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HYDROSPHERE AND CRYOSPHERE

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THE HYDROSPHERE

In the most general sense, the hydrosphere refers to all the water in our ecosphere, including oceans, land water masses (glaciers, lakes, rivers, and underground), and moisture in soils, biota, and the air (i.e. all water molecules on, under, and over the surface of our planet), although it is often restricted to aggregated water in the liquid state, separately considering the cryosphere (frozen water), atmospheric water (humidity, dissolved and in suspension), and soil and biota moisture.

The hydrosphere interacts with the atmosphere and lithosphere mainly through the water cycle (controlling meteorology and hydrology, respectively). The pioneering oceanographer Matthew Maury said in 1855 "Our planet is invested with two great oceans; one visible, the other invisible; one underfoot, the other overhead; one entirely envelopes it, the other covers about two thirds of its surface." Water plays a dominant role in global and local energy balances through its radiative, phase change and thermal properties. The ocean/atmosphere interface is taken as part of the hydrosphere, including its motion and

properties (i.e. sea surface is here studied, whereas 'air surface' properties are usually measured at around 2 m over the surface). The main cause of motion in both, the atmosphere and the ocean is the non-uniform solar heating: a collimated energy beam, pointing to different places as the Earth rotates around its axis daily, and revolves around the Sun yearly.

The ocean is the set of interconnecting water basins (e.g. including the Black Sea, which is connected through the Mediterranean, but not the Caspian Sea, which is endorheic). Looking from the Moon, the Earth appears as a bluish planet mostly made of water, with small land patches (the continents), all under a changing pattern of clouds (covering in average half of the sphere). Thus, it is often said that Earth should more properly be called 'the water planet'. A fortunate energy balance (controlled by the greenhouse effect) keeps water on Earth not too cold to freeze and not too hot to boil-off and escape.

Oceanography is the branch of science dealing with the physical, chemical, geological, and biological features of the oceans and ocean basins. Oceanography deals with salt waters, whereas limnology (Gr. *limn*, lake) usually deal with fresh waters, in spite of the fact that some lakes are saltier than sea (e.g. Great Salt Lake, in Utah, has 100..300 g/L of salts, whereas seawater has 35 g/L).

Oceans were formed early in Earth history, by condensation of steam emanating from the hot interior, and the global amount of water is constant since geological times; it originated from outgassing during the first $1 \cdot 10^9$ years or so after Earth's formation ($4.5 \cdot 10^9$ years ago). It was a gas at the time, mixed with CO_2 and NH_3 , and no O_2 , condensing to liquid state as the Earth cooled down. Life begun around $3.5 \cdot 10^9$ years ago, and consumed most of the CO_2 and NH_3 to synthesize living matter (generating all the O_2 in the atmosphere), diminishing the green-house effect and condensing most of the H_2O . At its origin, seawater might have been more salty (some 100 g/L instead of the present 35 g/L) and hotter (some 40 °C instead of 17 °C, at the surface). In spite of life originating in the ocean, it is a hostile environment to humans: we cannot breath inside, seawater is not drinkable, ocean water is cold, waves and tides may cause damage to vessels and shores, it is a corrosive environment, and down in the ocean there is no light (all electromagnetic radiations get quickly absorbed) and pressure is squashing.

From the total area of the globe, $510 \cdot 10^{12} \text{ m}^2$, around 72% is in liquid form (ocean basically, but also lakes and rivers), some 3% is water in solid form (ice caps basically, but also mountain glaciers), and the rest 25% is land. An 86% of the total evaporation and 78% of total precipitation take place over the ocean; in global average, evaporation equals precipitation equals $500 \cdot 10^{15} \text{ kg/yr}$, equivalent to 900 mm per year of liquid precipitation over the globe (to be compared with the 3000 m equivalent ocean water, and the 30 mm equivalent liquid water in the atmosphere).

We make use of the ocean as a transportation medium (the cheapest for long routes, with low infrastructure needs, the highest capacity, but the slowest), as a source of raw materials (fish, seafood, salts, water, metals...), as a sink of wastes (banned nowadays: no more nuclear waste, no plastic waste, raw sewage cannot be dumped within 20 km of the coast, etc., as established in the marine pollution regulations of the International Maritime Organisation, MARPOL-IMO), and for other services like

cooling, leisure, military strategy (e.g. aircraft carriers and submarines), sea cultures (marine farms), ocean floor mining...

The main subject here addressed is Ocean Thermodynamics, a branch of Physical Oceanography with focus on energy exchanges (heat) and energy distribution (temperature) in the ocean, including a aspects of ocean motions relevant to thermal convection. It can be applied to global environmental modelling (global climate, water cycle, dispersion of contaminants, and so on), and to local ocean exploitation problems (e.g. thermal effects on surface and underwater navigation, on mineral and biota extraction, on platforms and cultures, etc.), ocean-energy applications being included in the latter.

Measurement platforms

Physical details of the ocean were scarce until the advent of artificial satellites in the 1960s, which have completely changed the situation. The ultimate goal is to measure all the physico-chemical variables at all three-dimensional spatial points, and time. Ocean measurements can be taken from different platforms:

- Shore. Seawater analysis, tide gauges...
- Ships (on regular or dedicated trips). Besides meteorological variables (air temperature and pressure, humidity, wind speed and direction, precipitations, solar irradiance), seawater physico-chemical properties (T , S , p , x_{O_2} , x_{CO_2} , pH , nutrients, radioactivity...), and mechanical magnitudes (sea bottom depth, wave height and period, currents) can be measured at different sites:
 - Deck probes for meteorological data and waves.
 - Towing probes (just below sea surface).
 - Profiling probes (descending), usually measuring S, T, p at a time by conductivity, temperature and depth (CTD-probe), sampling at 15 Hz while going down at 0.5 m/s.. Samples of deep seawater can be achieved by Nansen and Niskin bottles.
 - Bathymetry (see floor topography).
- Underwater crafts. Research submarines (autonomous manned vessels), submersibles (supported by a surface vessel, either manned like a bathyscaph, or robotic). Robotic submersibles can be remotely operated via an umbilical (ROV), or autonomous (UAV); since 1990, underwater gliders are the best choice for endurance and cost, and are particularly suited for ocean profiling, since its propulsion is based on deflecting its up-and-down buoyancy-driven motion with some wings. With a typical 1 W of battery power, a glider can follow a slow (<0.5 m/s) saw-tooth trajectory from its design depth (several hundred to a few thousand meters) to surface (where it connects by satellite for data download and telecommand), for many months of endurance and the whole ocean range.
- Buoys. Usually a fix vertical array of sensors. Usually on mooring buoys, but may be on drifting free-floating buoys.
- Airplanes. Unmanned air vehicles (UAV) are nowadays taking over manned flights.
- Satellites. The best for surface measurements. Global two-dimensional maps of sea-surface temperature (SST), wave height and propagation speed, sea surface salinity (SSS), and other sea surface composition details.

The thermodynamic model of the ocean usually restricts to the seawater state equation and its thermal capacity, both in terms of temperature, salinity, and pressure. The ocean is the major water reservoir on Earth, providing fresh water for all terrestrial life, and for industrial use, through the water cycle. Water salinity is the key control in ocean circulation, in a similar manner as water-vapour in the air being the key factor in atmospheric dynamics (although the latent-heat effect has no counterpart in the former).

The ocean is the major thermal reservoir on Earth, maintaining a moderately-varying temperature both spatially and temporally. The relative uniformity in ocean temperatures (all within 290 ± 20 K) is not due to its thermal conductivity but to the ocean currents that mix all around. Ocean water is subjected to gravitation and rotational forces, and driven by the sun-heating and winds originated from atmospheric circulation, and from gravitational forces from the Moon and Sun (tides).

Seawater properties

Water is the chemical compound H_2O , usually assumed to be in its liquid phase since specific names are given to its solid and gas state (ice and steam). The [physico-chemical properties of water](#) are really amazing. Living beings are basically aqueous solutions of organic macromolecules and inorganic salts and gases, and water is a key environmental substance; furthermore photosynthesis and respiration processes in aquatic life have a strong influence on O_2 and CO_2 concentration. Engineering also makes use of water as a working fluid and cooling media. On our planet, some 97.5% of all water molecules are found as seawater in the ocean.

Seawater is an aqueous solution of ions, and gases (CO_2 , O_2 and N_2 , mainly), with suspended particles of different sizes (seawater is first filtered before analysis). Some physical properties are presented in Table 1.

Table 1. Properties of seawater and pure water at 15 °C and 100 kPa.

Property		Seawater	Water
Salinity ^a	S [-]	0.035	0
Density	ρ [kg/m ³]	1026	999
Molar mass	M [kg/mol]	0.018	0.018
Thermal capacity	c [J/(kg·K)]	3990	4180
Thermal expansion	α [1/K]	0.00021	0.00015
Compressibility	κ [1/Pa]	$0.44 \cdot 10^{-9}$	$0.47 \cdot 10^{-9}$
Sound speed	c [m/s]	1510	1490
Thermal conductivity	k [W/(m·K)]	0.583	0.594
Thermal diffusivity	a [m ² /s]	$0.144 \cdot 10^{-6}$	$0.142 \cdot 10^{-6}$
Kinematic viscosity	ν [m ² /s]	$1.2 \cdot 10^{-6}$	$1.1 \cdot 10^{-6}$
Dynamic viscosity	μ [m ² /s]	0.0012	0.0011
Prandtl number (ν/a)	Pr [-]	8.4	8.1
Diffusivity of NaCl	D_i [m ² /s]	$0.12 \cdot 10^{-9}$	$0.12 \cdot 10^{-9}$
Diffusivity of O_2	D_i [m ² /s]	$2.5 \cdot 10^{-9}$	$2.5 \cdot 10^{-9}$
Diffusivity of N_2	D_i [m ² /s]	$3.6 \cdot 10^{-9}$	$3.6 \cdot 10^{-9}$
Diffusivity of CO_2	D_i [m ² /s]	$2.1 \cdot 10^{-9}$	$2.1 \cdot 10^{-9}$
Freezing temperature	T_f [°C]	-1.91	0.00
Fusion enthalpy	h_{sl} [J/kg]	$334 \cdot 10^3$	$334 \cdot 10^3$

Boiling temperature	T_b [°C]	100.56	99.6
Boiling enthalpy	h_{lv} [J/kg]	$2260 \cdot 10^3$	$2260 \cdot 10^3$
Vapour pressure	p^* [Pa]	1670	1705
Interface tension	σ [N/m]	0.078	0.073
Acidity ^b	pH [-]	8.0	6.5
Electrical conductivity	σ [S/m]	4.29	$4 \cdot 10^{-6}$

- a) Salinity, S , is the mass fraction of total ionic solids dissolved in water, in parts per thousand (ppt or ‰), e.g. $S=0.035=35\text{‰}$ (meaning there is 35 g of solutes per kilogramme of solution, what is approximately 35 g/L for aqueous solutions). A slightly modified definition of salinity (more accurately measured) uses a practical salinity scale with units of 'psu' (practical salinity unit) roughly equal to [‰], i.e. $S=35 \text{ psu} \leftrightarrow S=35\text{‰}$.
- b) Acidity of rain water is quoted; distilled water at 25 °C and 100 kPa has $pH=7.0$.

Density. Equation of state. Speed of sound

Water is not a constant-density fluid, fortunately, because density-variations with temperature and salinity are fundamental to natural convection and ocean circulation, and density-variations with pressure are crucial to underwater acoustics and ocean circulation. Water density also depends on the amount of solutes, $\rho=\rho(T,p,y_i)$, solid, liquid and gases, but the effect of dissolved gases and liquids is usually neglected, and only that of salinity retained ($S=y_i$). Average ocean surface density is 1026 kg/m³, but ocean bulk average is 1035 kg/m³ (mainly due to the effect of pressure). Density of common salt (anhydrous) is $\rho(\text{NaCl(s)})=2165 \text{ kg/m}^3$, and density of saturated brine at 15 °C is $\rho(\text{NaCl(aq),sat})=1195 \text{ kg/m}^3$, with $y_s(\text{NaCl(aq),sat})=0.264$.

