermistors in a chain (suspended under drifting buoys or in a mooring)
2) the most standard use is the XBT: eXpandable BathyThermograph.
XBTs were developed to deliver temperature profiles of the oceanic surface layer without having to stop the ship.

Problem associated with these measurements include
- memory (see course Supplement - Reminder - Sampling)
- protection against pressure effects: depth measurements are affected by pressure,
hence the introduction of a potential temperature.

2. Conductivity by measuring seawater conductance between two electrodes

In the lab: These sensors make it possible to measure the conductivity ratio between standard seawater and the sample to be analysed at a known temperature. In order to avoid polarization of the electrodes, the measurement is made using alternating current.	
In the field: e.g.: SEABIRD conductivity sensor The cell consists of a quartz tube in containing 3 inert ring-shaped platinum electrodes. The electrodes near the ends of the tube are connected to each other. The conductance between the central electrode and the 2 others is connected in an oscillating circuit whose frequency varies depending on the conductance and therefore the conductivity of the sea water. As the two outer electrodes are connected, there is no current flowing outside the cell. **These cells need to be rinsed with demineralised water.**	 Capteur de conductivité à électrodes SEABIRD. **Precision 0.005** **but the pumping introduces an time offset between the T and C measurements which needs to be corrected**

3. Pressure

There are different types of instruments to measure pressure or pressure differences.
They typically fall into two categories: some use the principle of "hydrostatic equilibrium", others the principle of "elastic deformation under the action of pressure forces".

The SI unit is the Pascal (Pa). The unit of pressure most commonly used by manufacturers of pressure sensors is the PSI (pounds per square inch). 1 PSI = 6894.8103 Pa. Oceanographers mainly use decibars.

1 bar = 10 dbar = 10^5 Pa and 1 Pa = 1 kg $m^{-1}s^{-2}$ = 1 N/m^2

Standard atmospheric pressure is defined as the pressure at sea level which produces a 760 mm rise in a column of mercury, or a pressure of 1013 hPa, assuming ρ= 1.225 kg/m^3 as air density and T=15° C or 288 K as temperature.

1 atm = 1013 hPa = 1.013×10^5 Pa

The instruments based on the principle of "hydrostatic equilibrium" include mercury barometers, manometers, and piezometers. Instruments using the principle of "elastic deformation under pressure" include the mechanical and Bourdon manometers. There are also piezoelectric pressure sensors. Piezoelectricity refers to the ability of certain solid materials such as ceramics, crystals (quartz, barium titanate, etc.), or even biological matter such a bone, to become electrically charged when subjected to a mechanical stress (or pressure in this case) (for details see https://en.wikipedia.org/wiki/Piezoelectricity)

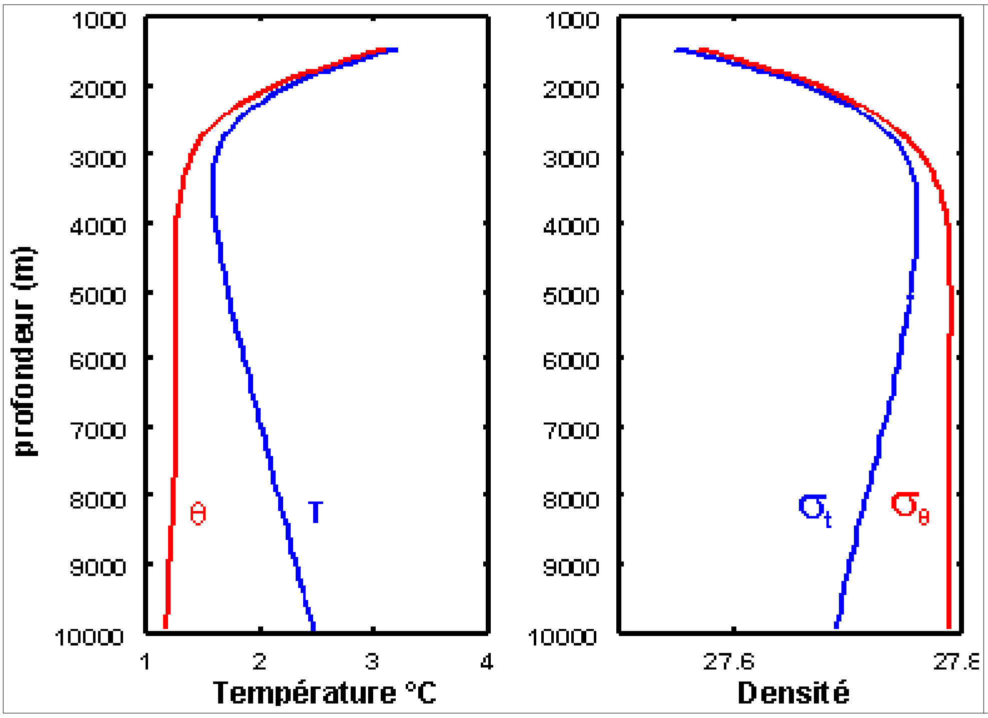
In Seabird CTDs, the pressure is typically measured by two sensor types:

- on “small” autonomous CTDs (type SBE19), the sensor measures mechanical deformation (“Paine strain-gauge sensor”);
- on “larger” CTDs (type SBE9+) used in combination with Niskin carousels, a piezoelectric sensor is used (“Paroscientific Digiquartz pressure sensor”).

4. Derived quantities

4.1 Potential temperature and conservative temperature

The potential temperature, θ , of a fluid parcel at pressure P is the temperature that this water parcel would attain if adiabatically (i.e., without heat exchange) brought to the surface (at standard reference pressure P_0=1013 hPa) (and without salt exchange).

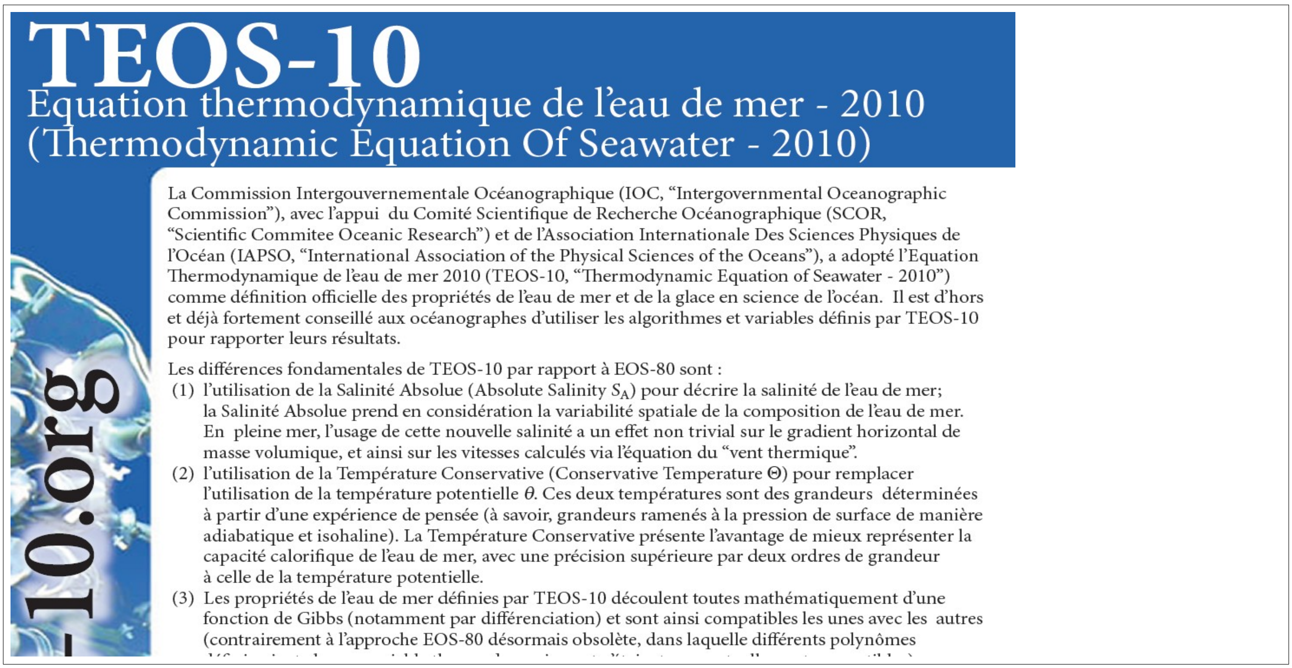


comparison of in situ *and potential temperatures in the Mindanao Trench*

http://lecalve.univtln.fr/oceano/fiches/fiche3C.htm

Since 2010, the IOC, SCOR, and IAPSO recommend the use of TEOS 10 (Thermodynamic Equation Of Seawater) and the evaluation of quantities commonly used in oceanography with new algorithms.

They recommend the use of Conservative Temperature, Θ (also denoted CT), to replace potential temperature, θ. Both quantities can be determined from a thought experiment (see above).



TEOS-10

Equation thermodynamique de l'eau de mer - 2010
(Thermodynamic Equation Of Seawater - 2010)

La Commission Intergouvernementale Océanographique (IOC, "Intergovernmental Oceanographic Commission"), avec l'appui du Comité Scientifique de Recherche Océanographique (SCOR, "Scientific Commitee Oceanic Research") et de l'Association Internationale Des Sciences Physiques de l'Océan (IAPSO, "International Association of the Physical Sciences of the Oceans"), a adopté l'Equation Thermodynamique de l'eau de mer 2010 (TEOS-10, "Thermodynamic Equation of Seawater - 2010") comme définition officielle des propriétés de l'eau de mer et de la glace en science de l'océan. Il est d'hors et déjà fortement conseillé aux océanographes d'utiliser les algorithmes et variables définis par TEOS-10 pour rapporter leurs résultats.

Les différences fondamentales de TEOS-10 par rapport à EOS-80 sont :

(1) l'utilisation de la Salinité Absolue (Absolute Salinity S_A) pour décrire la salinité de l'eau de mer; la Salinité Absolue prend en considération la variabilité spatiale de la composition de l'eau de mer. En pleine mer, l'usage de cette nouvelle salinité a un effet non trivial sur le gradient horizontal de masse volumique, et ainsi sur les vitesses calculés via l'équation du "vent thermique".

(2) l'utilisation de la Température Conservative (Conservative Temperature Θ) pour remplacer l'utilisation de la température potentielle θ. Ces deux températures sont des grandeurs déterminées à partir d'une expérience de pensée (à savoir, grandeurs ramenés à la pression de surface de manière adiabatique et isohaline). La Température Conservative présente l'avantage de mieux représenter la capacité calorifique de l'eau de mer, avec une précision supérieure par deux ordres de grandeur à celle de la température potentielle.

(3) Les propriétés de l'eau de mer définies par TEOS-10 découlent toutes mathématiquement d'une fonction de Gibbs (notamment par différenciation) et sont ainsi compatibles les unes avec les autres (contrairement à l'approche EOS-80 désormais obsolète, dans laquelle différents polynômes

The potential enthalpy* can be evaluated in the same way (i.e., brought to the surface pressure in an adiabatic and isohaline way). The conservative temperature is simply the potential enthalpy divided by the heat capacity Cp (Cp = 3991.86795711963 $J\,kg^{-1}\,K^{-1}$).

