

Chapter I

AQUATIC ENVIRONMENT

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The most important chemical compound that enables the existence of living organisms on Earth is water. Circulating in the solid (ice), liquid, and gaseous (vapor) states, water supports the activities of living organisms on Earth. A hydrosphere is that in which water exists in either of the phases, and is described as the mass of water found on, under, and over the surface of the Earth. When we accept this definition, and call organisms living in hydrosphere as aquatic organisms, most of the organisms on Earth can be regarded as aquatic organisms. In this book, we therefore define the aquatic organisms as those living mainly in liquid water bodies such as ponds, lakes, rivers or oceans.

It is said that Earth was born 4.6 billion years ago. Compared to the age of the cosmos, 13.7 billion years old, birth of the Earth was a relatively recent event. On the contrary, when we compare the birth of life on Earth, the history of the Earth can be said to be very long. In the early days of the Solar System after its birth, gases and dusts clumped together and formed the micro-planets. The Earth was born from the fusion of these micro-planets. Heat released during this process created the ocean of magma on the Earth, while the gases from the magma, of which, vapor, carbon dioxide and nitrogen were the main components, formed the hot atmosphere on the Earth. With a decrease in the temperature of the Earth, dense cloud covered the upper layer of the atmosphere leading to rain. The rain falling on the hot magma produced vapor in the lower layer of the atmosphere. This process repeated over and over again for a long period. After the fusion of the micro-planets subsided, the temperature decreased gradually and solid crust of the Earth was formed. This is a generally accepted theory on the early history of the Earth. Ocean was produced ca. 4 billion years ago with the formation of the crust. Life originated in the ocean shortly after its formation. Appearance of photosynthetic organisms which produced free oxygen molecules changed the atmosphere of the Earth from reductive to oxidative. After its origin in the ocean, living organisms diversified, adapted to various aquatic environments, and further moved on to the terrestrial environments. This short outline story may make one feel that living organisms evolved along a simple straight line from its origin until the present. Modern ecosystems on Earth have developed through continuous environmental changes, extinction of species, appearance of new

species and adaptation to new environments. Today, in the debate on global warming, the rapidness of the shifts is seriously concerned about its impact on ecosystems, as the living organisms have so far not experienced such a rapid change. Can living organisms adapt to these changes? How will the ecosystems be reorganized? How can human being reap the benefits from the ecosystems, which are indispensable for their existence? We need multi-disciplinary efforts to find a way to answer these questions. For this, understanding of biological activities in present environment is a prerequisite. Let us study aquatic environments, biological activities therein, and its relationship with human society.

(Ken Furuya)

1. FORMATION AND DISTRIBUTION OF AQUATIC ENVIRONMENTS

1.1 *Aqua planet*

The Earth was called “the planet of water”. When we see the picture of the Earth floating in the real black cosmos, the Earth is totally bluish even in land parts just like its name. In fact, 75% of the Earth’s surface is covered with water when we include the ice-covered areas. Water is indispensable for living organisms. Life originated in water and all living organisms are dependent on it. Among the planets of the Solar System, liquid water exists only on the Earth. The distance between the Earth and the Sun enables the existence of liquid water on the Earth. Venus, moving on an inner orbit around the Sun than the Earth, receives much higher solar energy than the Earth, and the heat is trapped by its atmosphere as a greenhouse effect. The surface temperature of the Venus is about 500°C. Mars, the planet circulating on an outer loop than the Earth, has only solid water (ice) because its surface temperature is –60°C. Since the Earth moves on an orbit between Venus and Mars, it receives intermediate flux of solar energy, and its surface temperature calculated from the heat from the Sun is –18°C. In reality, mean temperature is about 15°C because of the greenhouse effect of water vapor and other gasses. Water is a unique matter which can exist in three phases, namely gaseous, liquid and solid, at the surface temperature and pressure of the Earth. Water on the Earth’s surface absorbs solar energy and changes to vapor. Vapor forms clouds in the air and returns to the Earth’s surface as rain or snow. This is a global circulating system of water. If the Earth was smaller than its present size, the gravitational force of the Earth would be too weak to maintain the vapors around the Earth, and consequently the Earth would lose its water. This presumption is plausible by seeing the Moon, which likely had air and water when it was formed, and lost not only the vapors but also the air because of insufficient gravitational force.

Total amount of water in the Earth is $137 \times 10^9 \text{ km}^3$, 97% of which exists in the oceans (Table 1.1). As the second largest pool, 1.7% is permanent ice/glacier, and another 1.7% is underground water. Thus, it is not possible for us to use most

Table 1.1. Distribution of water in aqua sphere and average residence period.