Equation of state

In the range of interest, $30\text{‰}<S<40\text{‰}$ and $-2 \text{ °C}<T<32 \text{ °C}$ seawater density increases monotonically with S , p , and $1/T$ (there is no anomaly similar to the pure water anomaly at 3.98 °C), and a good approximation is $\rho=\rho_0+a(T-T_0)+b(p-p_0)+c(S-S_0)$ with $\rho_0=1026 \text{ kg/m}^3$ at $T_0=15 \text{ °C}$, $p_0=100 \text{ kPa}$, $S_0=35\text{‰}$, and $a=-0.21 \text{ (kg/m}^3\text{)/K}$, $b=0.45 \cdot 10^{-6} \text{ (kg/m}^3\text{)/Pa}$ and $c=770 \text{ kg/m}^3$ ($c=0.77 \text{ (kg/m}^3\text{)/(g/kg)}$); e.g. for abyssal waters $\rho(0 \text{ °C}, 100 \text{ MPa}, 35\text{‰})=1071 \text{ kg/m}^3$. A better approximation, but only valid for $T>5 \text{ °C}$, is got with the thermal expansion coefficient $\alpha\equiv(-1/\rho)\partial\rho/\partial T|_{p,S}=(77.5+8.7 \cdot T) \cdot 10^{-6} \text{ K}^{-1}$ (with T in °C), and the salinity expansion coefficient $b\equiv(-1/\rho)\partial\rho/\partial S|_{p,T}=(-779+1.66 \cdot T) \cdot 10^{-6}$ (with T in °C and S in ‰).

Because small density changes are so important, it is customary to work with a shift density scale, $\sigma(T,p,S)\equiv\rho(T,p,S)-\rho_0$, usually with $\rho_0=1000 \text{ kg/m}^3$ (e.g. $\rho=1026 \text{ kg/m}^3$ corresponds to $\sigma=26 \text{ kg/m}^3$).

Ocean acoustics is crucial for telecommunications and teledetection, since electromagnetic fields are drastically absorbed by water and not transmitted. An accurate speed of sound correlation is $c=1449.2+4.6T-0.055T^2+0.00029T^3+(1.34-0.01T)(S-35)+0.016z$, in m/s, with T in °C, salinity S in ‰, and depth z in m (Medwin-1975, in Lurton, X., 1998, "Acoustique sous-marine", Ifremer). Down to 500 m depth temperature effects prevail over pressure effects. An increase of 1 °C increases c in 3.5 m/s, an increase of 1‰ in salinity increases it in 1.4 m/s, and an increase in depth of 1 km increases it in 17 m/s. Water acoustics has some advantages over air acoustics: more than 4 times quicker (around 1500 m/s instead of some 340 m/s), and less attenuation, but it has no comparison with the EM radiation: speed

ratio 300 000 to 1.5 (km/s), one main steady source (the Sun) to help detect everything by passive reflexion, much wider spectrum, far less resolution (some kb/s at several kilometres; very poor real-time imaging)... Sonar stands for Sound Navigation and Ranging (radar for Radio Detection and Ranging).

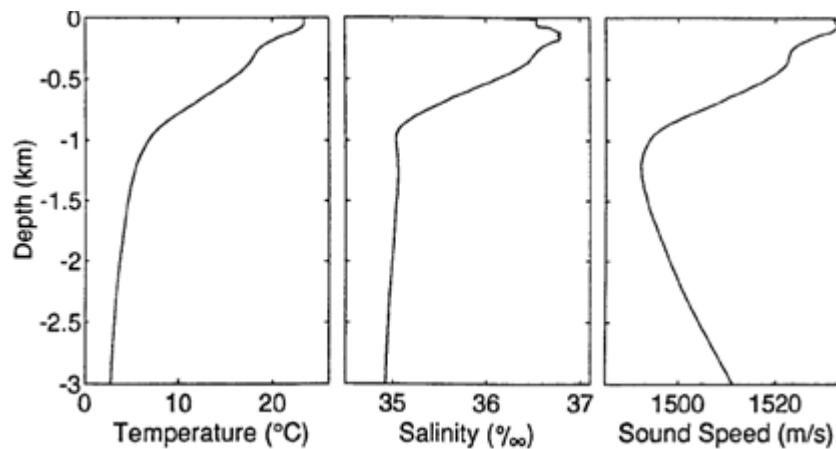


Fig. 1. Corresponding temperature, salinity and sound speed profiles from a CTD-probe in the Sargasso Sea. Sound speed increases toward the surface because of increasing temperature and toward the bottom because of increasing pressure.

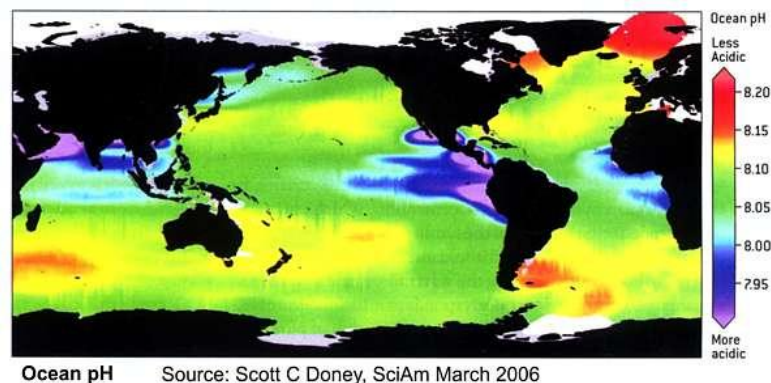
Vertical stability in the ocean: lapse rate and potential temperature

Vertical stability of seawater is studied in a similar way as for [vertical stability in the atmosphere](#). The adiabatic lapse rate in the ocean, $\Gamma_a = -\partial T / \partial z$, is obtained from the hydrostatic equation $dp/dz = -\rho g$, and the isentropic process, $ds = cdT/T - (\alpha/\rho)dp$, yielding $\Gamma_a = \alpha Tg/c_p$, ranging near the sea surface from $\Gamma_a = 0.017$ °C/km at -2 °C, to $\Gamma_a = 0.22$ °C/km at 25 °C, and $\Gamma_a = 0.12$ °C/km at 2 °C and 4 km depth (recall that in the atmosphere it was much larger and nearly constant $\Gamma_a = g/c_p = 9.8/1000 = 9.8$ °C/km). The actual lapse rate is always negative (buoyancy stable), except for polar regions (which are buoyancy unstable). Potential temperature in the ocean, θ , is defined in a similar manner than in the atmosphere, from $ds = 0 = c_p dT/T - (\alpha/\rho)dp$, but, in the ocean case it does not reduce to the simple $\theta_s = T(p_{ref}/p)^{(\gamma-1)/\gamma}$ with $p_{ref} = 100$ kPa and $\gamma = 1.4$, being an empirical correlation $\theta = \theta(T, p, y_s)$.

Dissolved gases

- O₂. Oxygen from the air gets dissolved in water up to an equilibrium amount of 10 g/m^3 at 15 °C, decreasing with temperature and salinity (it dissolves a little less in seawater than in fresh water), and increasing with gas partial pressure; oxygen solubility is $0.048 \text{ (kg/m}^3\text{)}/\text{bar}$ at 15 °C ($0.059 \text{ (kg/m}^3\text{)}/\text{bar}$ at 0 °C). Oxygen concentration is highest in surface waters, both from exchange with the atmosphere, and from photosynthesis ($\text{CO}_2 + \text{H}_2\text{O} \rightarrow \text{CH}_2\text{O} + \text{O}_2$), increasing with phytoplankton activity, and decreasing with temperature and with salinity. A typical vertical profile may have 5.7 g/m^3 of O₂ at the surface layer (the first 200 m or so well mixed), a rapid decrease to 1 g/m^3 or so near the thermocline (from 200 m to 600 m) due to biota respiration, and a gentle increase to some 5 g/m^3 at 4000 m or below due to oxygen-rich water sinking at high latitudes. Dissolved oxygen, in spite of its low concentration, allows aqueous living matter (flora and fauna) to breath. With an average of 1 g/m^3 , there is around $1.7 \cdot 10^{15} \text{ kg}$ of O₂ dissolved in the ocean, whereas in the atmosphere there is around $1.0 \cdot 10^{18} \text{ kg}$ of O₂ dissolved in the air.

- CO₂. Carbon dioxide from the air gets dissolved in water up to an equilibrium amount of 0.8 g/m³ at 15 °C, decreasing with temperature and salinity (it dissolves some 10% less in seawater than in fresh water), and increasing gas with partial pressure; CO₂ solubility is 2.0 (kg/m³)/bar at 15 °C (3.5 (kg/m³)/bar at 0 °C). Oceans hold about 60 times more CO₂ mass than the atmosphere (3.0·10¹⁵ kg of CO₂ dissolved in the air), and 20 times more than the land biosphere. Oceans can take up a lot of CO₂ because it doesn't all stay as a dissolved gas; it reacts to form a weak acid: CO₂(g)+H₂O(l)→H₂CO₃(aq). Cold seawater at the poles dissolves a lot of carbon dioxide from the air, which is carried slowly around the world by deep ocean currents. Carbon dioxide concentration in surface waters is low, being used for photosynthesis, increasing with depth due to biota respiration. Decay of organic matter, a reverse process of photosynthesis, adds CO₂ into deep waters. A typical vertical profile may have 90 g/m³ of CO₂ at the surface layer (the first 200 m or so well mixed), a rapid increase to 100 g/m³ or so in the thermocline (from 200 m to 600 m, and a gentle further increase to some 105 g/m³ at 4000 m or below. Notice that, in spite of the acidity caused by CO₂ ionization in solution, CO₂+H₂O=HCO₃⁺+OH⁻, seawater is mildly alkaline, with pH=8.1 due to H⁺ consumption by carbonate decomposition to bicarbonate, CaCO₃+H⁺+OH⁻=HCO₃⁻+OH⁻+Ca²⁺ in deep ocean. Rain water is mildly acid, pH=5.6, due to atmospheric CO₂ dilution.



Ocean pH Source: Scott C Doney, SciAm March 2006

Fig. 2. Ocean surface pH.

Dissolved solids

Dissolved solids are mostly ionic solids, i.e. salts in solution that have come from the chemical weathering of the crust and from volatiles released by volcanic gases. Salinity is the amount of ionic compounds in mass-fraction basis, S (usually in ‘parts per thousands’=ppt=‰=g/kg, i.e. grams of solute per kilogramme of seawater). It is difficult, however, to accurately measure total dissolved solids, TDS (mostly inorganic, but also organic), because solid residue depends on evaporation temperature, so that other measurement methods are preferred (see Salinity, below). For non-ionic solid solutes, symbol y_s is often used instead of S for solute mass fraction.

Average ocean salinity is around $S=35‰$, but average surface value is a little lower (34.7‰) showing ample geographical variations. The main ionic constituent of seawater is Cl⁻ brought by outgassing of Cl₂ from the sea floor (in drinking water it is HCO₃⁻ from Ca(HCO₃)₂ brought by rock weathering). Assuming all the $S=35‰$ (35 g/L) of salts being NaCl ($M=0,0233+0,0355=0.0558$ kg/mol), the NaCl concentration would be $c_s=(y_s/M_{NaCl})\rho=(35/55.8)\cdot 1026=640$ mol/m³=0.64 mol/L (a molar fraction of $x=0.64/(0.64+57)=0.0111$ of dissolved salt, which, known to fully ionise when dissolved, yields a

$x=0.0222$ of solutes, corresponding to a cryoscopic descent of $\Delta T=-RT_f^2x_s/h_{sl}=-1.9$ °C, an ebullioscopic elevation of $\Delta T=RT_b^2x_s/h_{lv}=0.57$ °C, and an osmotic pressure of $\Delta p=c_sRT=2.5$ MPa.