* Reminder: the enthalpy corresponds to the total energy of a thermodynamic system. It includes the internal energy, i.e., the energy necessary to create the system, plus the work this system must exert against the external pressure to occupy its volume. Enthalpy is a state function, whose unit of measurement in the International System of Units (SI) is the Joule (J).

The conservative temperature, Θ, has the advantage that it is a better representation of the heat capacity of seawater, with an accuracy two orders of magnitude better than potential temperature θ. Turbulence destroys the potential temperature but generally leaves enthalpy intact; therefore enthalpy is more conservative than potential temperature by about two orders of magnitude in general (McDougall, JPO, 2003, Potential Enthalpy: A Conservative Oceanic Variable for Evaluating Heat Content and Heat Fluxes). This is why the use of conservative temperature, Θ, has been recommended by TEOS.

The new functions to calculate these values can be found on the site http://www.teos-10.org/software.htm#1 (Note: If you use programs from the GSW Oceanographic Toolbox you should include a citation to McDougall and Barker (2011), in the form: McDougall, TJ and PM Barker, 2011: Getting started with TEOS-10 and the Gibbs Seawater (GSW) Oceanogr. Toolbox, 28pp., SCOR/IAPSO WG127, ISBN 978- 0-646-55621-5).

Θ can be evaluated from the *in situ* temperature with the **gsw_CT_from_t** function; if the absolute salinity is known: CT = gsw_CT_from_t(SA,T,p), with p in decibars.
The difference between conservative and potential temperature can be as much as $\theta-\Theta=-1.4\,^{\circ}C$ for very warm fresh water ($\Theta>30\,^{\circ}C$). For typical seawater this difference is much smaller, of the order of +/- 0.1 °C.

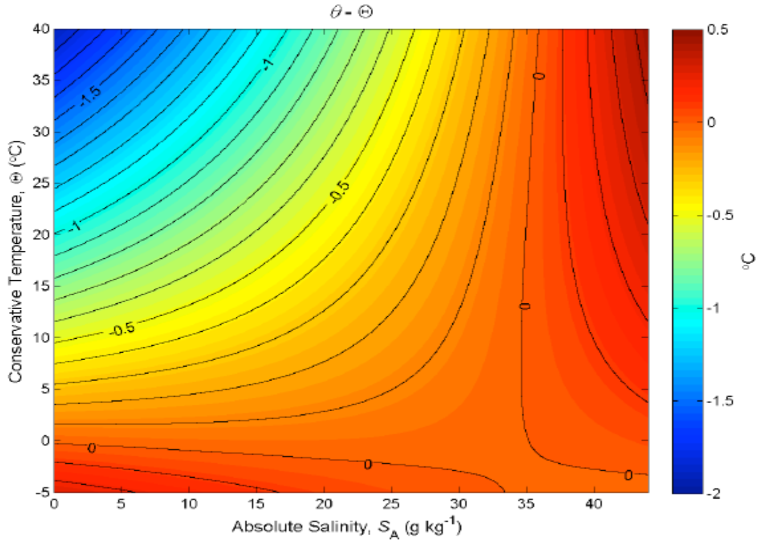


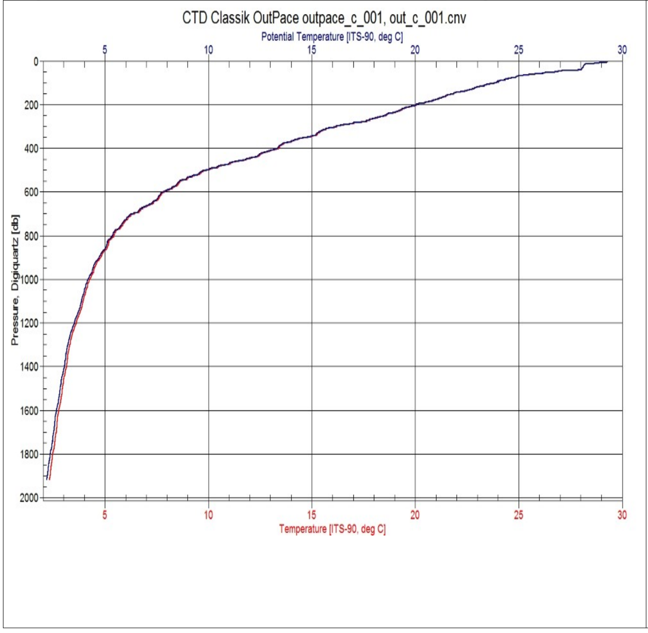
Figure A.17.1. Contours (in °C) of the difference between potential temperature and Conservative Temperature $\theta - \Theta$. This plot illustrates the non-conservative production of potential temperature θ in the ocean.

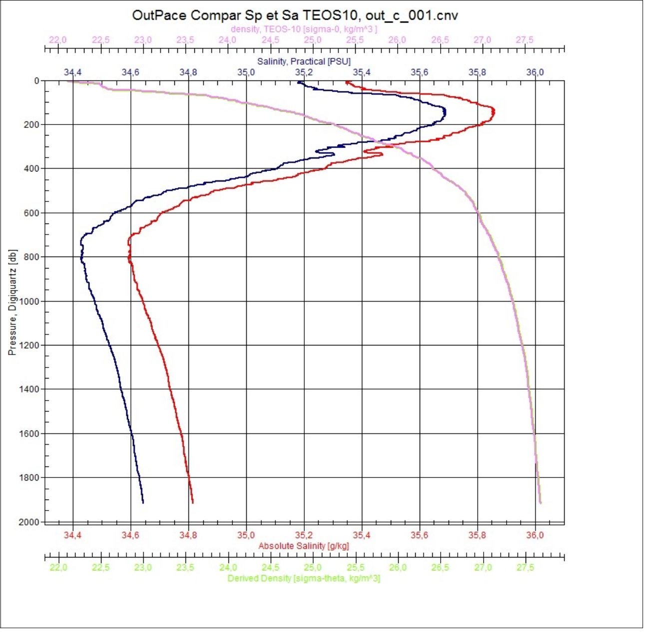
Example comparisons (courtesy G. Rougier) between conservative temperature, denoted CT, potential temperature, Theta, and the *in situ* temperature, T:

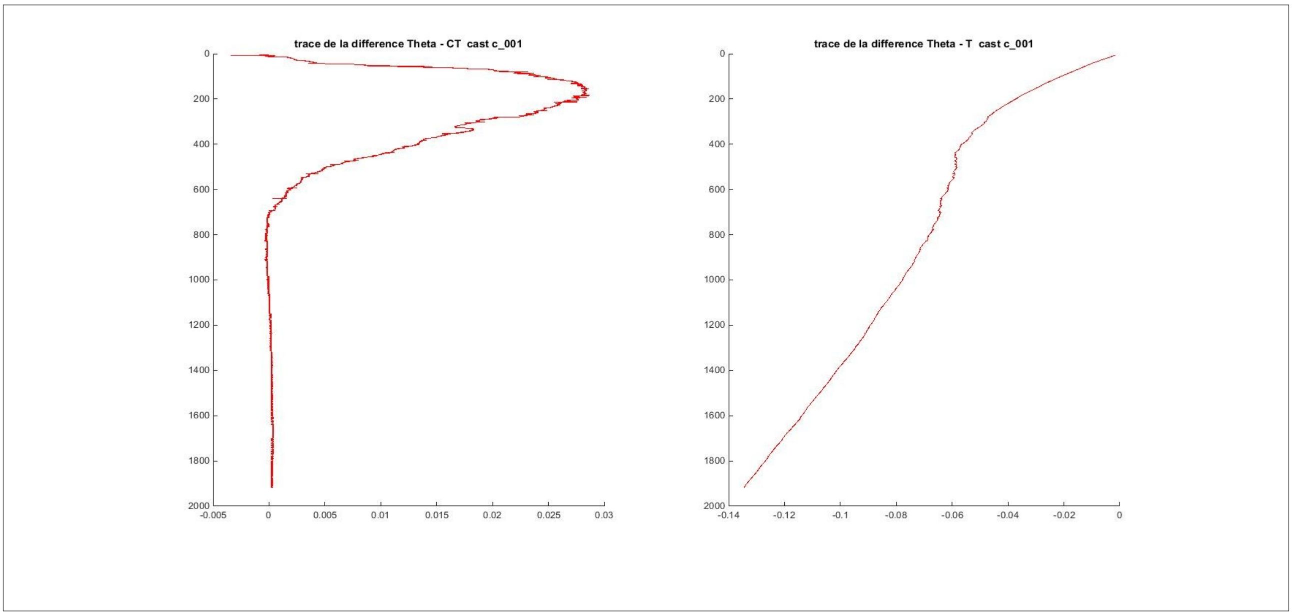
p= 100 dbars; S=35 T(ITS90)=20°C Theta=19.982°C and CT=19.974°C
p=1000 dbars; S=35 T(ITS90)=10°C Theta= 9.879°C and CT= 9.869°C
p=2000 dbars; S=35 T(ITS90)= 2°C Theta= 1.862°C and CT= 1.860°C
p=2000 dbars; S=35 T(ITS90)= 7°C Theta= 6.789°C and CT= 6.780°C
p=6000 dbars; S=35 T(ITS90)= 2°C Theta= 1.414°C and CT= 1.412°C

Typically, the differences between T and Theta increase with depth while the differences between Theta and CT decrease with depth.

E.g.: Station C_001 Outpace Campaign 2015 SW Pacific (PIs: T. Moutin and S. Bonnet)







4.2 Salinity

Theoretical definition:

Salinity is the weight in grams of solid residue contained in one kilogram of seawater when all carbonates have been transformed into oxides, bromine and iodine replaced by chlorine, and all organic matter completely oxidized.

It is very difficult to estimate the salinity by direct chemical analysis (drying and weighing of the solid residue), because certain components, in particular the chlorides, tend to escape during the last stage of drying. Since this determination is difficult to achieve, a simpler method has been devised.

Definition of Practical Salinity:

Since 1978, the practical salinity scale defines salinity in terms of conductivity. The practical salinity, S, of a sample of sea water was defined as a function of the ratio K of the electrical conductivity of this sample at 15°C at standard atmospheric pressure and the conductivity of a potassium chloride solution in which the mass fraction of KCl is 0.0324356, at the same temperature and same pressure. A value of K equal to 1 corresponds, by definition, to a practical salinity of 35 (**without units** – although you may sometimes see it with the suffix psu or PSU for practical salinity units).

The corresponding formula is (valid for $2 \le S \le 42$):

$$S = 0.0080 - 0.1692\,K^{1/2} + 25.3853\,K + 14.0941\,K^{3/2} - 7.0261\,K^{2} + 2.7081\,K^{5/2} + \text{term}(T).$$

Since 2010, the IOC, SCOR, and IAPSO recommend the use of the TEOS 10 (Thermodynamic Equation Of Seawater) standard for calculating absolute salinity (Sa) and practical salinity (Sp), which differs from the previous method.