Category	Volume (10^3 km^3)	Percentage (%)	Average residence period
Ocean	1,338,000	97	3,700 years
Polar eternal ice/glacier	24,100	1.7	16,000 years
Underground water	23,400	1.7	300 years
Fresh water lake	91	0.007	10–100 years
Salt water lake	85	0.006	10–10,000 years
Soil water	16.5	0.001	280 days
Water vapor	12.9	0.001	9 days
River	2.12	0.0002	12–20 days

The modified by the author from data in previous literatures including Gleick (1996).

of the water directly. Water in the pond, lake and river is only 0.007%. Residence time of water in each pool, defined as mean time for replacement of whole water in a pool, is listed in Table 1.1. Residence time of the vapor in the air is 9 days. This implies that, if there was no evaporation from the surface water, there would be no vapor in the air within 9 days because of loss of vapor by precipitation. Actually, evaporation occurs constantly from the surface of the oceans and land receiving heat from the sun, and thus a certain level of vapor content is maintained in the air. How much does water evaporate from the ocean? As shown in Table 1.1, residence time of the seawater is 3700 years. The average depth of the ocean is 3800 m. From these values, we can calculate the annual evaporation of the water, which is about 1 m, and ocean would be dried up within 3700 years, if there was only evaporation. The amount of evaporation is recovered through water from the land as riverine flow and direct rain on the ocean. Observed value of annual mean global precipitation is 1 m and there is a balance between evaporation and precipitation. In other words, we can consider the Earth as a huge distillation apparatus driven by solar energy.

The residence time of the river water is 2–3 weeks. We can generalize that it takes 2–3 weeks for water precipitated as rain or snow to flow into the sea, or to evaporate from the river. Our life mainly depends on the river water, which has a short residence time and which is less abundant. Dynamics in water circulation among air, oceans and lands causes an imbalance between the input and output, leading to draught and flooding, and occasionally having serious impacts on our life.

1.2 Nature of water

Water is dipolar, with highly electro-negative oxygen at the negative pole and hydrogen at the positive one. In addition, there exist two single electrons, which are not used for covalent binding. As a result, the force of attraction among water molecules is very strong because of hydrogen bond. The Van der Waals force is a general attracting force between molecules and the force is stronger in

Table 1.2. Nature of water as a medium compared to air at 20°C.

Character	Water	Air	Influence to aquatic organisms
Density	1 g/cm ³	1/800 of water	Get buoyancy
Isobaric specific heat	4.18 J/k/g	1/4 of water	Stability of temperature
Viscosity	1.002×10^3 Pa	2 order smaller	Difficult to settle down
Light absorption	large	small	Attenuated with depth
Oxygen density	6–8 mg/l	20% of air	Risk of oxygen deficiency
Sound	1,500 m/s	340 m/s	Transmission of signal
Phases	3	1	Stability of temperature

heavier molecules. Although the molecular weight of water is 18, the force of attraction between water molecules derived from the hydrogen bond is comparable to Van der Waals force in molecules with molecular weight of 100. This strong attractive force makes boiling point and melting point of water 100°C and 0°C in water, respectively. If there was no hydrogen bond among water molecules, the boiling and melting temperature would be lower than –90°C and –110°C, respectively. This means that all water on the Earth would exist as vapor and there would be no liquid or ice water. In addition to this, water is a good solvent because of hydrogen bond, and can easily break ion bonding of salts such as sodium chloride using polarized power. Furthermore, water bind with other various molecules by hydrogen bond as shown in dissolution of sugar. Using this nature water dissolves many biological substances and provides the environment for chemical reaction.

A comparison of some fundamental physical characteristics of air and water are listed in Table 1.2. Firstly, there is a large difference in density. Organisms can remain afloat in water, while they cannot do so in the air without constant movements. This fact regulates vertical distributions of organisms in water and air. Organisms cannot exist in the air constantly and their habitat cannot depart far from ground. While it is well known that birds can fly across the Himalayas, their flying heights rarely exceed 1 km from the ground. On the contrary, organisms are distributed three dimensionally from the surface to depths exceeding 10 km. This is possible because organisms can float in water. If organisms required large amount of energy to float in water, it would deplete a large portion of their energy storage. This would bring huge disadvantages in living strategy of organisms.