A water mass is said to be saline if it has more than 3 g/L of salt ($S>3\%$), because this salinity is: i) near the calcite branch point, ii) near the low points between modes when the frequency distribution of salinity of all lakes over 100 km² area is plotted logarithmically, iii) near the salinity at which most humans first begin to taste salt, and iv) the salinity below which biota typical of higher salinities are not found and above which the freshwater biota begins quickly to disappear or not extend. Average ocean salinity is 35‰, with the average solutal concentration presented in Table 1. When seawater evaporates (or is boiled), first the small carbonate content (CaCO₃) precipitates; then, when more than 80% of water has evaporated, sulfates (CaSO₄ and Na₂SO₄) precipitates (some 4 kg/m³); the mayor precipitate, some 24 kg/m³ of NaCl, starts when more than 90% of water has evaporated, later on, some 5 kg of MgCl₂ precipitates, and finally the remaining of salts (<2 kg out of the total 35 kg/m³ in seawater) precipitate as KCl, MgSO₄, CaCl₂ and other components (see Table 1). In the lab, artificial seawater may be prepared as a rough approximation, adding 35 g of NaCl to 1 litre of water, or as a more precise combination per litre of solution with 24 g of NaCl, plus 5 g of MgCl₂ (from MgCl₂·6H₂O), plus 4 g of Na₂SO₄, plus 1.1 g of CaCl₂ (from CaCl₂·2H₂O), plus 0.7 g of KCl, plus 0.1 g of KBr, plus 0.03 g of H₂BO₃. In contrast, the main ion in river water is HCO₃⁻, typically some 0.1 g/L; notice that its concentration in seawater may be nearly equal or even larger, but its contribution to salinity is negligible in the latter, and dominant in the former.

Table 2. Solute concentration in seawater ($S=35\%$).

Solute ion	by 1000 kg of seawater	by 1000 kg of seawater	Contribution to salinity (% in mass)	Content (by evaporation)	by 1000 kg of seawater
Cl ⁻	19.1 kg	546 mol	55%	H ₂ O	965 kg
Na ⁺	10.6 kg	469 mol	31%	NaCl	24 kg
SO ₄ ⁻²	2.6 kg	28 mol	7.7%	MgCl ₂	5 kg
Mg ²⁺	1.3 kg	53 mol	3.7%	Na ₂ SO ₄ and CaSO ₄	4 kg
Ca ²⁺	0.40 kg	10 mol	1.2%	CaCl ₂	1.1 kg
K ⁺	0.38 kg	10 mol	1.1%	KCl	0.7 kg
HCO ₃ ⁻	0.14 kg	2 mol	0.4%	NaHCO ₃ , H ₂ BO ₃ ...	<0.2 kg

Ocean salinity has been constant at about 35‰ for a very long time. Salinity stability is important as most cells require a rather constant salinity and do not generally tolerate values above 9%. Synthetic mixes of seawater salts are commonly used to make seawater for marine aquaria.

Scientific theories behind the origins of sea salt started with Sir Edmond Halley in 1715, who proposed that salt and other minerals were carried into the sea by rivers, having been leached out of the ground by rainfall runoff. However, although some salts (bicarbonates) are carried on to sea with the rain streams (rock weathering, including Na⁺), most ions come from the bottom crust by initial leaching or by excess volatiles from hydrothermal vents and volcanos in the floor, including Cl⁻ from chlorine, sulphur, magnesium salts, fluorine, etc. Na⁺ and Cl⁻ are so strongly enriched in seawater because they are not used by organisms and do not precipitate out very easily, except under highly evaporative conditions in salt

ponds or isolated basins. At its origin, some $3.5 \cdot 10^9$ years ago, seawater might have been more salty (up to 95 ‰) and hotter (up to 44 °C). Although seawater salinity varies with location (typically within $30\text{‰} < S < 40\text{‰}$), its composition is rather invariant (what allows a close correlation of salinity with chloride-ion concentration).

Salinity and electrical conductivity

Salinity is the quantification of the amount of dissolved salts in a given body of water. It may be given in mass fraction, molar fraction, mass concentration, molar concentration... Besides, raw salinity values are often corrected to be comparable under certain rules to an equivalent NaCl content. The broad classification of saline waters is: fresh water if total dissolved solids < 1 g/kg, brackish water if in the range 1..30 g/kg, saline water if in the range 30..50 g/kg, and brine if > 50 g of salt by 1 kg of solution.

There are several experimental methods to measure salinity:

- By residual weight after vaporization at 103 °C (100 kPa). When left to evaporate in air, first a CaCO_3 precipitate forms (0.1 g/kg), afterwards a CaSO_4 precipitate (3 g/kg), then a NaCl (24 g/kg), then MgCl_2 and KCl (6 g/kg), etc.).
- By densitometry, $\Delta\rho(c_s)$, with a floating weight, a pycnometer, or by Coriolis resonance.
- By refractometry, $n(c_s)$. There is a quasi-linear concentration dependence for the refractive index, from $n=1,333$ (pure water) to $n=1,35$ (seawater). Temperature must be finely regulated.
- By osmometry, $\Delta p = cRT$; e.g. for seawater $\Delta p = cRT = 2 \cdot 0,625 \cdot 1000 \cdot 8,3 \cdot 288 = 3$ MPa (it is 2.5 MPa in practice).
- By neutralization with AgI, one can determine y_{Cl^-} (Cl^- is only 55.0% of total dissolved solids, but the concentration does not vary). Approximately $S = 0.03 + 1,805 \cdot [\text{Cl}^-]$ (Knudsen-1902 formula). Notice that pH-metry does not work, since $\text{NaCl}(\text{aq})$ does not change $[\text{H}^+]$ or $[\text{OH}^-]$.
- By conductometry. Electrical conductivity, $\sigma = L/(RA) = \sigma(S, T, p)$, is almost linear for small salinities, with a minimum value (ultrapure water) of $\sigma = 4 \cdot 10^{-6}$ S/m (i.e. $\sigma = 0.04$ mS/cm, or a maximum resistivity of $\rho = 25$ M $\Omega \cdot \text{cm}$; $\text{S/m} = 1/(\Omega \cdot \text{m})$), and a slope of 0.2 S/m per kg/m^3 (). The present standard is the practical salinity scale (PSS) of 1978, which specifies salinity as the conductivity ratio $\sigma(\text{salt_water})/\sigma(\text{calibre_solution})$, where the calibrated solution has 32.4 g/kg of KCl (not NaCl) in water at 15 °C and 100 kPa, and labelled as psu (practical salinity unit), often not distinguished from the mass fraction in thousands (1 psu = 1 g/kg = 1‰). To make it more precise, all organic matter must be first fully oxidised, all bromine and iodine ions substituted by chlorine ions, and all carbonates converted to oxides.

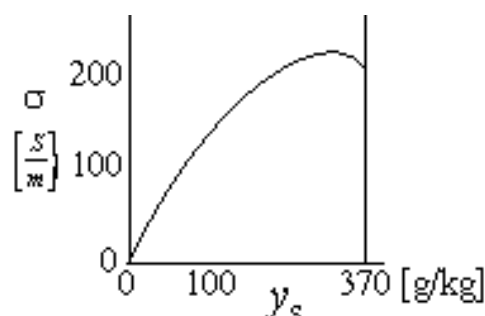


Fig. 3. Electrical conductivity vs. salinity.

Table 3. Salinity, S (mass fraction), and corresponding electrical conductivity, σ .

Type of salt water	S	σ	Notes
Saturated NaCl brine 26.4%wt at 15 °C	264 000 ppm	22 S/m	(adding 359 g to 1 litre) *
Typical brine 3:1	250 000 ppm	22 S/m	(adding 1 kg _{salt} to 3 litres)
Sea water	35 000 ppm	6 S/m	(60 000 μ S/cm)
Brackish water	3 000 ppm	0.6 S/m	
Drink water 100< S <1000 ppm (bicarbonates, not chlorides) (river water: 10..2000 ppm)	300 ppm	0.06 S/m	(600 μ S/cm at 25 °C +2% per excess degree)
Rain water (bicarbonates)	50 ppm	10 000 μ S/m	
Industrial water (treated)	30 ppm	6000 μ S/m	
Distilled water	3 ppm	600 μ S/m	CO ₂ (aq) from air affects σ
Distilled de-ionised water	<1 ppm	<200 μ S/m	
Ultra-pure water ISO 3696 type II (boilers)	0.5 ppm	100 μ S/m	<1 μ S/cm \leftrightarrow ρ >1 M Ω ·cm
Ultra-pure water ISO 3696 type I (analytical)	0.05 ppm	10 μ S/m	Dry residue at 110 °C <0.01 kg/m ³
Ultra-pure water (practical limit)	0.02 ppm	4 μ S/m	0.04 μ S/cm, 25 M Ω ·cm

*There are other aqueous solutions with higher electrical conductivity; e.g. water with 10% of H₄SO₂ has σ =50 S/m.

Sea surface salinity

Surface seawater salinity far from the coastline is determined by the balance between evaporation and precipitation, which in turn is controlled by solar heating. In general, seawater salinity is largest at subtropical regions because: 1) in the tropics and mid-latitudes, precipitation is greater than evaporation, and 2) in the Polar Regions, ice thawing is greater than evaporation. The saltiest ocean is the North Atlantic, with $S_{\text{surf,mean}}=37\text{‰}$ (against (35‰ for North Pacific); it is also warmer ($T_{\text{surf,mean}}=5.1$ °C against 3 °C).

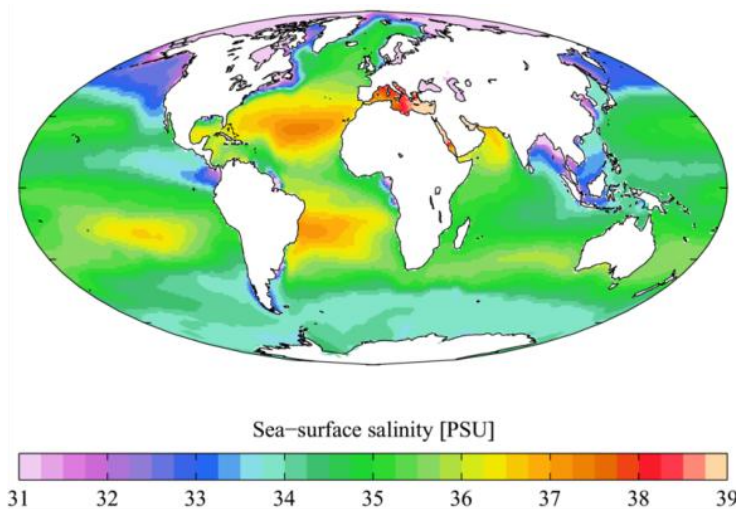


Fig. 4. Annual mean sea surface salinity, SSS (1 PSU \approx 1 g/kg).
(en.wikipedia.org; World Ocean Atlas 2005).

ESA's satellite SMOS (Soil Moisture and Ocean Salinity) in 2010 was the first to measure sea surface salinity. It was based on salinity effect on the dielectric constant of water, using the instrument, MIRAS (Microwave Imaging Radiometer with Aperture Synthesis), operating at L-band, 1413 MHz.

Enclosed seas have larger salinity excursions, according to river supply, evaporation, and precipitation regimes (Table 4). Differences in evaporation rate and salinity between the Mediterranean Sea and the Atlantic Ocean give rise to a crossed seawater exchange: superficial ocean water enters the Mediterranean to compensate evaporation, while denser (more saline) water flows out underneath (Fig. 5).

Table 4. Salinity, S (in ‰ or g/L) and other data of large water masses.