All the details concerning these salinities have been introduced in the Chemical Oceanography 1 course. You can also download the programs to do the calculations from this website: http://www.teos-10.org/software.htm#1

e.g.: `[PSAL_SA,in_ocean]=gsw_SA_from_SP(PSAL,p,longitude,latitude);`
`knowing the pressure in decibars, longitude, and latitude.`

Knowing the location of a measurement is necessary to calculate the absolute salinity as this calculation depends on an archive of previously measure data (at that or nearby locations).

4.3 Density (also known as specific mass or volumetric mass density)

This is the most widely used derived quantity and can be calculated by an algorithm $\rho(S, T, p)$ known as the International Equation of State of Seawater. For a long time, it was used in its 1980 version (EOS-80).

Since 2010, the IOC, SCOR, and IAPSO recommend the use of a different version $\rho(S_A, T_\Theta, p)$. These recommendations are also valid for the following derived quantities (all new functions can be found on this website: http://www.teos-10.org/software.htm#1).

4.4 Specific volume α

This is the inverse of density: $\alpha(S,T,p) = 1/\rho(S,T,p)$

4.5 Specific volume anomaly

This is the difference between the specific volume of sea water at S,T,p and normal water with a salinity of 35.00 and a temperature of 0°C subjected to the same pressure, i.e.: $\delta(S,T,p) = \alpha(S,T,p) - \alpha(35,0,p)$ in $m^3\,kg^{-1}$. This quantity is used in dynamic methods (calculation of geostrophic currents from vertical profiles of S, T, and p; c.f., Ocean Dynamics L3).

4.6 Relative density

The relative density of a fluid is defined as the ratio between the density of the fluid and the density water (or air when the fluid is a gas) at standard conditions (P = 1 atm, T = 3.98 °C for water, T=0°C for air).

$$d = \frac{\rho}{\rho_{st}}$$

It is a dimensionless number.

Since the density of seawater is typically very close to the density of standard water (density of standard water ρ_{st} = 999,975 kg m^{-3}) the relative density of seawater is usually close to 1. For example, if d = 1.02872, it is important to include digits up to the fifth decimal.

4.7 Sigma

Sigma is a type of notation used in oceanography to measure the density of seawater. As density is usually close to 1000 $kg\,m^{-3}$ in SI units (with at least 2 or 3 important decimals) the quantity of σ was introduce to simplify the notation:

$$\sigma(S,T,p) = \rho(S,T,p) - 1000\,kg\,m^{-3}$$

This allows a density of $\rho = 1025.748\,kg\,m^{-3}$ to be written quite simply as: σ = 25.748

There is also:

$$\sigma_t = \sigma(S,T,0) = \rho(S,T,0) - 1000\,kg\,m^{-3}$$

$$\sigma_\theta = \sigma(S,\theta,0) = \rho(S,\theta,0) - 1000\,kg\,m^{-3}$$ and

$$\sigma_\Theta = \sigma(Sa,\Theta,0) = \rho(Sa,\Theta,0) - 1000\,kg\,m^{-3}$$

Summary:

	English
ρ	density
d	relative density
σ	sigma

4.8 The speed of sound

The speed of propagation C of a vibration in a fluid is given by the Newton-Laplace equation:

$$C=\sqrt{\frac{1}{\rho\,\kappa_a}}$$

where κ_a is the adiabatic compressibility coefficient of seawater, related to the isothermal compressibility κ_T , the thermal expansion coefficient α_V , density ρ , and heat capacity C_p .

The speed of sound increases with increases in water temperature, or in pressure (depth), or in - to a lesser degree - salinity.

(these variations are going in the same direction as density and sigma except for temperature: when temperature increase, the speed of sound increases while density and sigma decrease).

Calculating the speed of sound in seawater

Method by Chen et Millero (1977):

$C(S,t,p) = C_w(t,p) + A(t,p)\, S + B(t,p)\, S^{3/2} + D(t,p)\, S^2$

S salinity, t temperature, p pressure

parameters: Cw, A, B, D defined below:

$Cw(t,p) = C_{00} + C_{01}\, t + C_{02}\, t^2 + C_{03}\, t^3 + C_{04}\, t^4 + C_{05}\, t^5$

$+ (C_{10} + C_{11}\, t + C_{12}\, t^2 + C_{13}\, t^3 + C_{14}\, t^4)\, p$

$+ (C_{20} + C_{21}\, t + C_{22}\, t^2 + C_{23}\, t^3 + C_{24}\, t^4)\, p^2$

$+ (C_{30} + C_{31}\, t + C_{32}\, t^2)\, p^3$

C_{00} = + 140388

C_{01} = +5.03711

C_{02} = - 5.80852 E-2

C_{03} = +3.3420 E-4

C_{04} = -1.47800 E-6

C_{05} = + 3.1464 E-9

C_{20} = + 3.1260 E-5

C_{21} = - 1.7107 E-6

C_{22} = + 5974 E-8

C_{23} = - 5335 E-10

C_{24} = + 1.0405 E-12

C_{10} = + 0.153563

C_{11} = + 6.8982 E-4

C_{12} = - 8.1788 E-6

C_{13} = + 1.3621 E-7

C_{14} = - 6.1185 E-10

C_{30} = -9.7729 E-9

C_{31} = + 3.8504 E-10

C_{32} = - 3643 E-12

$A(t,p) = A_{00} + A_{01}\, t + A_{02}\, t^2 + A_{03}\, t^3 + A_{04}\, t^4$

$+ (A_{10} + A_{11}\, t + A_{12}\, t^2 + A_{13}\, t^3 + A_{14}\, t_4)\, p$

$+ (A_{20} + A_{21}\,t + A_{22}\,t^2 + A_{23}\,t^3)\,p^2$
$+ (A_{30} + A_{31}\,t + A_{32}\,t^2)\,p^3$

A_{00} = + 1.389
A_{01} = - 1.262 E-2
A_{02} = + 7.164 E-5
A_{03} = + 006 E-6
A_{04} = - 3.21 E-8
A_{20} = - 3.9064 E-7
A_{21} = + 9.1041 E-9
A_{22} = + 1.6002 E-10
A_{23} = + 7.988 E-12

A_{10} = + 9.4742 E-5
A_{11} = - 1.2580 E-5
A_{12} = - 6.4885 E-8
A_{13} = + 1.0507 E-8
A_{14} = - 0122 E-10
A_{30} = + 1.100 E-10
A_{31} = + 6.649 E-12
A_{32} = - 3.389 E-13

$B(t,p) = B_{00} + B_{01}\,t + (B_{10} + B_{11}\,t)\,p$

B_{00} = - 1.922 E-2
B_{01} = - 4.42 E-5

B_{10} = + 7.3637 E-5
B_{11} = + 1.7945 E-7

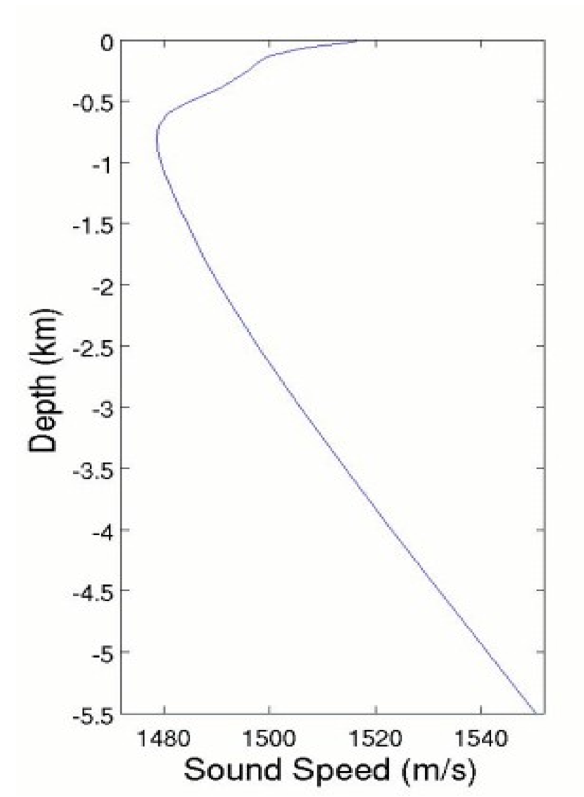
$D(t,p) = D_{00} + D_{10}\,p$

D_{00} = + 1.727 E-3
D_{10} = - 7.9836 E-6

To verify (S unitless, t in °C, and p in bars):
C (30,0,0) = 1445 m/s
C (35,10,500) = 1573.4 m/s
C (40,40,1000) = 1730 m/s

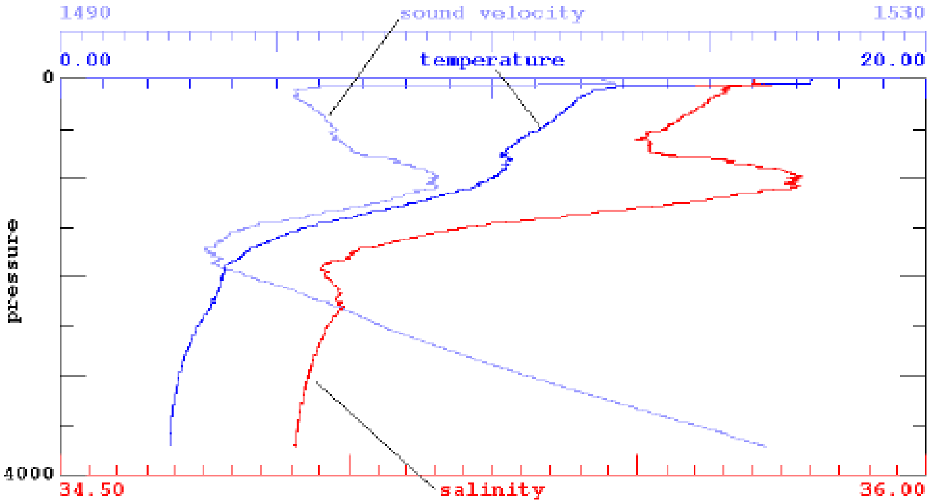
The absolute error is about: Error in C(S,t,p) = 0.8 S + 0.3 t + 6 p
speed is in m/s, S unitless, t in °C, and p in bars

The SOFAR channel (**SOund Fixing And Ranging**) is the layer in the water column where the sound velocity is minimal and where sounds can therefore travel great distances.



Vertical profile

Along a vertical profile, salinity can vary by a few units, temperature by a few tens of degrees, and the pressure by a few hundred bars. **The speed of sound varies mostly due to changes in pressure, some due to changes in temperature (especially near the surface), and a little due to changes in salinity**. When depth increases, c increases with pressure. But, as the temperature decreases significantly between the surface and approximately 1500 m, this effect dominates and leads to a decrease in velocity. For depths beyond 1500m, the temperature changes only very little and the effects of the increasing pressure become dominant leading to an increase in sound velocity. **There is therefore a minimum speed typically located around a depth of 1500 m (the so-called SOFAR channel)**.