Secondly, comparatively water has higher specific heat, which can be easily deduced from the strong hydrogen bonding of water. Temperature change with unit transfer of heat in the water is smaller than that of air, and accordingly the temperature of water is more stable than that of air. Existence of the three phases of water on the surface of the Earth also contributes to temperature stability in the aquatic environment. For example, let us consider the floating ice bergs in the Antarctic Ocean. When heated by the solar energy, water temperature would increase due to the absorption of heat energy, if there was no ice. However, water

temperature does not increase unless all the ice thaws away, because the heat energy absorbed is used as energy for melting of ice. Inversely, in the case of loss of heat energy from the sea by cooling, temperature decrease is hindered by the formation of ice. In the warm waters in tropical and subtropical areas, heat energy of the sun is lightened up by the evaporation of water. Thus, water helps in buffering the temperature change with heat exchange on the ocean surface.

Viscosity is also higher in water. We can experience the high viscosity of water by feeling the resistance of the water when we walk in the pool. Viscosity works as a resistance to moving and settling of floating bodies. In other words, it is advantageous for floating and disadvantageous for movement of plankton in water, which have limited swimming capability. Thus plankton life is strongly controlled by viscosity of water.

Density of water affects its light absorption and sound propagation. Light is attenuated and sound propagates rapidly in water. Light is absorbed also by air in atmosphere, but it is absorbed more acutely by water, and is attenuated rapidly with depth. Even in the most transparent sea, light intensity at 100 m depth is about 1% of that at the surface. On the other hand, sound can propagate deeper in water than in air.

Concentrations of oxygen, which is indispensable for aerobic organisms, differ considerably between water and air. Anoxic condition rarely occurs on land, except under exceptional conditions, although it is often common in water. We will discuss the details of anoxic conditions in water in the later part of this book.

Seawater contains several inorganic salts as compared with fresh water, and almost all natural elements exist in the sea. Among these, Sodium chloride and Magnesium Sulfate account for 95% of dissolved matters. Major ions in seawater are Na^+ , K^+ , Ca^{++} , Sr^+ , Cl^- , SO_4^- , Br^- , and F^- , with lower concentrations in coastal area and higher in the offshore. The relative ionic composition however, is constant among the different seas.

Because of the presence of inorganic salts, specific gravity of seawater is higher than that of fresh water, and concentrations of dissolved gases such as dinitrogen and oxygen are lower in seawater. Salt content regulates the osmotic pressure which is an important environmental condition for aquatic organisms in determining the distribution of marine, fresh and brackish species.

As summarized in Table 1.2, there are various differences in the characteristics of water and air. Therefore, we can easily infer that aquatic organisms have different nature from terrestrial organisms. Creatures that shift their habitat from land to waters such as whale evolved their life in water. The largest creature on the Earth is blue whale. Blue whales are much huge compared to the African elephant, which is the largest terrestrial creature. Shifting its habitat to the aquatic environment enabled the whale to increase the body weight.

1.3 Vertical zonation of sea

Considering the sea surface as the border, the surface of the Earth can be

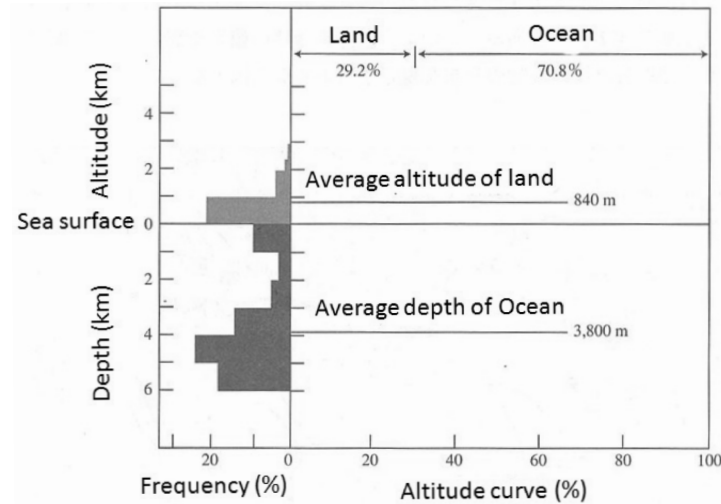


Fig. 1.1. Distribution of altitude from sea surface (left) and altitude curve.

divided in two, the upper region is land and air, and the lower is ocean (Fig. 1.1). The height of Mount Everest, the highest summit on the Earth, is 8,848 m and mean height of the land is 840 m. Pond, lakes, river and inland seas are distributed between 0 m to 8,848 m altitude. The deepest part of the ocean is the Mariana Trench in North Pacific Sea. Its depth is 11,035 m and mean depth of the ocean is 3,800 m. If we trim the undulation on the Earth's surface and fill the depression evenly by cutting and scraping, an ocean with a depth of 2,340 m will cover the surface of the Earth. At the margins of the continent, the shoreline slopes gently offshore and steeps down at the point called the shelf break. The slope before this point is called the continental shelf and that after the shelf break is called the continent slope. The average depth of the continental shelf is 130 m. It is considered that continental shelf was a coastal plain in the glacial age when the shore line was 100 m lower than the present shore line.