Water basin	S [‰]	Surface [km ²]	Surf. altitude [m]	Volume [km ³]	Max. depth [m]
Dead Sea ^a	340 ^b	800	-425	150	370
Great Salt Lake	50..300 ^c	4 400	1280	19	15
Persian Gulf	55	250 000	sea level	12 000	90
Red Sea	40	440 000	sea level	230 000	2 200
Mediterranean Sea	38	2 500 000	sea level	3 700 000	5 300
Ocean (mean)	35	360 000 000	sea level	$1.3 \cdot 10^9$	10 900
Black Sea	20 ^d	440 000	sea level	550 000	2 200
Baltic Sea	13 ^e	380 000	sea level	20 000	460
Caspian Sea	12 ^f	370 000	-28	75 000	1 000
Lake Tanganyika	5	33 000	770	19 000	1 500
Baikal Lake	0.1	32 000	450	24 000	1 600
Lake Superior	0.06	82 000	180	12 000	400

a) $\rho=1120 \text{ kg/m}^3$. b) 30%NaCl+50%MgCl+16%CaCl₂, c) Depends a lot on lake level. d) 18‰ above, 22‰ below the pycnocline. e) 10‰ above, 15‰ below the pycnocline. f) $S=12 \text{ g/L}$, mainly constituted by carbonates and sulfates.

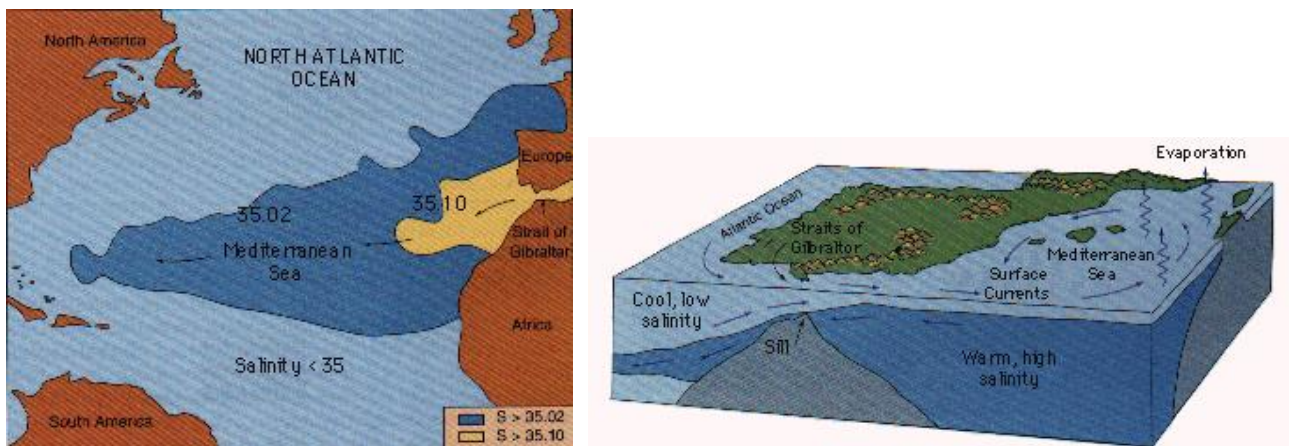


Fig. 5. Seawater exchange at the Gibraltar Strait.

Salinity profiles

Salinity levels are very uniform in deep ocean, with much wider spatial and temporal oscillations at the sea surface layer (some 100 m of well stirred water). A transition between these surface and bottom zones may show a salinity gradient, called halocline region, although not as marked as the thermocline (see Fig. 1 and Fig. 99). A combined plot of salinity, temperature, and density, is shown in Fig. 6, where the vertical profile (depth dependence) for the three variables can be traced from the warmer upper layer (in each of the major ocean basins), towards the abyssal waters represented by the Antarctic ocean bottom.

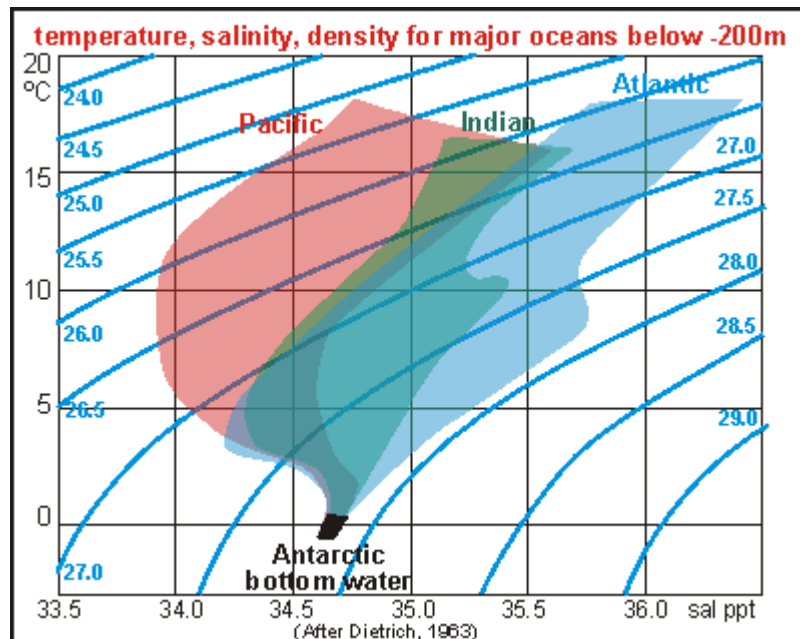


Fig. 6. Salinity, temperature, and density, vary from the ocean surface layer at each basin (top of the figure) to the ocean bottom (bottom centre in the figure).
<http://www.seafriends.org.nz/oceano/seawater.htm>

Effects of drinking seawater

Seawater is salty and bad for drinking, but an occasional small gulp does not harm, especially if a larger quantity of fresh water is drunk afterwards. However, consuming seawater to maintain hydration is counterproductive, and problems already appear by the litre because, in the long run, more water must be expended to eliminate the seawater's salt (through excretion in urine) than the amount of water that is gained from drinking the seawater itself.

This occurs because the amount of sodium chloride in human blood (around 4 g/L, or 4‰ by weight) is actively regulated within a very narrow range by the kidney, which can concentrate NaCl(aq) up to a maximum of 9 g/L. Drinking seawater (with 35‰ of dissolved sodium chloride) temporarily increases the concentration of sodium chloride in the blood, promoting sodium excretion by the kidney, but the sodium concentration of seawater is above the maximum concentrating ability of the human kidney. Eventually with further seawater intake the blood concentration of sodium will rise to toxic levels, removing water from all cells and interfering with nerve conduction ultimately giving seizures and heart arrhythmias which become fatal. Some animals can adapt to salty living conditions, like the desert rat, which is able to concentrate sodium far more efficiently than the human kidney, and therefore would be able to survive by drinking seawater. Sea animals do not dehydrate by osmosis because they segregate an oily mucus (that makes them slippery) to increase insulation.

Sea ice salinity depends on rate of formation and age. It grows with freezing rate from $S=5‰$ in slow processes (e.g. a sudden drop to $-15\text{ }^{\circ}\text{C}$ in air temperature) to $S=10‰$ in fast freezing (e.g. a sudden drop to $-40\text{ }^{\circ}\text{C}$ in air temperature), and slowly decreases with time due to leaching of the denser brine droplets (e.g. by summer, the top layer in the Arctic banquise is suitable to making drinking water). Salt is trapped as brine droplets, decreasing with time because of seepage (one-year-old sea-ice is good for drinking).

Distilled water is also bad for drinking; it is not bad for short periods (although it is tasteless), but for prolonged periods it dissolves out essential minerals from our body.

Freezing of seawater

An aqueous solution with only one salt species shows a simple eutectic point, like for NaCl-water. Seawater shows a different behaviour, with no eutectic point, but a monotonous salinity concentration increasing as temperature drops.

An accurate freezing temperature correlation is $T_f(p,S) = -0,0575S + 1.7105S^{3/2} - 2.155S^2 - 75.3 \cdot 10^{-9}(p-p_0)$, in °C, with salinity S in ‰, and p in Pa.

Table 5. Eutectic point in single solution and the start of precipitate in seawater

Salt	Eutectic point in single solution	Start of precipitate in seawater
CaCO ₃ ·6H ₂ O	0 °C, 0‰ (nearly insoluble)	-2.1 °C, 0‰
Na ₂ SO ₄ ·10H ₂ O	-3.6 °C	-8.2 °C, 150‰
MgCl ₂ ·8H ₂ O	-34 °C	-18 °C
NaCl·2H ₂ O	-21 °C, 230‰	-23 °C, 300‰
KCl	-11 °C	-37 °C

Sea ice (banquise) has a residual salinity of 6..10‰ the first year, decreasing to 0..4‰ after several years. Sea ice retains some brine (in tiny pockets), decreasing with temperature from the 10..40% at the bottom liquid-solid interface (the hottest part, at -2 °C), to 1% at -5 °C, 0.3% at -20 °C, to 0.1% at -30 °C, etc. For this reason, there is a large increase in ice tensile strength below -5 °C (ice-breakers have experienced it).

Desalination

Desalination means removing salt from saline water to render it drinkable (or at least suitable for irrigation). Saline water may come from the sea or from salty ponds on ground (brackish water). If the goal is to get the salt and discard the water, it is called salt works, or salt pans (the latter also referring to natural geological formations like Salar de Uyuni in Bolivia). The Earth's water cycle provides the basic natural desalination process in the world, evaporating seawater, forming clouds, blowing them over the continents, precipitation and run-off.

We restrict here to the production of fresh water from seawater (35‰ by weight of salts) by splitting an incoming flow into two streams, one with low salinity (typically $S < 1‰$, or < 500 ppm TDS for human consumption according to the World Health Organization, WHO), and another obviously more saline than the original. The ratio of mass flow rates between the fresh water obtained and the incoming one is an important parameter known as yield mass ratio (or desalination factor), typically less than 50%.

Desalination is expensive because it demands a large energy input. As for 2012, the best desalination plants consume a little over 6 kJ/kg of electricity (i.e. 6 kW per 1 kg/s of fresh water produced, or nearly 2 kWh/m³, which, at 5 c€/kWh for big consumers (at 25 kV) means 0.1 €/m³ just on electricity. The thermodynamic minimum is half of that: 3 kJ/kg (0.8 kWh/m³ ≈ 0.04 €/m³), but in real plants, to the 0.10

€/m³ of electricity must be added 0.15 €/m³ for depreciation, 0.10 €/m³ for maintenance, and 0.05 €/m³ for water pre- and post-processing, what adds to a minimum cost of 0.10+0.15+0.10+0.05=0.40 €/m³ (the best plants produce at 0.50 €/m³ and older ones at 1 €/m³). Drinking water at 0.50 €/m³ is very cheap just for drinking (we are satisfied with 3 kg/day), and not too expensive for other domestic uses like washing or toilet flush (most dwellings in large cities pay more than 1 €/m³, although the cost is not due to the supply of water but mainly due to sewage treatment); the real problem of desalination business is that more than 70% of water use is for irrigation, and present-day irrigation farming is not competitive with water price above ≈0.03 €/m³.

Desalination methods

- Thermal methods. By segregation in a phase-change, basically liquid-vapour, but more rarely liquid-solid transition. The vaporisation may be in multi-stage flash chambers (MSF), by multi-effect distillation (MED), by vapour compression, by solar collector condensation, etc.
- Membrane methods. By segregation through a semipermeable membrane, mainly by reverse osmosis (RO), but sometimes by electrodialysis.
- Chemical deionisation methods. By selective ion exchange, by ion precipitation, or by capacitive electric deionisation.

Desalination needs first appeared in besieged cities, boats, islands and dry coastal regions. Until late 20th century the solution was based on distillation. Aristotle described a method to evaporate impure water and then condense it to obtain potable water; he wrote “Salt water when it turns into vapour becomes sweet, and the vapour does not form salt water when it condenses again. This is known by experiment.” Alexander of Aphrodisias in AD 200, wrote that sailors at sea boiled seawater and suspended large sponges from the mouth of a brass vessel to absorb what is evaporated and then drawn potable water. In the Renaissance, Della Porta mentioned in his book *Magiae Naturalis* (1558) several methods of desalination, describing a solar distillation apparatus, and a method to obtain fresh water from the air by dehumidification. In 1872, an engineer from Sweden, Carlos Wilson, designed and built the first large solar distillation plant, in Las Salinas, Chile, to provide fresh water to workers in a saltpetre mine (KNO₃) and a nearby silver mine; it consisted of 64 bays having a total surface area of 4450 m², producing 23 m³ of fresh water per day.