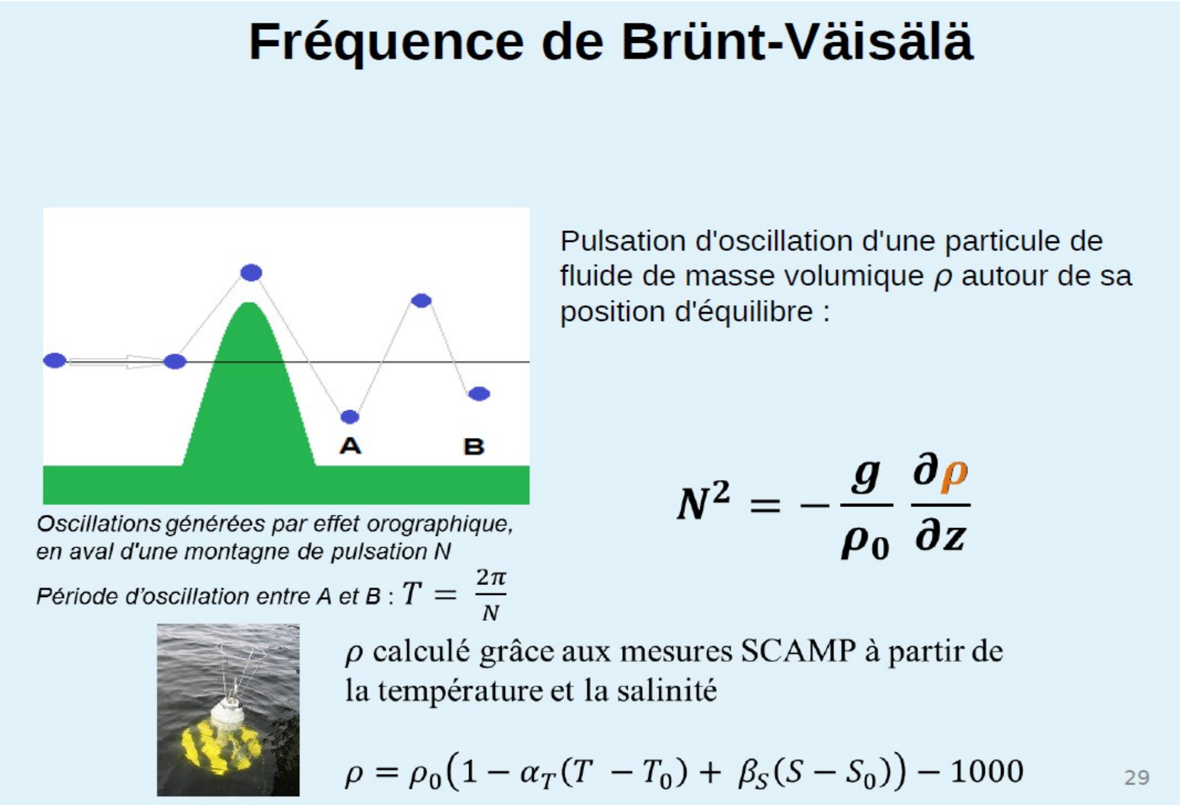
An atypical example of vertical temperature, salinity, and sound velocity profiles with depth (pressure).

A simplified formula for calculating the speed of sounds is (Urick, 1983):
$C = 1449.2 + 4.6T - 0.055T^2 + 0.00029T^3 + (1.34 - 0.01T)(S - 35) + 0.016D$
with:
T temperature in ºC
S salinity in parts per thousand (‰) [old salinity units]
D depth in metres

4.9 Measuring turbulence

E.g., with a SCAMP (Self Contained Autonomous MicroProfiler) manufactured by PME Precision Measurements Engineering, Inc. (http://www.pme.com/)
reminder – units and values

ν	$\nu = 1,9.10^{-6}$	Viscosité cinématique	$\mathrm{m^2\ s^{-1}}$
ρ_0	$\rho_0 = 1025$	Masse volumique moyenne	$\mathrm{kg\ m^{-3}}$
R_f	$R_f = \frac{N^2}{Pr_T\left[\left(\frac{\partial \bar{u}_1}{\partial x_3}\right)^2+\left(\frac{\partial \bar{u}_2}{\partial x_3}\right)^2\right]}$	Nombre de Richardson de flux	sans unité
Pr_T	$Pr_T \approx 1$	Nombre de Prandtl	sans unité
N^2	$N^2 = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial x_3}$	Fréquence de Brunt-Väisälä	$\mathrm{rad\ s^{-1}}$
κ_T	$\kappa_T = 1.10^{-7}$	Diffusivité thermique moléculaire	$\mathrm{m^2\ s^{-1}}$
K_B	measured	Batchelor wave number	$\mathrm{rad\ m^{-1}}$
L_B	$L_B = 2\pi/K_B$	Batchelor length scale	m
ϵ_B	$\epsilon_B = \frac{\kappa_T^2 \nu}{L_B^4}$	Taux de dissipation moyen de l'énergie cinétique moyenne (définition de Luketina and Imberger (2000))	$\mathrm{m^2\ s^{-3}}$



Fréquence de Brünt-Väisälä

Oscillations générées par effet orographique, en aval d'une montagne de pulsation N

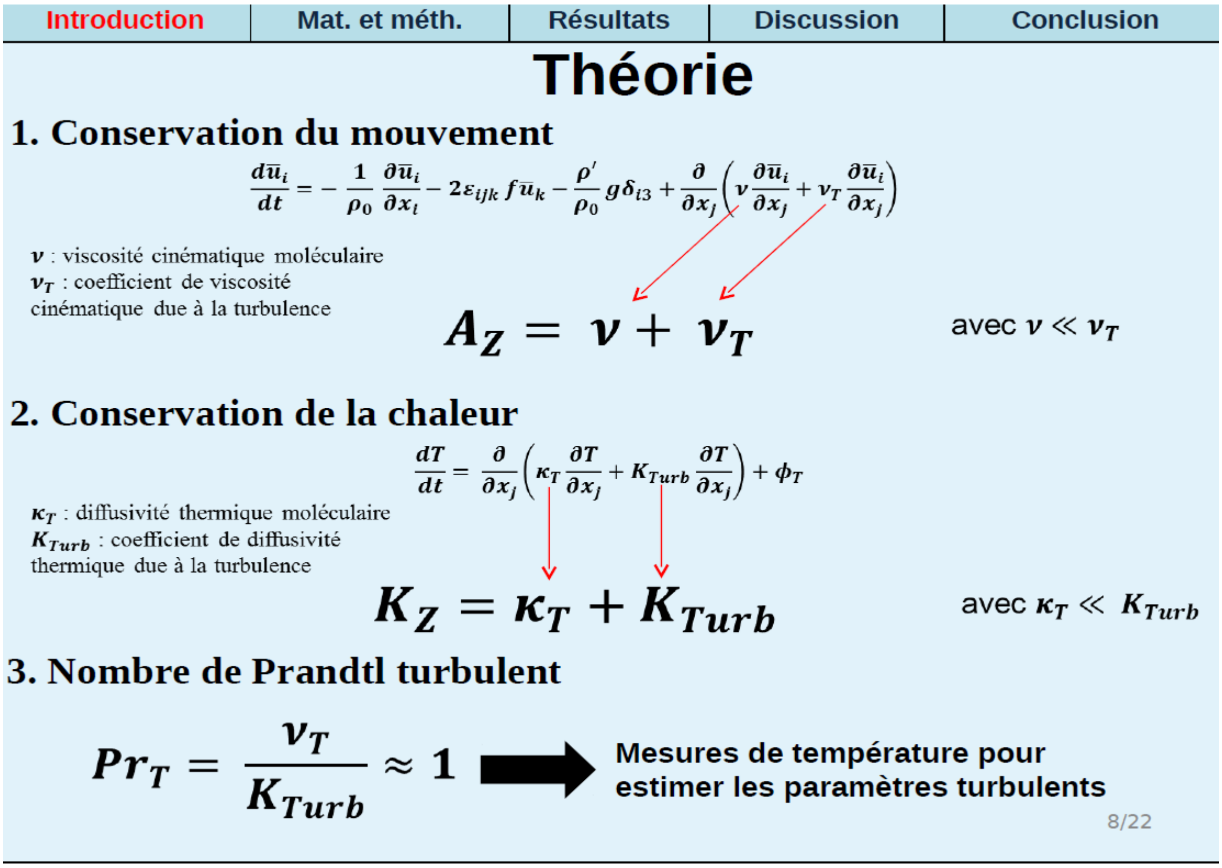
Période d'oscillation entre A et B : $T = \frac{2\pi}{N}$

Pulsation d'oscillation d'une particule de fluide de masse volumique ρ autour de sa position d'équilibre :

$$N^2 = -\frac{g}{\rho_0}\frac{\partial \rho}{\partial z}$$

ρ calculé grâce aux mesures SCAMP à partir de la température et la salinité

$$\rho = \rho_0\left(1 - \alpha_T(T - T_0) + \beta_S(S - S_0)\right) - 1000$$



Introduction | **Mat. et méth.** | **Résultats** | **Discussion** | **Conclusion**

Théorie

1. Conservation du mouvement

$$\frac{d\bar{u}_i}{dt} = -\frac{1}{\rho_0}\frac{\partial \bar{u}_i}{\partial x_i} - 2\varepsilon_{ijk} f \bar{u}_k - \frac{\rho'}{\rho_0} g \delta_{i3} + \frac{\partial}{\partial x_j}\left(\nu \frac{\partial \bar{u}_i}{\partial x_j} + \nu_T \frac{\partial \bar{u}_i}{\partial x_j}\right)$$

ν : viscosité cinématique moléculaire
ν_T : coefficient de viscosité cinématique due à la turbulence

$$A_Z = \nu + \nu_T$$

avec $\nu \ll \nu_T$

2. Conservation de la chaleur

$$\frac{dT}{dt} = \frac{\partial}{\partial x_j}\left(\kappa_T \frac{\partial T}{\partial x_j} + K_{Turb}\frac{\partial T}{\partial x_j}\right) + \phi_T$$

κ_T : diffusivité thermique moléculaire
K_{Turb} : coefficient de diffusivité thermique due à la turbulence