Based on the depth, the ocean can be divided into the epipelagic zone (0 m to 200 m), mesopelagic zone (200 m to 1000 m) and bathypelagic zone (deeper part than 1000 m). The bathypelagic zone may be further divided to 3 zones namely, bathypelagic (1000 m to 3000–4000 m), abyssopelagic (3000 m to 4000–6000 m) and the hadal zone (deeper than 6000 m). In the epipelagic zone, phytoplankton produce organic matter through photosynthesis in the presence of sun light, resulting in highest standing stock of organisms in this zone. The mesopelagic zone is located between the epipelagic and bathypelagic zones. It is characterized by diffused light, the intensity of which is insufficient for photosynthesis but allows visibility, and has a low temperature around 5°C. With no light penetration, the bathypelagic zone is dark and cold.

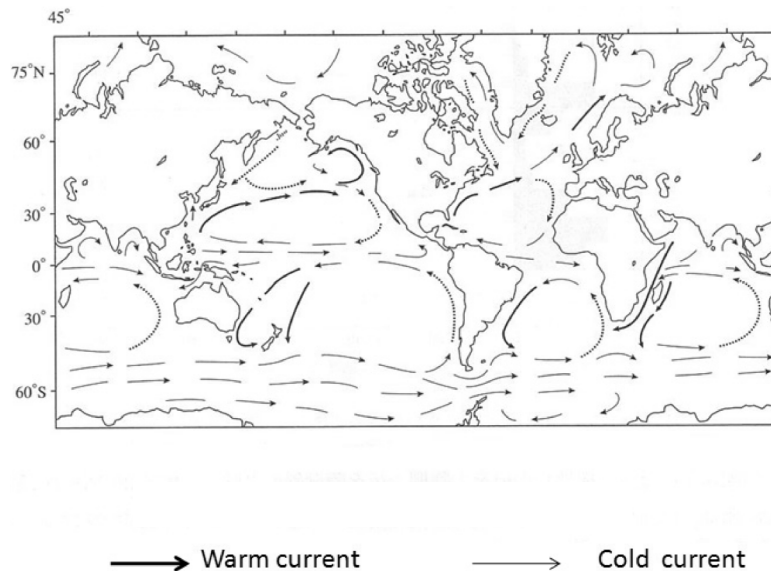


Fig. 1.2. Pattern diagram of ocean current.

1.4 Ocean current

When we see the surface of the sea on a global scale, we can observe the existence of regular flow pattern of the sea surface. These are the ocean currents, which are divided into warm currents, flowing from low latitudes to high latitudes, and cold currents flowing from high latitudes to low latitudes (Fig. 1.2). We can observe the existence of two circular current systems in the North Pacific Ocean formed by warm and cold currents. One consists of Kuroshio (Japan current) in the Western Pacific, California current in the eastern Pacific and the north equatorial current. The other consists of Oyashio current and Alaska currents. The velocities vary according to currents and location, with a velocity of more than 1 m/sec observed in some strong current such as the Kuroshio current. Strong currents caused by the rotation of the Earth are observed on the coast along the western boundaries of the ocean. Kuroshio current, Gulf stream, east Australian current and Brazilian current are some examples. Mechanism of ocean current is described in the later part of this book. Warm currents adds water vapor to the atmosphere and heats up the air in the process of cooling. Therefore, coasts along the warm current are generally humid. On the contrary, cold currents provide little vapor, and coasts along the cold currents tend to be dry and cold. As described above, currents influence the climate through transportation of heat and formation of clouds, and their temperature and salinity regimes regulate the distribution of organisms.

Currents are highly important for fisheries. Migratory fishes utilize the

currents effectively during their life history. Larval sardines that hatch out on the coast of Japan move towards the north through the Kuroshio current, grow up in the off Sanriku region rich in food organisms, and then return along the coast back to their spawning grounds in the south. Many other marine creatures such as the tuna, eel, sea turtles etc. are observed to depend on the currents. Currents are also important for the formation of fishing grounds. The offshore region of Sanriku is rich in plankton stocks and forms a very good fishing ground. This is due to the meeting of the Kuroshio and Oyashio currents and mixing up of the two water bodies in this region. Regions where two different water bodies, such as the warm and cold currents, or the coastal and oceanic water merges, form good fishing ground.