Nowadays, most desalination installations, from manual water-makers producing <10 kg/h (costing >1000 €) to large plants of >3000 kg/s, are reverse osmosis systems (sea water reverse osmosis, SWRO, and brackish water reverse osmosis, BWRO), with pressure recovery to increases efficiency (the first in 1975). The first large-scale municipal SWRO plant was installed in 1980 in Jeddah, Saudi Arabia. It consumed 8 kWh/m³, and demanded expensive maintenance. Energy recuperation is most important in SWRO because they must operate at high pressure, usually 7 MPa (BWRO systems have low operating pressures and low rejection rates, with less energy available for recovery). The first work exchanger on SWRO was installed in 1975 in Bermuda.

As freshwater is not available in their environment, seabirds like pelicans, petrels, albatrosses, gulls and terns possess a gland which separates salty water by counter-current exchange: the seawater inflow at the mouth gets more concentrated, by extraction of fresh water at the gland, and the brine is secreted near the nostrils above the beak, and 'sneezed' from time to time. Mangroves trees grow in seawater, and concentrate seawater at their roots, secreting the brine, which are then eaten by animals (usually crabs). Additional salt removal is done by storing it in leaves which then fall off.

Recycled urban water

At present, recycled municipal water from the sewage system can be used for (there are not piping systems in all cases):

- Irrigation of private and public gardens.
- Street cleaning.
- Ornamental fountains and lakes.
- Agriculture irrigation of food (except if in contact with food be consumed fresh).
- Car washing installations.
- Cooling towers.
- Fire fighting systems.
- Aquiculture (in some cases).
- Aquifer replenishment.
- Toilette flushing.

Ocean boundaries: surface and bottom topography

Everybody knows that the ocean upper surface, lower surface, and lateral coastline are irregular, although the upper surface is locally quasi planar in the average (and the whole ocean surface quasi spherical). The 'quasi' is due to non-equilibrium effects driven by the Sun and Moon (and other smaller perturbations), and the effect of Earth's rotation. The coastline only have very small changes; in the short term with tide phase (although it may be remarkable at some special sites, like Mount Saint Michel in France), and more secular at river deltas and estuaries (a delta is the alluvial area at the mouth of some rivers that carry alluviums into non-tidal seas, with mainstream splitting up into several distributaries, whereas an estuary is the widening channel of a river where it nears a tidal sea, with a mixing of fresh water and salt (tidal) water (living beings there must cope with large salinity variations within hours).

Before precise satellite radar-altimeter data were available, only rough absolute sizes (and detailed relative measurement) could be performed. Now, with satellites moving around the gravitational centre of the Earth, with minor external disturbances (which can be discounted), we can compute, with a typical resolution of 2 cm:

- Wave height, its distribution in time and space, and propagation speed.
- Reference sphere (fitting time-and-space averages to a sphere). $R=6371$ km.
- Reference ellipsoid (fitting to an ellipsoid of revolution around Earth axis). Earth's equatorial radius $a=6\ 378\ 137$ m, polar radius $b=6\ 356\ 752$ m, difference $21\ 385$ m, flattering,

$f=1-a/b=1/300$, eccentricity $e=1-(a/b)^2=(2-f)f=1/150$, gradually decreasing over geologic time scales. Moon is more spherical, $f<1/825$, and Jupiter more oblate, $f=1/15$.

- Reference geoid (i.e. reference equipotential surface, first defined by Gauss). From precise orbital mechanics, the series of geometrical surfaces with the same gravity potential can be computed; $V=(GM/r)F(r, \theta, \phi, t)$ (in a rotating frame, sum of the attractive potential and the centrifugal potential; local gravity is its gradient; at sea-level g varies from 9.781 m/s^2 to 9.832 m/s^2 from Equator to Poles). The reference geoid is the equipotential surface that best fits local tide-gauges (and its virtual extension across continents as with very narrow virtual canals), and the mean Earth volume from satellite altimetry, i.e. once the dynamic ocean motion is subtracted (in a rotating Earth). The geoid is a smooth surface horizontal to a local plumb (weight-line), considerably smoother than Earth's physical surface; e.g. while the latter has excursions of $+9000 \text{ m}$ (Mount Everest) to $-11\,000 \text{ m}$ (Mariana Trench), the total variation in the geoid differs less than 200 m from the reference ellipsoid (-106 in America and India, to $+85 \text{ m}$ in Europe and New Zealand). Ocean circulation and bottom topography can be inferred from satellite altimetry (e.g. a 2000 m floor mountain yields a 2 m bulge in the geoid).
- Nowadays, the reference ellipsoid is computed from geoid data (with same volume), which is the reference for standard-GPS coordinate system (latitude, longitude, and elevation). The reference for the precise-GPS coordinate system is the geoid. Both are geocentric coordinate systems co-rotating with the Earth (Z along the rotation North Pole, X along the Greenwich observatory's meridian plane), and Y as for a right-hand trihedral.

Sea surface and mean sea level

Sea surface altitude (height over the standard geoid) can be measured from satellite radar altimetry with an accuracy of 3 cm (as for 2010), and maps of wave height, tidal height, and ocean currents can be obtained from that.

Mean sea level (MSL) would not be a spherical surface even with a non-rotating, windless, isolated Earth, because of non-spherical internal mass distribution. Before satellites, only local mean-sea-levels LMSL (i.e. relative to a local land mark) could be measured (e.g. tide gauges in San Fernando and Alicante in Spain, in Cornwall and Liverpool in UK...). Monthly or yearly averaging gets rid of wave and tide fluctuations, but not of quasi-steady ocean currents, climate effects (mean temperature and pressure), and own land motion, so that actual MSL data (i.e. static ocean topography) may differ up to $\pm 2 \text{ m}$ from the current geoid model and even more from the ellipsoid model (i.e. GPS altitude at a LMSL line may show a value of $\pm 50 \text{ m}$ not because of accuracy but due to modelling). Extending MSL far from the measuring point adds more uncertainty. Nowadays, MSL may refer to the reference ellipsoid (for GPS altitudes) or to the reference geoid.

Global mean sea level is rising at a rate of 3 mm/yr in the 2010s (it was only 1 mm/yr in the 1950s), as measured with tide gauges in geological stable zones (since 1880), and from satellites (since 1993), although with marked regional differences (from -10 mm/yr near the Labrador peninsula, to 10 mm/yr near the Philippines), which appear to be transient. Mean sea level rise is due to global warming, mainly

due to thermal expansion of the ocean, but glacier thawing contribution is becoming significant. Mean sea level has risen around 0.15 m in the 20th century, but the estimation for the 21st century is 2 m. According to paleoclimate studies, such a level increase had not happened since the first tiers of Würm glaciation (115 000 BCE..10 000 BCE), where a 20 m rise occurred (from –20 m in 85 000 BCE to 0 m in 84 000 BCE).

Because of differences in seawater density due to salinity and temperature, there can be an averaged level jump at the two sides of an isthmus; e.g. in the Suez Canal, the Red Sea is 24 cm higher than the Mediterranean; in the Panama Canal, the Pacific is 18 cm higher than the Atlantic. For the same reason, sea level in Spain coast is some 82 cm higher in Cantabrian Sea than in the Mediterranean.

Tides are the rise and fall of sea levels caused by gravitational forces exerted by the Moon and the Sun, and the rotation of the Earth. Tides usually refer to ocean tides, but similar gravitational phenomena take place in the gaseous atmosphere (atmospheric tides), and in the solid Earth (Earth tides), because they are deformable continuous media too. They have a short 12 h period over a longer 29 d period. According to Strabo (1.1.9), Seleucus was the first to link tides to the lunar attraction, and that the height of the tides depends on the moon's position relative to the Sun. The first explanatory model was in 1687 by Newton (Galileo was wrong, attributing tide motion to Earth's translation sloshing). Small semidiurnal variations in water-surface can also be observed in large lakes (e.g. about 3 cm in Lake Michigan, 190×490 km² in size).

Ocean circulation (ocean currents) is inferred from dynamic topography of sea surface, i.e. the differences between local mean sea level and the geoid.

Sea bottom

The sea floor is a nearly uniform basin of 4 km depth (the abyssal plain), with some deeper zones of subduction (trenches, where lithospheric plates meet, the deepest is Marianas, 11 km below sea level), and spreading mountain ranges (ridges; usually in mid ocean, the largest is Mid-Atlantic Ridge, up to 2 km high and 20 km wide). The Atlantic Ocean is expanding because ridge accretion gains trench subduction, whereas the Pacific and Mediterranean are shrinking, all of the order of 2 m per century.

Near continents, the sea floor rises rather steeply in the so called continental slope, and then joins the shore in a more or less wide continental shelf.

Sea bottom altitude (height over the standard geoid) is measured by sonar bathymetry, but only 10% of the seafloor has been charted this way so far. It can be estimated from sea-surface satellite altimetry, based on the change in gravity due to mass differences (e.g. a deep ocean trench may cause a valley in the seas surface 10 m deep and 200 km wide).

Temperature at the sea floor is around 2 °C, but a geothermal gradient of some 25 °C/km soon develops below (e.g. some 150 °C 6 km underneath).

Ocean thermal behaviour

The Earth Radiation Budget (ERB)

Life on Earth is maintained by radiation from the Sun, which drives all other processes too, as atmospheric and oceanic motions (tides being the exception).

Solar radiation is mostly absorbed at the sea surface (84% by the ocean, 7% by the ice cover, 5% by land, and 4% by the atmosphere). Since the atmosphere is nearly transparent to solar radiation, the 'air ocean' is unstably heated from below, whereas the water ocean is stably heated from above, in Earth's gravitational field.

Both, the physics of the atmosphere and the physics of the oceans, are governed by thermodynamics and fluid-dynamics, the main difference being the key role of moisture in the air and of salinity in the ocean. The difference in thermal and mechanical inertia also yields different characteristic relaxation time and space scales: a few days and say 1 km for air in the atmosphere, and about 1 month and 30 km for ocean water.

The factors controlling the energy budget are

- Sun-Earth geometry. Daily, monthly and annual cycles.
 - Earth-Sun distance.
 - Earth shape: spherical; the sun does not shine equally on all parts.
 - Earth spinning tilt: the Earth axis is tilted relative to the plane of its orbit; the amount of sunlight reaching the poles varies with the seasons. Most sunlight reaches the tropics.
- Radiation absorption.
 - Solar radiation absorption mostly takes place at the ocean surface. Absorption is computed as the difference between incident and reflected radiation, since transmission is negligible down the few tens of metres.
 - Cloud shield. Clouds control the amount of sunlight reaching Earth's surface.
 - Greenhouse effect. Cloud and gas absorption of terrestrial radiation keep the atmosphere relative warm, rising the equilibrium surface temperature by some $\Delta T_{\text{greenhouse}}=33$ K (to $T_{\text{mean}}=288$ K instead of $T_{\text{mean}}=255$ K).
 - Aerosols in the air scatter sunlight.
- Fluid convection (ocean currents and winds contribute half-and-half to energy convection to the Poles).
 - Radiation absorption creates temperature gradients that set up convection currents in the atmosphere: winds.
 - Winds carry heat, mostly as water vapour, toward the poles.
 - The global ocean circulation has about 1000 yr period.
 - The ocean surface is usually about 1.5 °C warmer than the overlying atmosphere. Since the ocean and atmosphere are in contact at the air-ocean interface, heat energy passes from the warmer ocean to the colder atmosphere. About 10% of ocean cooling is through this process, plus another 50% carried out by water vapour; the 40% rest is by IR radiation.