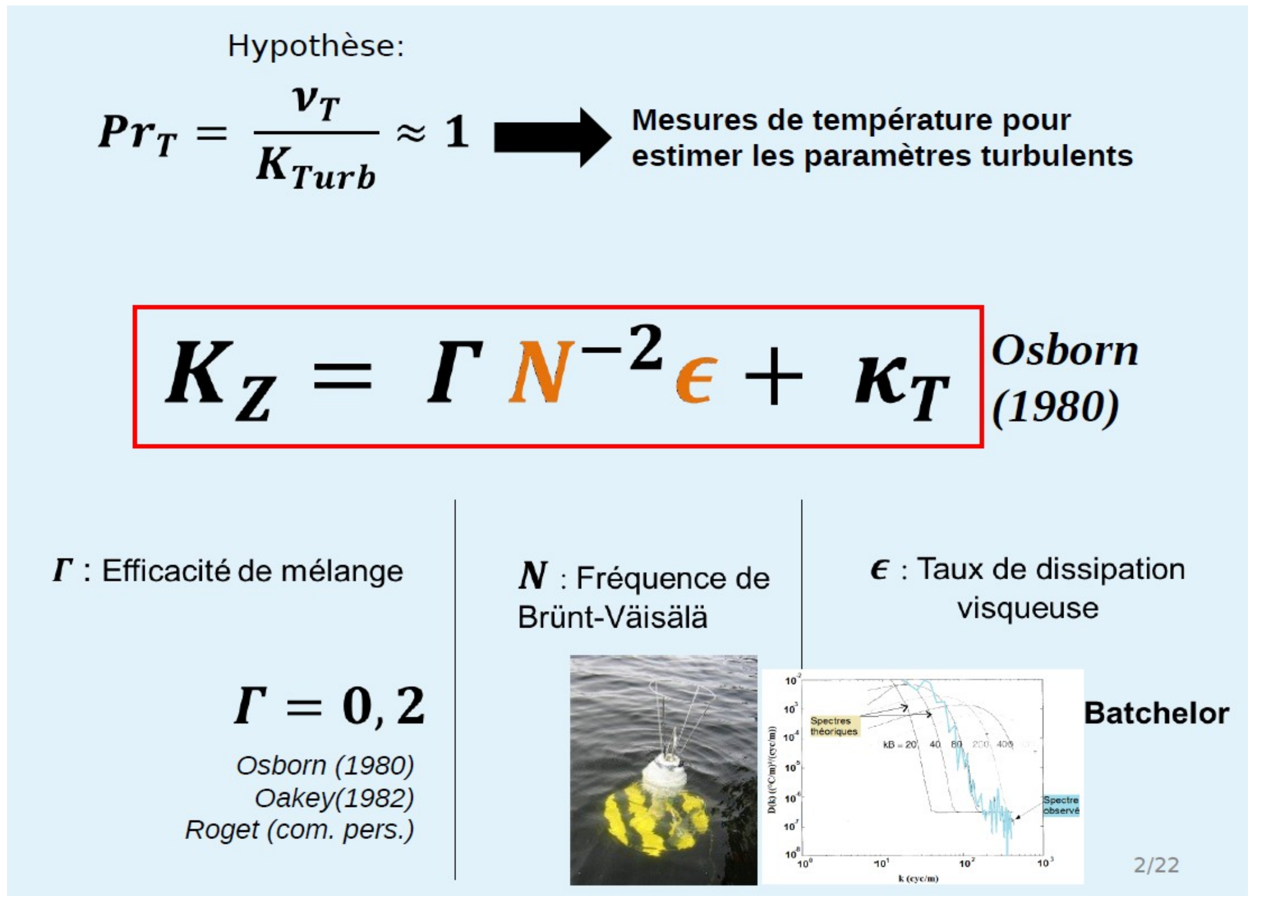
$$K_Z = \kappa_T + K_{Turb}$$

avec $\kappa_T \ll K_{Turb}$

3. Nombre de Prandtl turbulent

$$Pr_T = \frac{\nu_T}{K_{Turb}} \approx 1 \quad \Rightarrow$$

Mesures de température pour estimer les paramètres turbulents



Hypothèse:

$$Pr_T = \frac{\nu_T}{K_{Turb}} \approx 1$$

Mesures de température pour estimer les paramètres turbulents

$$\boxed{K_Z = \Gamma N^{-2} \epsilon + \kappa_T}$$

Osborn (1980)

Γ : Efficacité de mélange

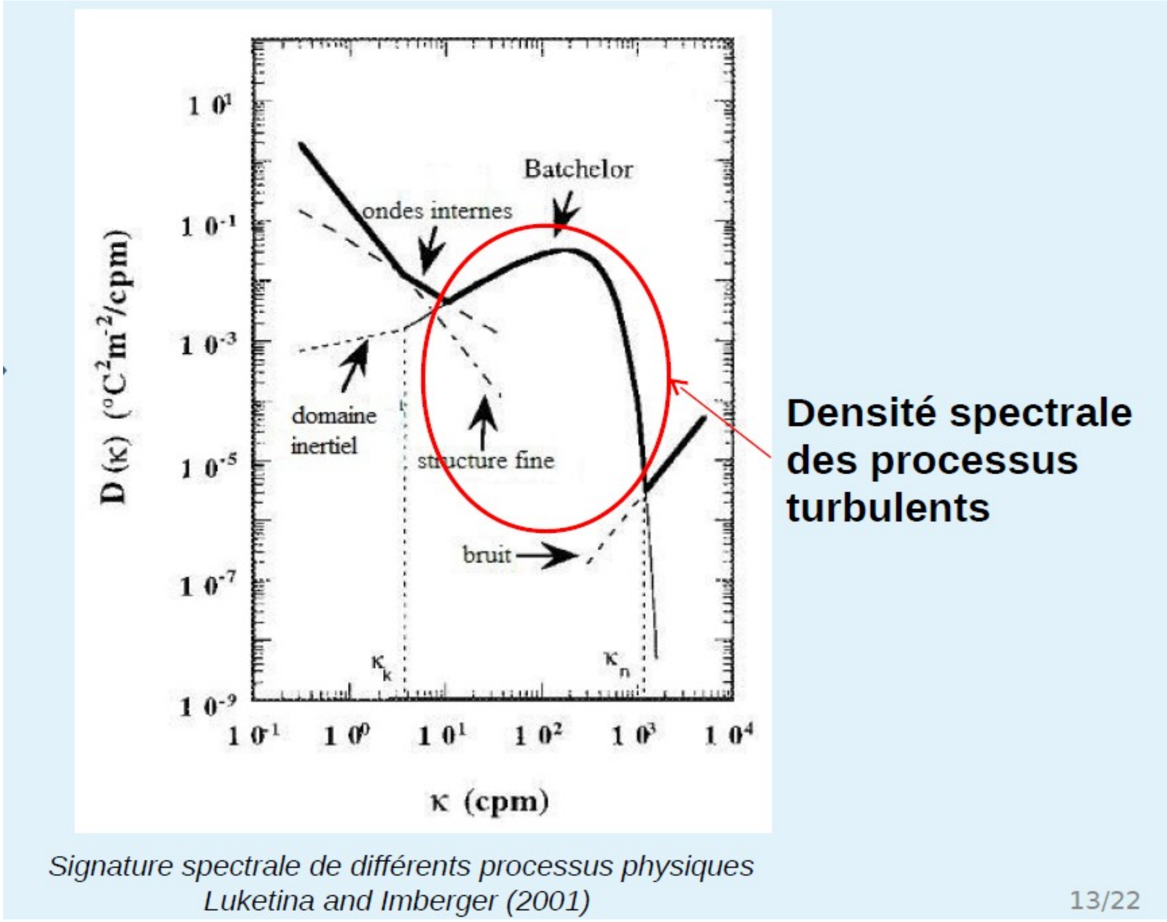
$$\Gamma = 0,2$$

Osborn (1980)
Oakey(1982)
Roget (com. pers.)

N : Fréquence de Brünt-Väisälä

ϵ : Taux de dissipation visqueuse

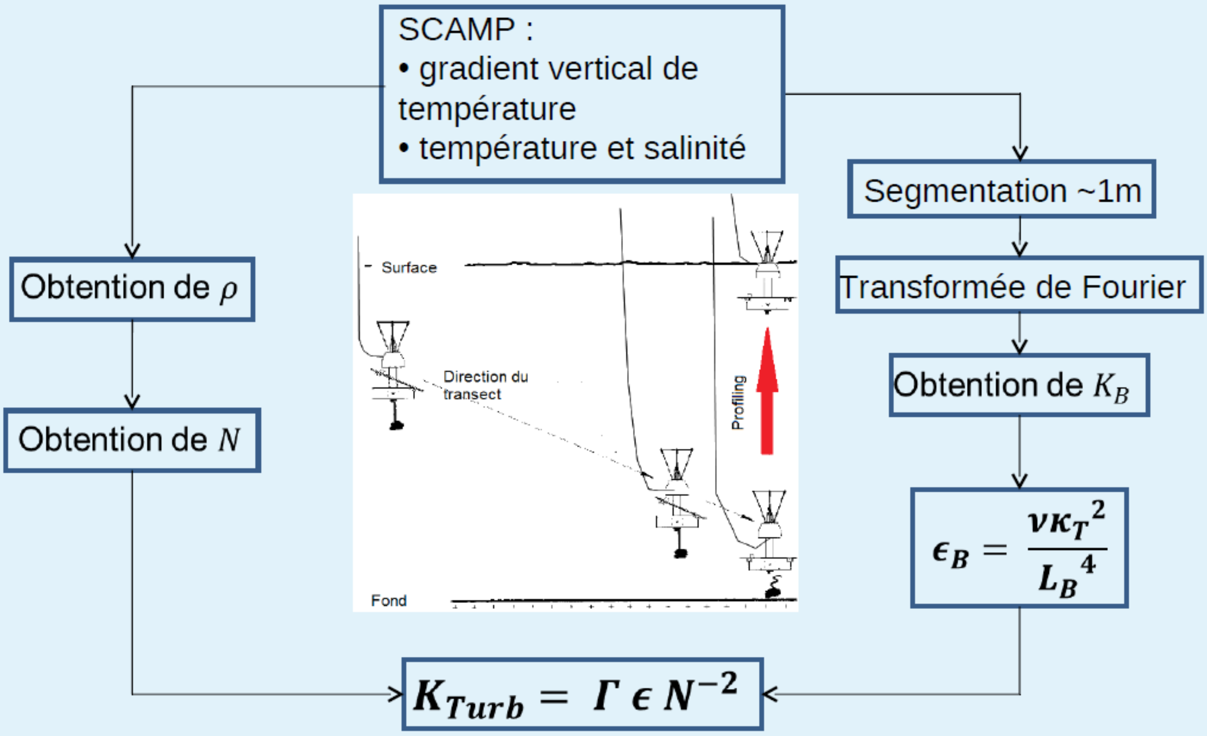
Batchelor



Densité spectrale des processus turbulents

Signature spectrale de différents processus physiques
Luketina and Imberger (2001)