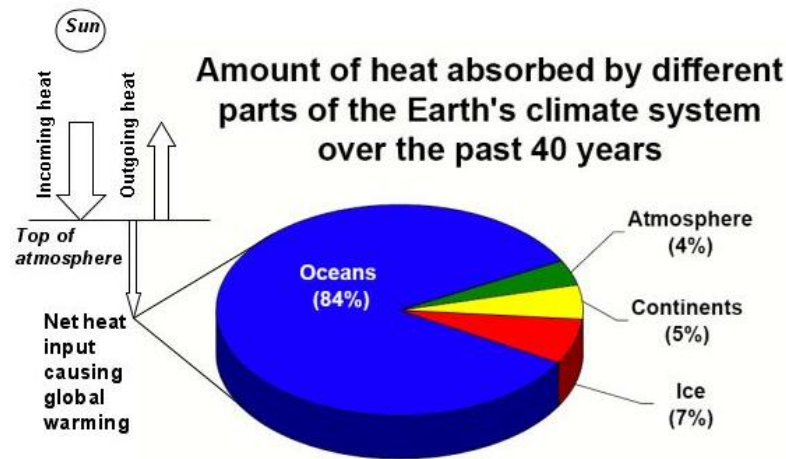


Fig. 7. Solar absorption (from Levitus et al., Geophysical Research Letters, 2004).

Sea surface albedo

Albedo is the fraction of the Sun's radiation reflected by an opaque object back into space. Earth's albedo is $\rho=30\%$ in the average, with large variations: highest values ($\rho=70..90\%$) correspond to ice and snow covered surfaces, while lowest values correspond to water surfaces ($\rho=2..10\%$, although water albedo varies with the angle of incidence and approaches 100% when the Sun is very low on the horizon; but the angle dependence is not linear, and $\rho<10\%$ after the Sun is more than 25° above the horizon).

Chlorophyll concentration has an influence on solar absorption, transmission and heating of the oceans. Marine phytoplankton controls the air-sea exchanges of carbon dioxide, sulfur compounds, and ocean radiation transfer as well as albedo of ocean surface water, which affect Earth climate. A higher abundance of chlorophyll increases absorption of solar irradiance and heating rate in the upper ocean

Energy transport in the water cycle

The 1360 W/m^2 of normal solar irradiance outside Earth's atmosphere averages to 342 W/m^2 around the planet, from which 30% is reflected back (albedo), 50% is absorbed at the surface (mostly at sea surface, 45%, the 5% rest at the ice and land surfaces), and the 20% rest is absorbed by air (16%) and clouds (4%) in the atmosphere. From the $342 \cdot 0.45 = 154 \text{ W/m}^2$ absorbed by the ocean in the average, 71 W/m^2 are transferred to the atmosphere as latent heat of vaporization. The rising water vapour and its condensation into clouds are entrained by the winds and give rise to precipitations and run-off that close the water cycle.

Water is evaporated mostly from the oceans. Much less is evaporated from lakes, rivers, soils, and vegetation on land. The evaporated water eventually condenses and falls as rain or snow over both land and ocean. Approximately 93% of the evaporated water comes from the oceans. Only 71% of the rain or snow (global precipitation) falls on the oceans. In other words, more water evaporates from the ocean than re-enters it as rain or snow. This is good for us as that is where most of the freshwater supply for Earth comes from. Only 3% of the total water on Planet Earth is fresh water.

Air moisture follows the winds. Around the Earth there are several major bands where high or low pressure dominates. At the equator, heating from the Sun causes the air to rise leaving an area of low

pressure. This air sinks back down to earth at $\sim 30^\circ$ N and $\sim 30^\circ$ S resulting in bands of high pressure. At around 60° N and 60° S, cold dense air moving away from the poles meets the warmer air from nearer the equator. The warm air is less dense so is forced to rise, leading to areas of low pressure. This air cools and sinks forming a high pressure zone around the poles.

An ocean thermal equator may be defined as the latitudes dividing North

The Oceanographic Thermal Equator is a direct reflection of the ocean's ability to balance temperature. It is located between 5 and 10 degrees North latitude, near, but not coincident with, the thermal equator for the following reasons. More water exists in the Southern Hemisphere (81 percent is covered by water) than in the Northern Hemisphere (61 percent is covered by water). The high specific heat of water allows the Southern Hemisphere to absorb a great deal of heat without a corresponding large temperature increase. The large ice covered continent of Antarctica acts as a cooling mechanism for the entire Southern Hemisphere.

Surface maps: sea surface temperature

Temperature, as any other local variable in the ocean, varies horizontally, vertically, and with time. Overviews are usually split in these three components, separately considering time-average horizontal map values, time-average vertically profile values, and space-average time series values. Vertical zoning of the ocean is dealt with below. Horizontally, the ocean can be split in several different zones according to land proximity (directly related to other parameters like ocean life):

- Littoral zone: the coastal environment (up to some 100 m offshore and down to some 10 m depth). Sunlit penetrates to the floor, which is rich in seafood (e.g. mollusc).
- Neritic zone (Gr. *Nereus*, a sea God): the continental-platform sea (up to 200 km offshore and down to 200 m). Sunlit scarcely penetrates to the bottom (<1% visibility below 100 m). Most fish live in this region, which occupies some 8% of ocean surface.
- Pelagic zone (Gr. *πελαγος*, far sea): the deep sea region beyond the continental platform (92% of sea surface), which is vertically divided in the pelagic zone properly (the first 200 m down the surface), the bathial or demersal zone (from 200 m to 4000 m), and the and abyssal zone (below 4000 m depth). Bottom waters are generally called the benthic zone (Gr. *βενθος*, deep sea), not only in deep ocean but on the neritic regions and large lakes.

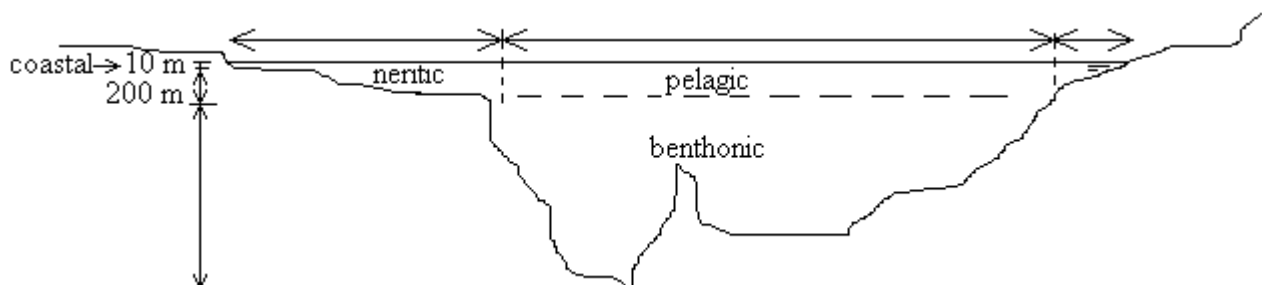


Fig. 8. Sketch of ocean zones.

Temperature in the ocean varies less than temperature in the air, but, as the interest is on energy and not only in temperature, it is important to be accurate and give specify details. Three measurements zones on the ocean surface can be distinguished:

- Skin temperature (or Sea Surface Temperature, SST). It is the water surface temperature measured by IR emission from the air side, notably by satellite radiometers.
- Ship trailing temperature, measured at some 0.05..0.1 m below sea surface.
- Buoy-depth temperature, measured at about 0.5 m below sea surface by buoys moored or drifting in the ocean.
- Bulk surface temperature, measured at about 3 m below sea surface from ships.

Bulk surface temperature is traditionally the favourite measurement of oceanographers because they want a measure of the heat content of the ocean. At night, when the first 5 m of the ocean are well mixed, all these temperatures will be within 0.2° C of each other. However, during the day, with strong solar insolation and a stably stratified upper ocean warm layer in the top 1 m, there can be as much as a 2..3 °C difference between the skin and ship numbers. This difference can be very significant when calculating ocean heat fluxes, budgets and contents.

The high latitude oceans have a much larger variability in sea surface temperatures than tropical oceans, taking an active part in the global thermohaline circulation with the sinking of cold, salty waters.

Sea surface temperature should not be confused with air surface temperature. Global average temperature of the ocean surface waters is about 17 °C (varying from –2 °C in polar waters to 35 °C in the Red Sea and Persian Gulf), whereas global average temperature of the air at sea level (directly measured over the ocean and corrected from ground measurements) is about 15 °C (varying from –90 °C in Antarctica to 60 °C in Libya desert).

Global mean surface temperatures have risen over the past 100 years by about 0.6 °C, with half of the increase in the last 25 years. Locations in Alaska and northern Eurasia have warmed by nearly 6 °C in the winter months over the past 30 years, and some cooling has occurred in the North Atlantic and central North Pacific.

Vertical profiles: thermocline

Ocean depth can be split in several different layers according to temperature, composition, sunlit penetration, or currents (pressure increases almost linearly, without significant features). Most classifications are based on a three-layer structure (e.g. according to light: the well-lit euphotic zone, the shadowy disphotic zone, and a lightless aphotic zone).

According to temperature, the ocean can be conceptually divided in three layers:

1. A thin (a few hundred metres) surface layer more or less well-mixed because of high convective currents. Temperature in this surface layer oscillates seasonally with solar irradiance, with amplitude decaying to nearly zero at its bottom.
2. An intermediate gradient layer, more or less thin, that matches the other two. The '-cline' suffix indicates a slope in the magnitude that characterises this layer, thus the thermocline is the layer where temperature quickly drops from the warm surface layer to the cold bottom layer, and similarly the pycnocline in the density profile and the halocline in the salinity profile.

3. A thick bottom layer (from 1 km to the sea bed; i.e. typically several kilometres depth).

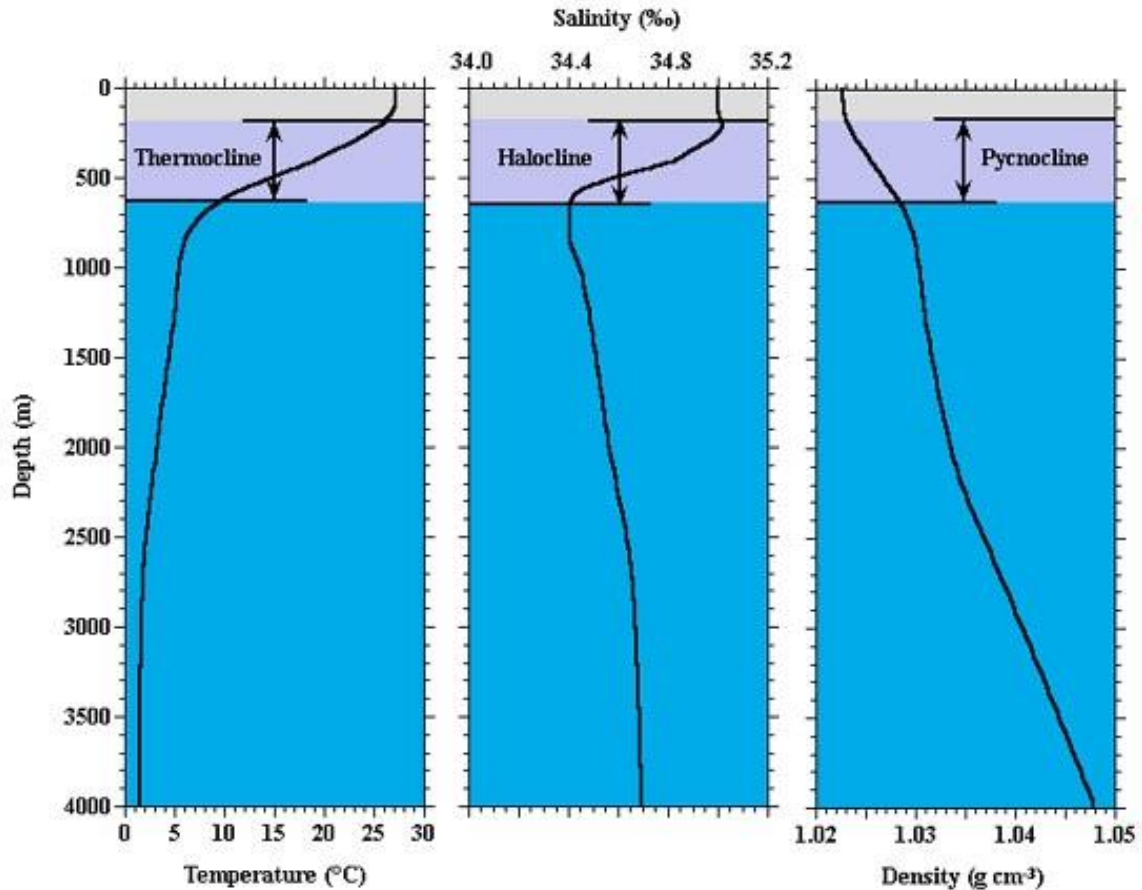


Fig. 9. Typical profiles of temperature, salinity and density (Pacific ocean) shown the mixed surface layer, the gradient zone (thermocline, halocline, and pycnocline), and the main bulk zone underneath.

Low-frequency sounds transmitted across the ocean basin at a particular depth are being used for tracking long-term changes in ocean temperature (acoustic thermometry).

Underwater gliders are most appropriate to gather ocean profiles, because of their inherent up-and down trajectory.

Ocean motions

Unlike the atmosphere, all oceans are laterally bounded by continents except in the Southern Ocean where the ocean extends all the way around the globe.

Thermohaline circulation

Water motion in the ocean can be split in 3 types: waves, tides, and currents:

- Surface waves are caused by wind stresses, and consist of oscillations, both in the vertical direction and the wave-propagation direction, rarely beyond 10 m in amplitude, with no average speed.
- Tides are caused by moon and, to a lesser extent, by sun attraction (first explained by Newton in 1687; tide correlation to moon phases was known from Antiquity). The highest vertical excursion is 19.6 m Fundy Bay (C). In Spain 5.3 m in Santander and 1 m in Alicante.

- Currents are caused by horizontal thermal gradients (equatorial/polar), by vertical and horizontal saline gradients and by entrainment of waves and tides, all of it deflected by the Coriolis force (turning clockwise in the North and anti-clockwise in the S). Surface currents (<100 m deep) are caused by permanent winds (e.g. from Canary Islands to Caribbean Sea by the trade winds). The Gulf Stream, some width and with around 1 m/s (and 27.5 °C near Florida), carries 100 times more water than all the rivers in the world. Ocean currents make the West coast of continents at mid-latitudes ($\pm 40^\circ$) warmer than their East coast, but cooler at subtropical latitudes (at $\pm 20^\circ$).

Ocean dynamics have characteristic times larger than atmospheric dynamics, but much shorter than lithosphere dynamics (plate tectonics).

THE CRYOSPHERE

The cryosphere (Gr. *κρυωο*, cold) refers to water in solid form, including:

- The two polar ice caps, formed by polar glaciers (thick ice sheets over bedrock) and by thin sea ice sheets floating on the sea (banquise). The two main polar ice sheets (Antarctica and Greenland), are also named inlandsis.
- Other glaciers, including circumpolar glaciers (in all continents except Australia), and mountain glaciers (e.g. Himalayas, Andes). The largest glacier after Antarctica and Greenland is in Patagonia. Mountain glaciers are quickly retreating due to global warming: Alpine glaciers have lost a third of their mass in the last 100 years, and Kilimanjaro's ice cover shrunk from 20 km² in 1880 to 2 km² in 2003. All glaciers (including the inlandsis) are formed by accumulated annual snowfall, and they cover about 10 percent of the world's total land area.
- Icebergs, floating glacial ice detached from polar glaciers.
- Seasonal ice: lake ice, river ice, snow cover, and banquise ice that thaws in summer. About 98% of the annual snow cover is located in the Northern Hemisphere (Siberia, Canada, Alaska...)
- Permafrost, i.e. permanent (two consecutive years) frozen ground (sometimes including a seasonal active layer that melts on summer, if it exists). The lower limit of permafrost is dictated by the geothermal gradient (25 °C/km in average near the surface). Seasonally frozen ground covers the largest area of the cryosphere (50 % of the land area in the northern hemisphere is frozen in January). Soil becomes waterlogged and swamps and thaw lakes are abundant because the iced ground below is impermeable. Freezing of water in the active layer causes upward expansion of ground, called frost heave, forming mounds along the edges of the permafrost zone.
- Although not usually included, high clouds are made of ice crystals, and clathrates are frozen water at $T > 0$ °C stabilised by gas inclusions at high pressure in the sea floor. As already said, the [physico-chemical properties of water](#) are really amazing.

The banquise is frozen seawater, a relatively thin (<10 m) continuous or patched layer, floating over liquid seawater, which forms mainly in the Arctic Ocean (more than 3000 km in diameter, but rarely more than 3 m thick), but also around Antarctica (only a few hundred kilometres width, in the average), with some of their mass thawing and freezing seasonally. At the present rate of sea ice retreat, the Arctic Ocean may become ice free by 2050. Sea-ice thickness is measured nowadays with a few centimetres accuracy from satellite radar altimetry and density of floating ice.

A glacier is identified as a creeping flow of ice bounded by surrounding relief at both banks and the bottom. The motion is forced by dislocations within the lattice of ice crystals caused by weight, with speeds of the order of 100 m/year. Polar glaciers sit in the Arctic (Greenland, Canada, and Siberian Russia), and at Antarctica on the South Pole. These large ice sheets stand mainly over bedrock, but they may bridge over large trapped water areas (otherwise, ice pressure would push water away). The largest glacier is in Antarctica, being 200 km long, 20 km wide, and 2 km deep. Glacier-ice thickness is measured locally by ice-penetrating radar; different radar frequencies are used to image internal layers at different depths and deeper layers can be better imaged using lower frequencies.

The vertical temperature profile in most thick ice sheets is similar, with temperatures growing downwards with a typical geothermal gradient of 25 °C/km at the bottom (Fig. 10), what yields a geothermal flux of $k\nabla T = 2.4 \cdot 0.025 = 0.06 \text{ W/m}^2$. Density of ice can be modelled by $\rho = \rho_0(1 - c_s \exp(-\gamma_s z))$, where $\rho_0 = 917 \text{ kg/m}^3$ is the compact ice value, c_s is the porosity of surface ice (% by weight of air) usually taken as $c_s = 0.69$, and γ_s a densification exponent factor (the inverse of the characteristic compaction depth, $z_c = 1/\gamma_s = 1/0.02 = 50 \text{ m}$); i.e. surface ice has $\rho = 300 \text{ kg/m}^3$, and reaches $\rho = 850 \text{ kg/m}^3$ some 100 m below surface (corresponding to some 3500 years old ice).

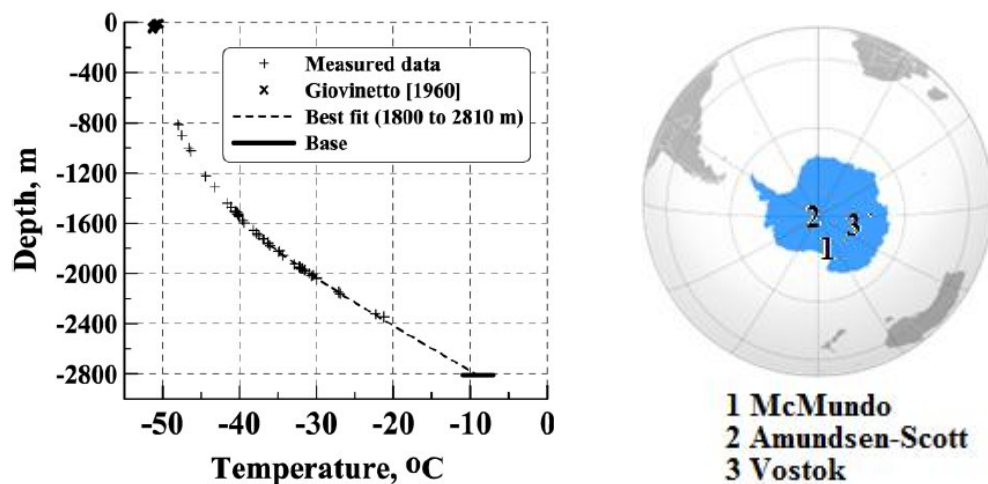


Fig. 10. Temperature profile in Antarctica's Amanda boreholes near the South Pole (2002), and location of some Antarctic bases.

Relevance of cryosphere studies:

- Influence on Earth's radiation budget, by the high albedo of fresh snow (90%) and ice (80%) in comparison with water (5%) or soil (20%..30%), and by changing ice-coverage extension. Sea ice insulates the relatively warm ocean water from the cold polar atmosphere, and it is a very sensitive indicator of climate changes.
- Influence on general thermohaline circulation, by supply of cold and fresh water on melting. The cryosphere exchanges heat, moisture and salinity with its environment.
- Influence on mean sea level. Present estimation is that from 2000 to 2100 sea level will rise about 2 m as a result of global climate warming, with Greenland's contribution alone estimated in 0.5 m; some 180 million people around the world in low lying lands might be displaced.

- The best site for geological-times dating, because the trapped water and gases retain the characteristics they had in ancient times. Ice cores several kilometres long and some 8 cm in diameter have been extracted, yielding climate data for the past hundreds of thousands of years.
 - Deuterium concentration in trapped H₂O molecules can be correlated with Earth's global mean temperature at the corresponding stratigraphic epoch, since the warmer the ocean, the lesser evaporation of this heavier hydrogen isotope.
- The largest reservoir of fresh water.

The two largest ice masses are in Antarctica and Greenland. Ice coverage is best determined by passive microwave detection from satellites, because frequent cloud cover in the polar regions and the fact that the sun remains below the horizon for continuous periods in winter, make visible sensing unreliable. The extent of the cryosphere has changed a lot over geological times, with periods of no ice at all, and others, like recently in the Pleistocene (the last two million years) where the Arctic ice sheet covered the North of Europe and North America (the Fennoscandian and Laurentian ice sheets, respectively); there seems to be a 100 000 year glaciation cycle with 90 000 years of cold and 10 000 years of warm weather; the last ice age finished some ten thousand years ago, and perhaps the Little Ice Age (lasting from 1500 to 1800) was the start of a new ice age which has been overturned by the Industrial Revolution. Since 2009, the Northwest Passage is open to marine shipping most of the year. Anthropogenic pollution of the cryosphere is to be watched out (Antarctic tourism is already about 50 000 pax/yr).

Antarctic ice

- Ice volume is estimated to be almost $30 \cdot 10^{15} \text{ m}^3$ (some $25 \cdot 10^{18} \text{ kg}$), i.e. about 90% of the world's ice mass (about 70% of the world's fresh water mass). Estimated ice loss is about $100 \cdot 10^{12} \text{ kg/yr}$.
- Ice covered area is $13.7 \cdot 10^6 \text{ km}^2$, which is 98% of Antarctica's emerged area.
- Ice thickness average may be 1.6 km, being up to 4 km at some places. The highest peak is at 4892 m.
- Precipitation: mean 100 mm/yr (20 mm/yr at Vostok, 200 mm/yr at the coast).
- Next to the continental shore are many ice shelves. These are floating extensions of outflowing glaciers from the continental ice mass. They gain mass primarily through flow from grounded ice sheets and glaciers; they lose mass through iceberg calving and melting. Most of the world's ice shelves, including the largest, are in Antarctica.
- Offshore, temperatures are so low that there is a layer of sea ice floating on seawater, with its spatial extent changing a lot seasonally; in Antarctic's Southern Ocean, from some $18 \cdot 10^6 \text{ km}^2$ in late winter to $4 \cdot 10^6 \text{ km}^2$ remaining at summer's end (in the Arctic, the oscillation is from $15 \cdot 10^6 \text{ km}^2$ to $8 \cdot 10^6 \text{ km}^2$).