Résumé de la méthode



SCAMP :
- gradient vertical de température
- température et salinité

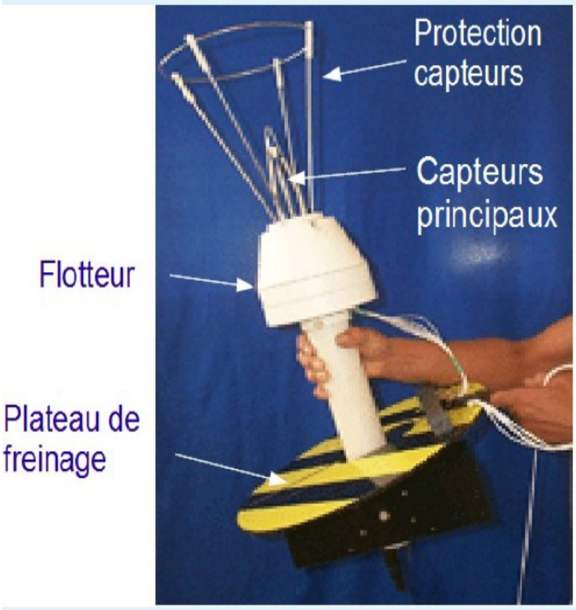
Obtention de ρ → Obtention de N

Segmentation ~1m → Transformée de Fourier → Obtention de K_B → $\epsilon_B = \dfrac{\nu \kappa_T^{2}}{L_B^{4}}$

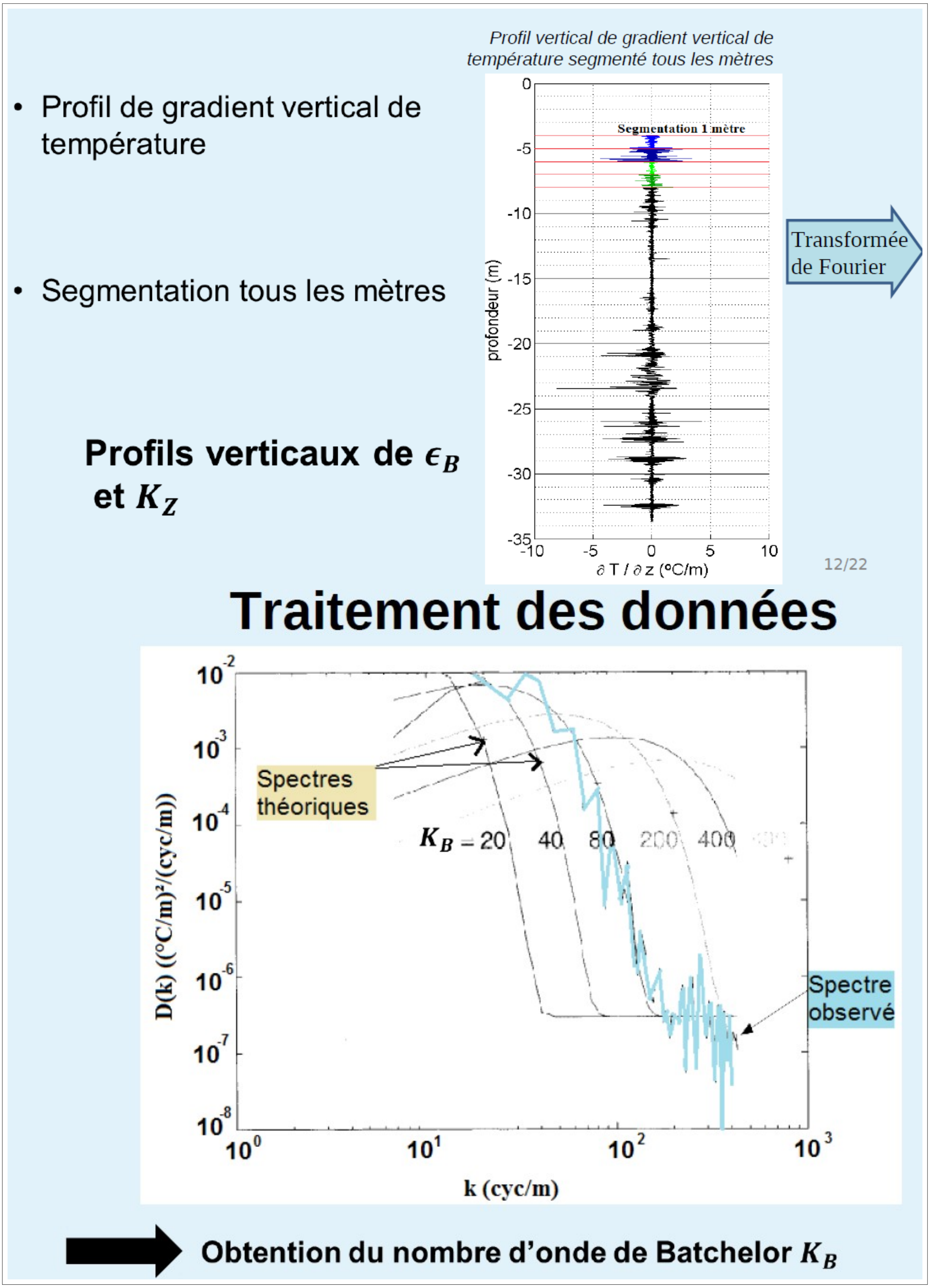
$$K_{Turb} = \Gamma\, \epsilon\, N^{-2}$$

Le SCAMP

Self Autonomous Micro Profiler



- Mesures à fine échelle de température et conductivité
- Fréquence : 100 Hz
- Vitesse de mesure : $10 - 20\ cm\ s^{-1}$
- Profondeur maximale : 100 m
- Analyse des données : Matlab