Vostok sub-glacial lake

This is the largest of more than 140 sub-glacial lakes, found in 1993 beneath Vostok Station which is 3480 m above sea level (as for 2012, a drilling borehole is approaching the lake surface). The lake surface is in the average 500 m below sea level, although it is not horizontal (it is not a free surface), with an average depth of 340 m and a maximum one of 510 m; it has $16 \cdot 10^3 \text{ km}^2$ of surface area, and 5400 km^3 of fresh water. It seems that the lake freezes every 13 000 years, swept away as typical glacier-bottom ice,

and formed again. Water temperature is estimated to be $-2.7\text{ }^{\circ}\text{C}$ roughly the solid-liquid equilibrium of water at 32 MPa (notice that the minimum Earth's surface ever measured is $-89\text{ }^{\circ}\text{C}$ precisely at Vostok Station, where the year average is $-55\text{ }^{\circ}\text{C}$). Geothermal flux at the lake bottom is estimated to be 0.054 W/m^2 .

Greenland ice

- Ice volume is estimated to be almost $3 \cdot 10^{15}\text{ m}^3$ (some $2.5 \cdot 10^{18}\text{ kg}$), only a tenth of that in Antarctica, but in comparison nearly half of the all the air mass in the atmosphere. It is presently thawing at a higher pace than ever, some $50 \cdot 10^{12}\text{ kg/yr}$, but notice that ice would still last for millennia at this melting rate. If the whole ice sheet melted away completely, the world's sea level would rise near 7 m (Antarctic ice would add 70 m more); just with this sea-level rise, countries like Bangladesh, and big cities like London, New York, Shanghai, and Los Angeles would flood.
- Ice covered area is $1.8 \cdot 10^6\text{ km}^2$, which is 85% of Antarctica's emerged area.
- Ice thickness: average 1.6 km, up to 2 km. Average Greenland's surface height is 2.3 km. The weight of this ice sheet has depressed the central land area to form a basin lying more than 300 m below sea level.
- It can be noticed that other lands at same latitude, like in Canada and Siberia, do not have ice covers as extensive and thick; the reason may be that underneath the ice on Greenland there are some high mountains that might have provided a permanent ice cover that cooled all around.

Analysis of ice cores and marine sediments show that dozens of cyclic natural warmings occurred in the last 2 to 3 million years, without totally melting polar ice. During the 13 000 years' warm period before the most recent ice age, the Eemian interglacial period 120 000 years ago, roughly half of the Greenland ice sheet melted, contributing some 3 metres to the 5 metres sea level rise observed during that period. It seems that Greenland was ice-free 3 million years ago, when concentration of CO_2 in the air was around 400 ppm (it is 395 ppm in 2012).

Ice sheet formation

There are two basic types of ice sheet formation (we are not considering ice formation on cold solid surfaces, neither evaporative cooling by a pressure drop).

- Water ice. It forms on the water surface, or at the bottom surface of an ice sheet standing over water. It has a grain structure that is uniform through the ice thickness. It contains relatively little dissolved air.
- Snow ice. It forms on any surface by falling snow. It has a grain structure comprised of the ice granules from the original slush. It contains a lot of air trapped between the ice gains.

If the formation of ice over a water layer at $T_w \geq 0\text{ }^{\circ}\text{C}$ by contact with ambient air at $T_a < 0\text{ }^{\circ}\text{C}$ is considered, the following process takes place. First, after some undercooling, small ice crystals form; then grow and get noticeable. If they can join quietly, they form transparent ice sheets; if the growth is disturbed (agitated or running water), it forms lumps of loose ice crystals (called frazil ice), a kind of foam ice that

may get almost dry and white as snow (e.g. snow that seems to fall only on the river), or wet and dark like ice slurry.

River freezing

Frazil ice stays at the surface of natural streams, and accumulates on the borders and on cold solids (may break wood footbridges). When frazil ice from the borders grow and meet in the centre of the stream, sheet ice forms (its thickness is very irregular, contrary to sheet ice over quiet water). Water continues running under the ice sheet. As the water begins to rise during the spring melt upstream, the ice layer gives way, cracking into blocks that float down, what may jam the river and cause flooding.

Glacier ice accumulation

Forgetting about beautiful glacier caves, glacier ice is a layered accumulation of snow ice, dull and whitish, with most characteristic parameters increasing with depth: pressure, density, temperature, age...

Fallen snow, with $\rho=(50..70)$ kg/m³, is being slowly compacted by its own weight and sublimation, which removes the air and increases the density. After a few days, snow settles by snow-crystals breaking their loose structure and collapsing, forming a wet foam with $\rho=(200..300)$ kg/m³. After a few years' time, and perhaps following some thawing and recrystallization process, ice may have $\rho=(400..800)$ kg/m³ and may be 10 m to 50 m below surface. This granular porous ice is called firn (old ice), and can be sometimes found at the surface (with $\rho=(300..400)$ kg/m³), where it has the appearance of wet sugar, and it is very hard to shovelling.

After decades and centuries of residence time, glacier ice properly is formed, with $\rho=(600..900)$ kg/m³ and a non-porous solid structure is formed by grain sintering, bluish in colour and containing little air in tiny bubbles (which can be analysed for isotopic content for paleoclimatic studies, together with stratigraphic models). Typical downward motion of ice parcels is 0.1 m/yr.

At the bottom of a glacier, usually bedrock but in some parts liquid water (seawater in between Greenland rock islands, or fresh water in Antarctica's Vostok lake), some melting may take place by pressure, friction, and geothermal heating, with regelation or water sliding. In any case, besides this downward ice movement, there is also a lateral glacier downhill creeping, with speed increasing towards the lower end (the terminus) where it may be from 50 m/yr to 5 km/yr (i.e. in the range 0.2..20 m/day). With global warming, thaw water from glacier top pours through fissures and gets to the bottom, acting like a lubricant to allow the ice to race ever-faster toward the sea. In addition, when snow melts at high altitudes and then refreezes, it can absorb up to four times more sunlight, creating even more melting the next year. On the other hand, humidity ratio and precipitations increase with warmer temperatures.

The marked stability of polar climate is due to a layer of very cold air just above ground level, air that is always heavier than the upper layers of the troposphere. The result of this is a strong, gravity-driven air flow down the slopes of ice sheets (i.e. catabatic winds), generating strong gusts of up to 100 m/s at ground level.

Summer thaw depth

The summer heating may cause some superficial thawing in not-too-cold iced surfaces. The water formed (slush really) may freeze in place in autumn, or may slide down on inclined surfaces (with or without re-freezing downstream). [In July 2012, almost all Greenland ice cover thaw](#), superficially, in a matter of weeks, a rare event that had never happened in the last century (normally less than half of the ice-covered area in Greenland thaws every summer). No figures were available for the depth of thawing, and we wanted to make a guess of how fast this active layer may grow (the active layer concept is much used in permafrost studies, as the soil thickness that thaws and re-freezes annually).

To find the growth rate of the thickness thawed, $\delta_w(t)$, a one-dimensional energy balance may be established; as for times, we only care for day-averaged values, to have a steady forcing (the only unsteady term is the thaw depth). Taking the whole liquid layer as the system, we equate its energy increase (latent plus sensible, although the latter is negligible), to the heat gain through the upper surface by solar radiation, and heat losses by convection and radiation (we neglect heat conduction at the bottom surface, and expansion effects).

$$\rho_s h_{sl} \frac{d\delta_w(t)}{dt} + \rho_w c_w \delta_w(t) \frac{dT_w(t)}{dt} = \alpha_w \gamma_{atm} \gamma_{day} E \cos \beta - h_a (T_f - T_a) - \varepsilon_w \sigma (T_f^4 - T_{sky}^4)$$

$$\rightarrow \frac{d\delta_w(t)}{dt} = \frac{E' - h_a (T_f - T_a) - \varepsilon_w \sigma (T_f^4 - T_{sky}^4)}{\rho_s h_{sl}}$$

The solar heating, starting from the solar constant $E=130 \text{ Wm}^2$ (really, the extra-terrestrial solar irradiance in the Northern summer should be taken as $E=1320 \text{ W/m}^2$ because the Sun is 1.7% far away), is reduced by the inclination of the Sun rays relative to the local vertical, $\beta=70-23.5=46.5^\circ$ (for a mean Greenland latitude of 70° , and solar declination of 23.5° at summer), $\cos\beta=0.7$; it is also reduced by the atmospheric filter by the amount $\gamma_{atm}=0.6$ (assuming clear sky); it is reduced by day averaging in the approximate amount $\gamma_{day}=0.4$ (there is sunlit 24 h/day, but $\beta=46.5^\circ$ at noon and $\beta=86.5^\circ$ at midnight); and finally we have to account for the solar absorptance of water, that we take as $\alpha_w=0.9$; in total, solar heating is $E'=\alpha_w \gamma_{atm} \gamma_{day} E \cos\beta=0.9 \cdot 0.6 \cdot 0.4 \cdot 1320 \cdot 0.7=200 \text{ W/m}^2$. Heat transfer with ambient air is $h_a(T_f-T_a)$, where the convective coefficient may be $h_a=10 \text{ W/(m}^2 \cdot \text{K)}$ without wind or $h_a=30 \text{ W/(m}^2 \cdot \text{K)}$ with 3 m/s permanent wind; typical summer values of ambient air temperature may be $T_a=(-5 \pm 5)^\circ \text{C}$, what enlarges possible heat convection fluxes to $h_a(T_f-T_a)=(20 \pm 10) \cdot (-5 \pm 5)=(-100 \pm 150) \text{ W/m}^2$. Heat loss by radiation to the sky also has great uncertainty because of the equivalent sky temperature, which may be around 0°C on cloudy skies, or below -30°C in clear sky (daily averaged); with infrared emissivity of water $\varepsilon_w=0.9$ and Stefan-Boltzmann constant $\sigma=5.67 \cdot 10^{-8} \text{ W/(m}^2 \cdot \text{K}^4)$, the heat flux lost by radiation is $\varepsilon_w \sigma (T_f^4 - T_{sky}^4)=(53 \pm 53) \text{ W/m}^2$. We further take for the density of ice $\rho_s=400 \text{ kg/m}^3$ (notice that it is firm ice, i.e. recrystallized snow from previous seasons, with density in the range $\rho=300..500 \text{ kg/m}^3$, or a few-years-old ice, not compact ice with $\rho_s=917 \text{ kg/m}^3$), and for the melting enthalpy of ice $h_{sl}=334 \text{ kJ/kg}$.

With the above values the thawing speed would be:

- For clear sky conditions with $E'=200 \text{ W/m}^2$ and $T_{sky}=-30^\circ \text{C}$:

- On warm days ($T_a=0$ °C): $d\delta_w(t)/dt=(200+0-106)/(400\cdot 334\cdot 10^3)=0.7\cdot 10^{-6}$ m/s=60 mm/day.
- On cold windy days ($T_a=-10$ °C): $d\delta_w(t)/dt=(200-250-106)/(400\cdot 334\cdot 10^3)=-1.2\cdot 10^{-6}$, meaning that the ice cover does not thaw but the ice further cools down.
- For cloudy skies conditions with $E=20$ W/m² (assumed incoming diffuse solar radiation) and $T_{sky}=0$ °C:
 - On warm days ($T_a=0$ °C): $d\delta_w(t)/dt=(20+0-0)/(400\cdot 334\cdot 10^3)=0.15\cdot 10^{-6}$ m/s=13 mm/day.
 - On cold windy days ($T_a=-10$ °C): $d\delta_w(t)/dt=(20-250-0)/(400\cdot 334\cdot 10^3)<0$, i.e. no thaw.

In conclusion, this model of Greenland ice surface thawing shows that it occurs at places where daily-mean temperatures are not freezing, on both sunny and cloudy days, with the thaw layer growing at a rate of some (4±3) cm a day (i.e. some (4±3) m of ice melt in the three-month summer period); and thawing cannot take place in freezing days ($T_a<0$ °C), in spite of the 200 W/m² solar heating.

The above model is rather crude, with several parameters having large uncertainty as mentioned, but it is able to clearly show the effect of each of the participating parameters. The model could be checked against measurements if, besides the image-sequence of Greenland ice thaw above mentioned, the threshold used by satellite data processing to define thaw was known, since detailed meteorological data can be found in the Internet.

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