- Profil de gradient vertical de température
- Segmentation tous les mètres

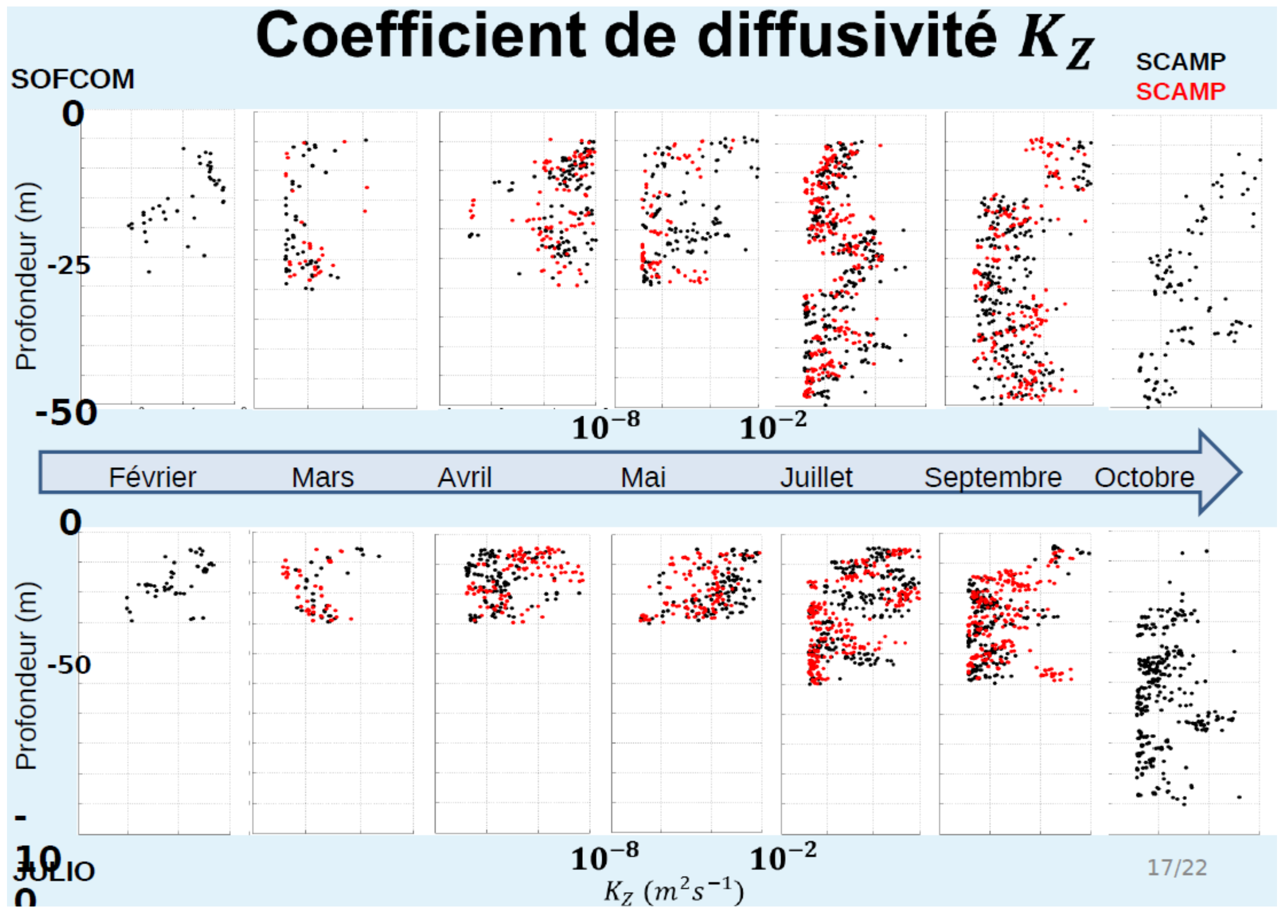
Profils verticaux de ϵ_B et K_Z

Profil vertical de gradient vertical de température segmenté tous les mètres

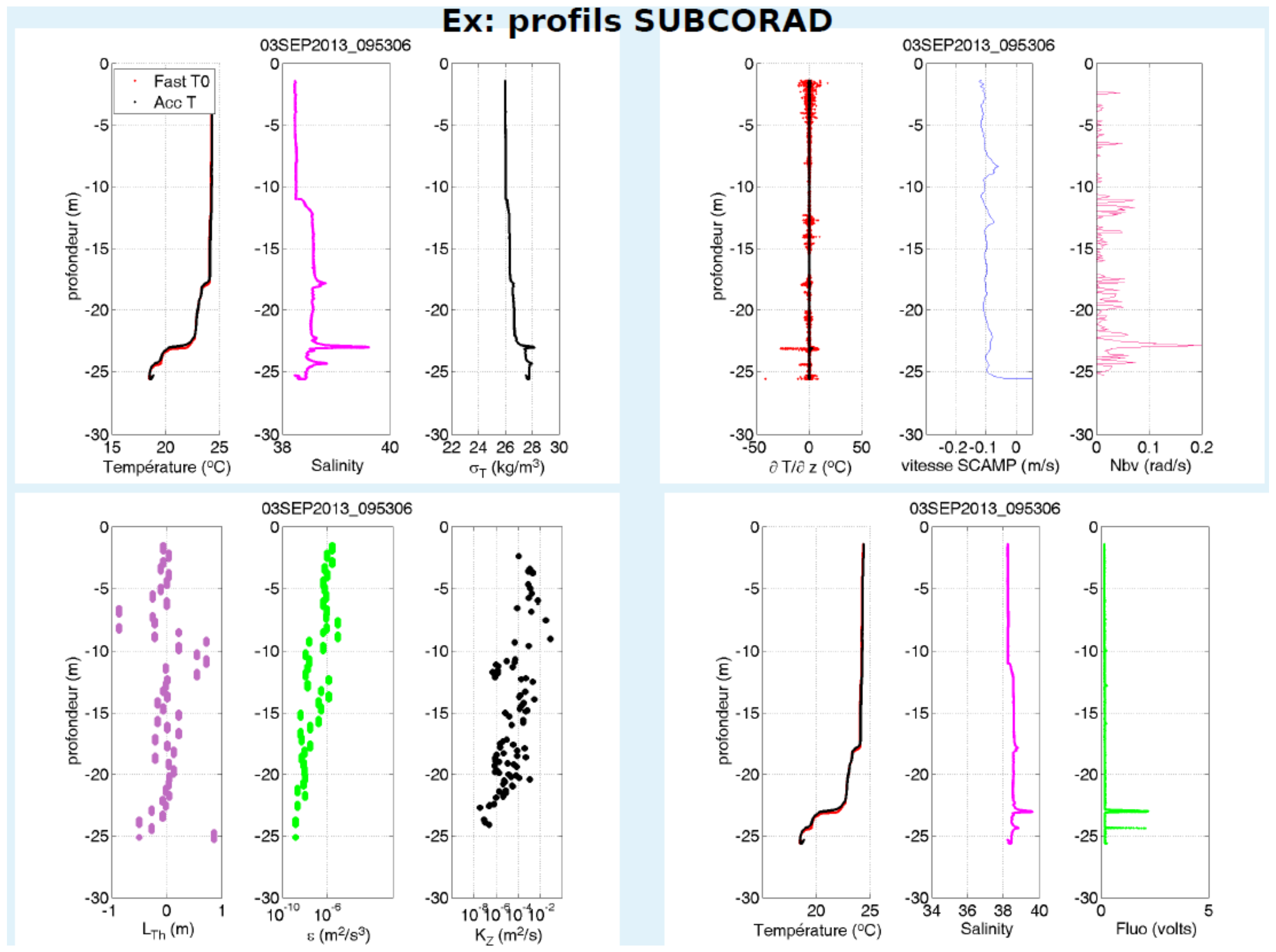
12/22

Traitement des données

Obtention du nombre d'onde de Batchelor K_B



Coefficient de diffusivité K_Z



Ex: profils SUBCORAD

5. Time

The international unit of time is the second.

Be careful to check that the time is in "Coordinated Universal Time" (UTC, sometimes UT) - which refers to universal atomic time - or "local time" (with variations during the year in some countries; example in France we have UTC+1h in winter and UTC+2h in summer).

UTC time corresponds to the time in London during winter:
at 15h UTC in winter, it is 3 p.m. in London and 4 p.m. in Paris;
at 15h UTC in summer, it is 4 p.m. in London and 5 p.m. in Paris.

Before, times would often be given in Greenwich Mean Time (GMT) instead of UTC.

Some History (see https://en.wikipedia.org/wiki/Greenwich_Mean_Time):

Greenwich Mean Time (GMT) is the mean solar time at the Royal Observatory in Greenwich (location of the 0° prime meridian), London, counted from midnight. The term 'GMT' is also used as one of the names for the time zone UTC+00:00.

English speakers often use GMT as a synonym for Coordinated Universal Time (UTC): in modern usage, this is incorrect – GMT is now a time zone (equivalent to Western European Time), not a time standard. For navigation, ~GMT could be considered equivalent to UTC; but it can differ from UTC by up to 0.9 s. The term GMT should thus not be used for purposes that require precision.

Because of Earth's uneven angular velocity in its elliptical orbit and its axial tilt, noon (12:00:00) GMT is rarely the exact moment the Sun crosses the Greenwich Meridian and reaches its highest point in the sky there. This event may occur up to 16 minutes before or after noon GMT, a discrepancy described by the equation of time. Noon GMT is the annual average (i.e. "mean") moment of this event, which accounts for the word "mean" in "Greenwich Mean Time".

For long time series measurements, the day is used. It is therefore necessary to check the conventions of measurement. To link the days one after the other, we use the notion of ordinal day or ordinal date.

An ordinal date is a calendar date usually consisting of the year and a day ranging from 1 to 366 (starting with January 1), although sometimes the year is omitted.

The two numbers are then written YYYY-DDD.

To the day of	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
Add	0	31	59	90	120	151	181	212	243	273	304	334
Leap years	0	31	60	91	121	152	182	213	244	274	305	335

(http://en.wikipedia.org/wiki/Ordinal_date)

If there we also need to provide hour, minute, and second, a decimal part can be added to this ordinal day. If we have a sequence of several years, we can put this annual system end to end. The question is to choose a reference date for the zero.

Theoretically, the so-called Julian day is the basis of a dating system consisting of counting the number of days and fractions of a day that have elapsed since a specific date, namely 1 January of the year -4712 at 12 noon.

The term Julian day is also used by CNES and NASA to date various events. Here, the number of days is counted from January 1, 1950 at midnight for CNES and from May 24, 1968 at midnight for NASA.

In Matlab, *julianday* gives the day of a year, i.e., a number between 0 (corresponding to Jan 1) and 364/365 (for Dec 31). It is very similar to the Matlab function *dayofyear* except that it starts with 0 instead of 1. These functions have no reference date. The function *datenum* has 1 Jan 0000 as reference date.

A bit of history:

you should know that although we use the term Julian day, **since 1582 we have been using the Gregorian not the Julian calendar**.

The term Julian day was chosen because the reference date (zero) predates the establishment of the Gregorian day. But the date is still calculated in the current system of the Gregorian calendar.

See for example these netcdf file headers:

```
 time
          Size:      1957x1
          Dimensions:time
           Datatype:   double
           Attributes:
                       long_name  = 'acquisition time'
                       units      = 'days since 1899-12-30 00:00:00 UTC'
                       calendar   = 'gregorian'
```

The Julian calendar was established by the astronomer Sosigene (in 46 BCE) and contained a leap day every 4 years yielding an average length of a year of 365.25 days. However, the average tropical year lasts 365.24219 days (about 365 days 5 h 48 min 45 s). This induced a lag of about 8 days per millennium with respect to true time.

The Gregorian calendar is also a solar calendar and was conceived at the end of the 16th century to correct the secular drift of the Julian calendar which was in use then. It bears the name of its instigator, Pope Gregory XIII. First adopted in 1582 by Catholic countries, then by Protestant ones, its use gradually spread throughout the world at the beginning of the 20th century. It also includes leap years (every 4 years) but with one important addition: the final years of a century (XX00) are only a leap year if their number is divisible by 400 (e.g., 1600 and 2000 were leap years, but not 1700, 1800, 1900). With this caveat, the Gregorian calendar results in an average of 365.2425 days per year (better than the Julian calendar which gives 365.254219 days), still, this yields an excess of 3 days in 10,000 years. There is no point in trying to amend the rule to get closer to the figure for the tropical year because the uncertainties associated with the length of the year over a 10,000 year period are of the same order of magnitude. (https://en.wikipedia.org/wiki/Gregorian_calendar).

With Matlab:

datenum Convert a date into a numerical date:

```
>> datenum('30.04.2019 00:00:00','dd.mm.yyyy HH:MM:SS')
ans =      737545
```

Important: you have to specify the date format, without it the result is usually wrong:

```
datenum('30.04.2019 00:00:00')
ans =      738157 which is not the correct number (see below on datestr)
```

datestr Convert numerical date to string format

```
>> datestr(737545)                          ans = '30-Apr-2019'
```

If you need to get today's date, you can use:

```
>> datestr(now,'mm/dd/yy')                  ans = '04/06/21'

>> datestr(now,'dd.mm.yyyy')                ans =  '06.04.2021'
```

If you type *help datestr* you can see Table 1: Standard MATLAB Date format definitions

e.g., format 31 'yyyy-mm-dd HH:MM:SS' 2000-03-01 15:45:17

```
>>datestr(datenum('30.04.2019','dd.mm.yyyy'),31)     ans =   '2019-04-30 00:00:00'
```

likewise

```
datestr(737545,31)                          ans = '2019-04-30 00:00:00'
```

but (see below) datestr(738157,31) = '2021-01-01 00:00:00'

which is January 1st 2021 (and not April 30 2019)

%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%%

Appendix Using TS-diagrams in Matlab

Make sure you have the script files tsdiagram.m, swfreezt.m, and swstate.m in your directory. These scripts are part of a toolbox that can be downloaded from:
https://people.mio.osupytheas.fr/~petrenko/TEACHING/MATLAB/

TSDIAGRAM([Smin Smax],[Tmin Smax],P,SIGMA) draws contour lines of density anomaly SIGMA σ at a specific pressure P (= scalar) chosen from the pressure measurements at a given station (in dbars), in a figure where the axes are fixed by the minimum and maximum salinity S (x-axis) and temperature T (y-axis), given during the function call by 2-element vectors [Smin Smax] and [Tmin Smax]. SIGMA is a vector containing values $[\sigma_1 \sigma_2 \sigma_3 \sigma_4 \sigma_5]$ that do not necessarily need to be evenly spaced although it is recommended to use regular intervals; if σ is not provided, Matlab will add some default isolines. If the temperatures are negative, the freezing points will be indicated by an additional line (using swfreezt.m):

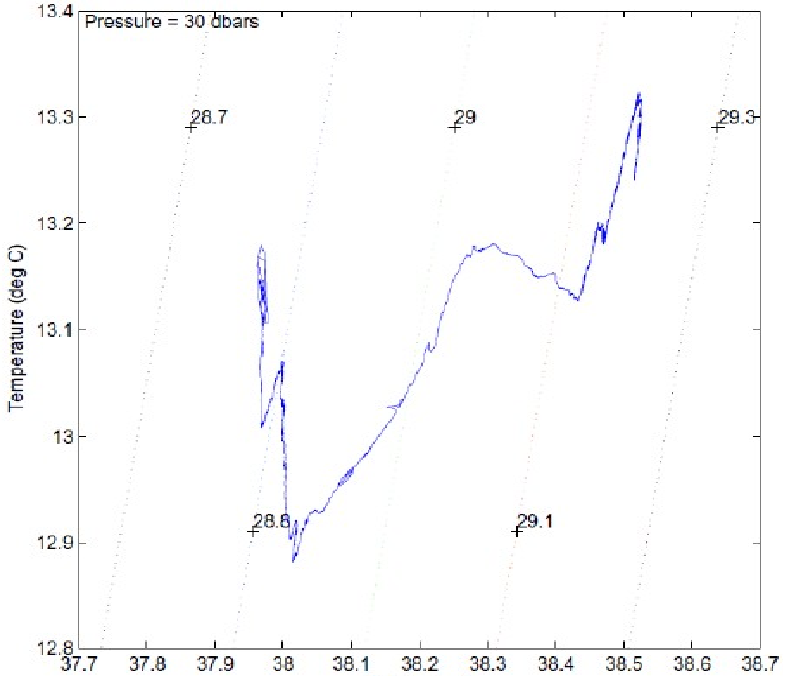
```
>> tsdiagram([0 40],[-2 30],0,0:10:30)
```

Once you have chosen adequate limits for T and S as well as the pressure at which you would like to create the graph, (and maybe the SIGMA isolines) you can add your T and S data points from the CTD profile to the same figure, for example:

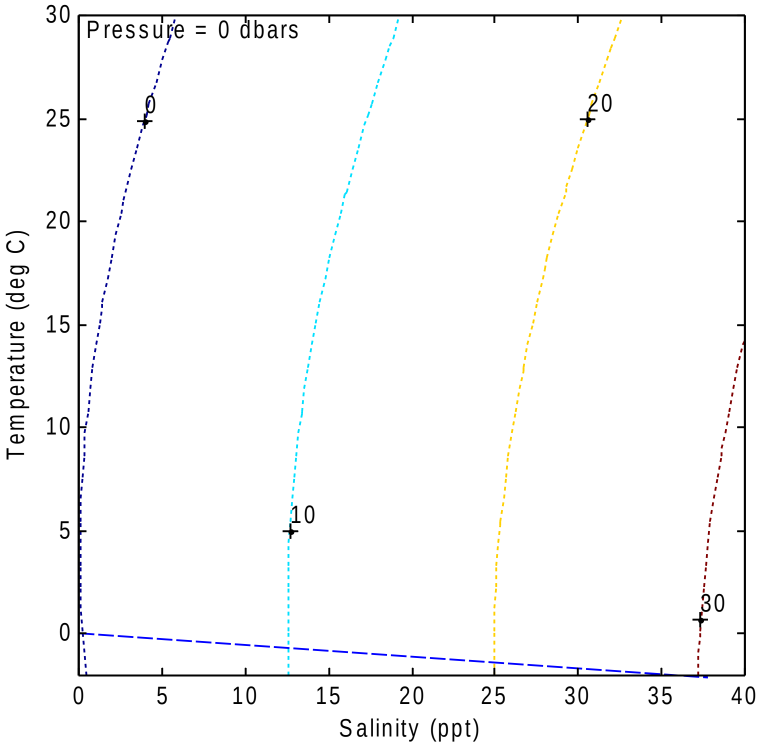
```
>> hold on
>> plot(data.CTD_statK_sal, data.CTD_statK_temp)
```

note: in Matlab, the help command yields the following:

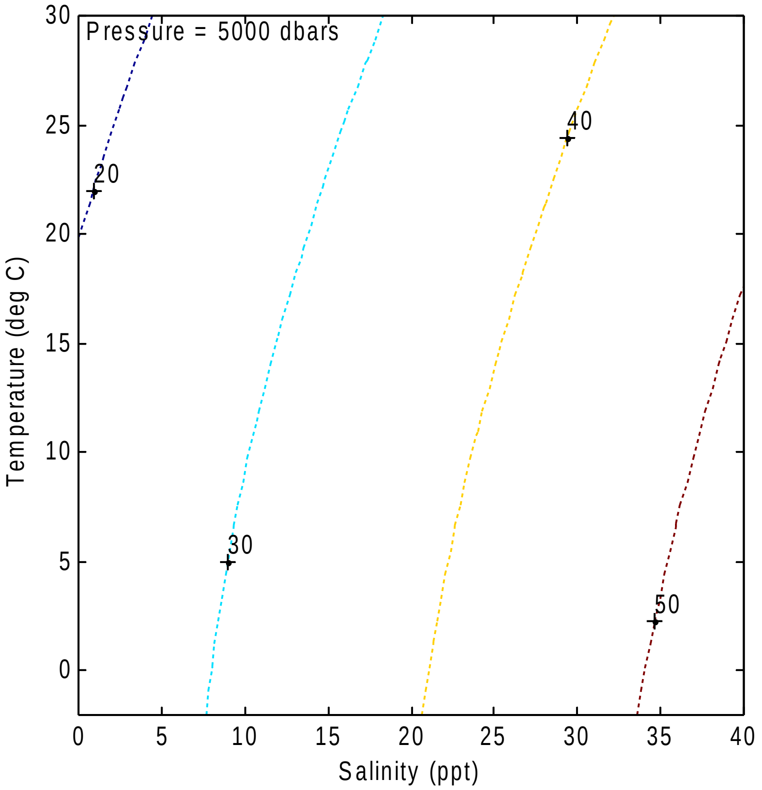
```
>>help TSDIAGRAM
TSDIAGRAM(S,T,P,SIGMA) draws contours lines of density anomaly (lines) SIGMA (kg/m^3) at pressure P (dbars), given a range of salinity and temperature (deg C) in the 2-element vectors S,T.
The freezing point (if visible) will be indicated.
```



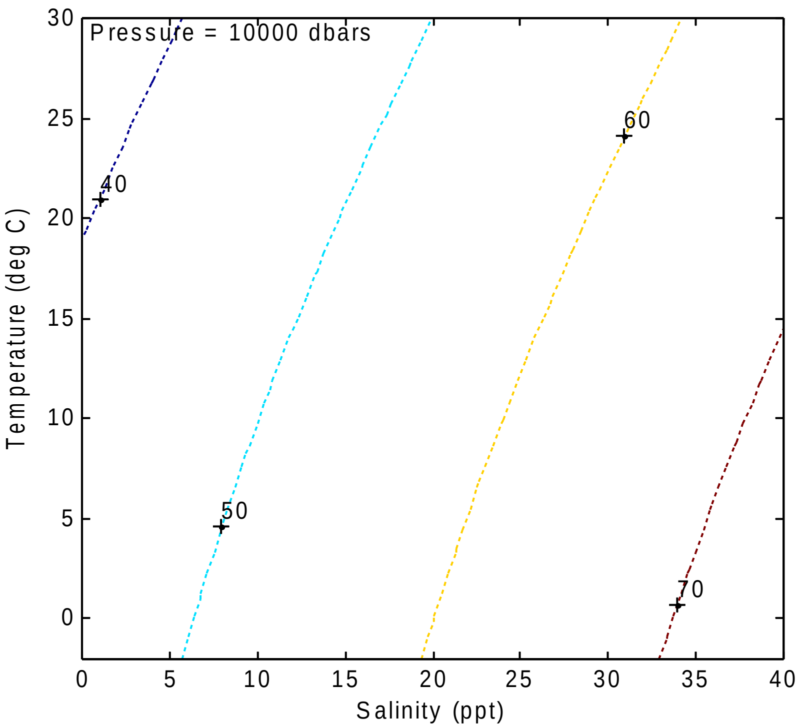
```
tsdiagram([37.7    38.7],[12.8   13.4],30)
hold on
plot(data.CTD_stat_sal, data.CTD_stat_temp)
```



tsdiagram([0 40],[-2 30],0,0:10:30) (includes a line indicating the freezing point)



tsdiagram([0 40],[-2 30],5000,20:10:50)



tsdiagram([0 40],[-2 30],10000,40:10:70)

- You can also make plots of Conservative Temperature and Absolute Salinity using the function gsw_SA_CT_plot.m (TEOS 10) described in detail here: http://www.teos-10.org/pubs/gsw/html/gsw_SA_CT_plot.html

Once you have downloaded the TEOS-10 toolbox (http://www.TEOS-10.org) its usage is similar to tsdiagram:

gsw_SA_CT_plot(SA,CT,p_ref,isopycs,title_string)

INPUT:
SA = Absolute Salinity [g/kg]
CT = Conservative Temperature [deg C]
p = sea pressure [dbar]
(i.e. absolute pressure - 10.1325 dbar)

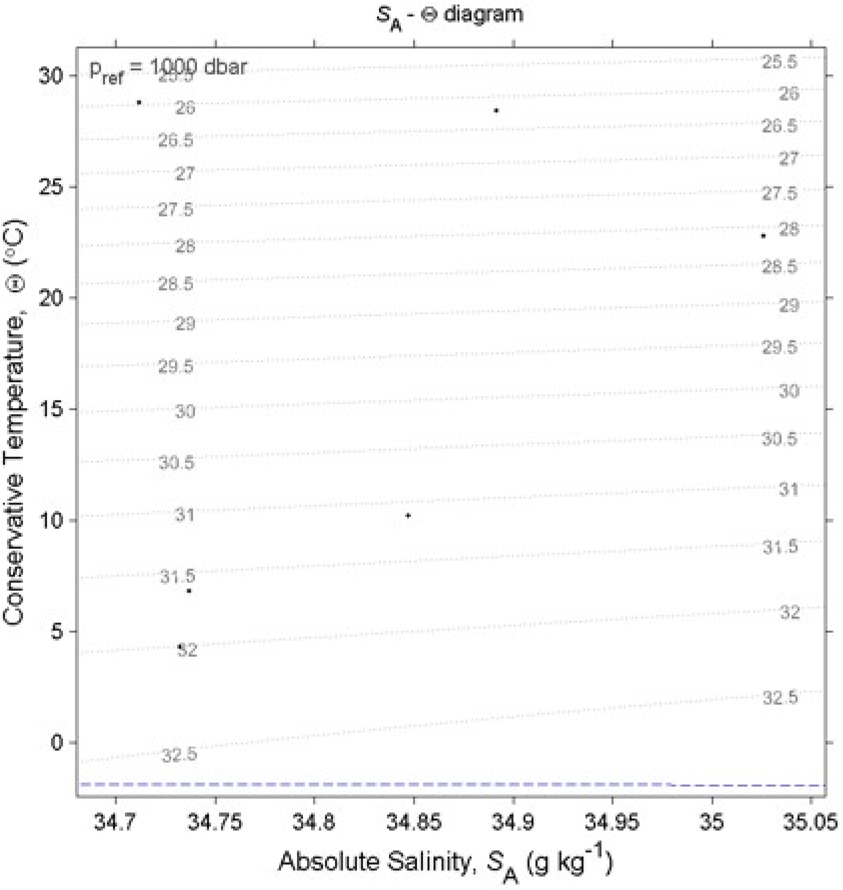
Optional:
p_ref = reference sea pressure for the isopycnals [dbar]
(i.e. absolute reference pressure - 10.1325 dbar)
If it is not supplied a default of 0 dbar is used.
isopycs = isopycnals, can be either an array of isopycnals or the number of isopycnals to appear on the plot. If it is not supplied the programme defaults to 5 isopycnals.
title_string = title text to appear at the top of the plot.

SA & CT need to have the same dimensions.
p_ref should be a scalar, (i.e. have dimensions 1x1).
isopycs can be either 1x1 or 1xN or Mx1

EXAMPLE:

```
SA = [34.7118; 34.8915; 35.0256; 34.8472; 34.7366; 34.7324;]
CT = [28.8099; 28.4392; 22.7862; 10.2262;  6.8272;  4.3236;]
p =  [     10;      50;     125;     250;     600;    1000;]
p_ref = 1000
isopycs = [24:0.5:33];

gsw_SA_CT_plot(SA,CT,p_ref,isopycs,'\it{S}\rm_A - {\Theta} diagram')
```



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