(Ken Furuya)

2. PHYSICAL ENVIRONMENT

2.1 *Physical property of seawater*

Physical nature of water is dependent on its temperature, salinity and pressure. Water density is a function of these variables. Salinity (S) represents the total content (in gram) of dissolved substances in 1000 g of water, and was expressed as parts per thousand (‰ per mil) until 1960s, when water samples were titrated by AgNO_3 and the weight of precipitated AgCl was measured. Salinity was calculated by multiplying the composition ratio to the weight of chloride ($S = 1.80655\text{Cl}$). This method is based on the fact that the composition of substances in sea water is constant. However, this method involves laborious sampling, titration and weighing, and the accuracy was limited at 0.02‰, which is insufficient for deep water studies. The titration method was then replaced by the measurements of conductivity, temperature and pressure as simpler and more sensitive methods. In 1978, PSS78 (Practical Salinity Scale 1978) was introduced, in which salinity was defined as a function of conductivity, temperature and pressure. With the modifications in the measurement techniques, salinity was liberated from the concept of relative weight. No unit is then added to salinity, or psu (Practical Salinity Unit) is added as the salinity unit. Conductivity, Temperature, Depth (CTD) sensors developed as monitoring equipment for pressure, conductivity and temperature, helped to considerably simplify the measurement of physical characteristics of sea water. However, performance of the sensors changes with time due to the attachment of settling (bio-fouling) organisms and other factors. Hence, it is necessary to calibrate the CTD using standard sea water, salinity of which is measured by accurate salino-meter under temperature regulated conditions.

Pressure of 1 m thick water is about 0.1 barometric pressure. The unit of barometric pressure is bar. The unit dbar (deci bar) is generally used in oceanography as an alternative to indicate water depth. Deci means one tenth. Pressure of seawater at 1000 m depth corresponds to about 1,010 dbar. 1 m seawater = 1 dbar = 10^4 Pa = 10^2 hPa (h represents hector or 100, Pa represents

Pascal).

Density of sea water varies from 1,020 to 1,070 kg/m³. Sigma (σ), the unit used for density of seawater, is obtained by subtracting 1,000 from the value of density. Sea water has various substances dissolved in it. Density of pure fresh water without dissolved substances reaches the maximum at 4°C, while seawater containing various dissolved substances reaches the maximum density at freezing temperature. The density of sea water is lower in higher temperature at the same salinity and pressure, higher in higher salinity at the same pressure and temperature, and is higher in higher pressure at the same temperature and salinity. When we place sea water at the same temperature and salinity under different pressure conditions, the densities of the sea waters are higher in higher pressure, due to compression under pressure.

The density at a certain salinity, temperature and pressure is called in-situ density and is represented as σ_{stp} . For example, the density of seawater; $\sigma_{s=35\text{psu}, t=0^\circ\text{C}, p=0\text{dbar}} = 28.13$, changes to $\sigma_{s=35\text{psu}, t=0^\circ\text{C}, p=4000\text{dbar}} = 48.49$. Refer to Appendix 1 for calculation of the in-situ density.

When seawater is displaced to different pressure (depth) without heat exchange, adiabatic expansion (in case of from deep to shallow), or adiabatic contraction (in case of from shallow to deep) occurs and seawater temperature decreases for the expansion or increases for the contraction. When the seawater at 0 dbar pressure, 35 psu salinity and 5°C temperature moves adiabatically to 4000 dbar, the temperature changes to 5.45°C. The temperature of seawater thus changes with the changes in pressure. Temperature values varying with the changes in pressure are not convenient for the identification or comparison of sea water. Potential temperature (θ) is sometimes used instead of in-situ temperature to identify the water. θ is defined as the temperature when the water is adiabatically moved to standard (for example at the surface) pressure. Potential temperature at the surface, where pressure $p = 0$ dbar can be calculated using equation in Appendix 2. Two water bodies that have the same θ and S could have the same origin.

The density obtained from the equation of in-situ density in Appendix 1 for the value of salinity, potential water temperature (θ) and standard pressure (Pr), $\sigma_{s,t=\theta,p=\text{Pr}}$, is called potential density. For ocean shallower than 1000 m, potential density at sea surface, σ_0 or σ_θ are often used. When sea water has no contact with air at the surface or bottom and mixing is weak, water tends to move along the same potential density keeping the same potential temperature and salinity. Therefore, tracing conservative properties such as potential temperature or salinity is a useful method to infer the distribution of currents or the origin of seawater in the ocean interior. For depths shallower than 500 m, sigma-t is also used ($\sigma_t = \sigma_{s,t=T,p=0}$), where the density is calculated for in-situ temperature, salinity and surface pressure.

2.2 Outline of ocean hydrographic structure

Driving force that produces currents and internal structure of ocean is mainly

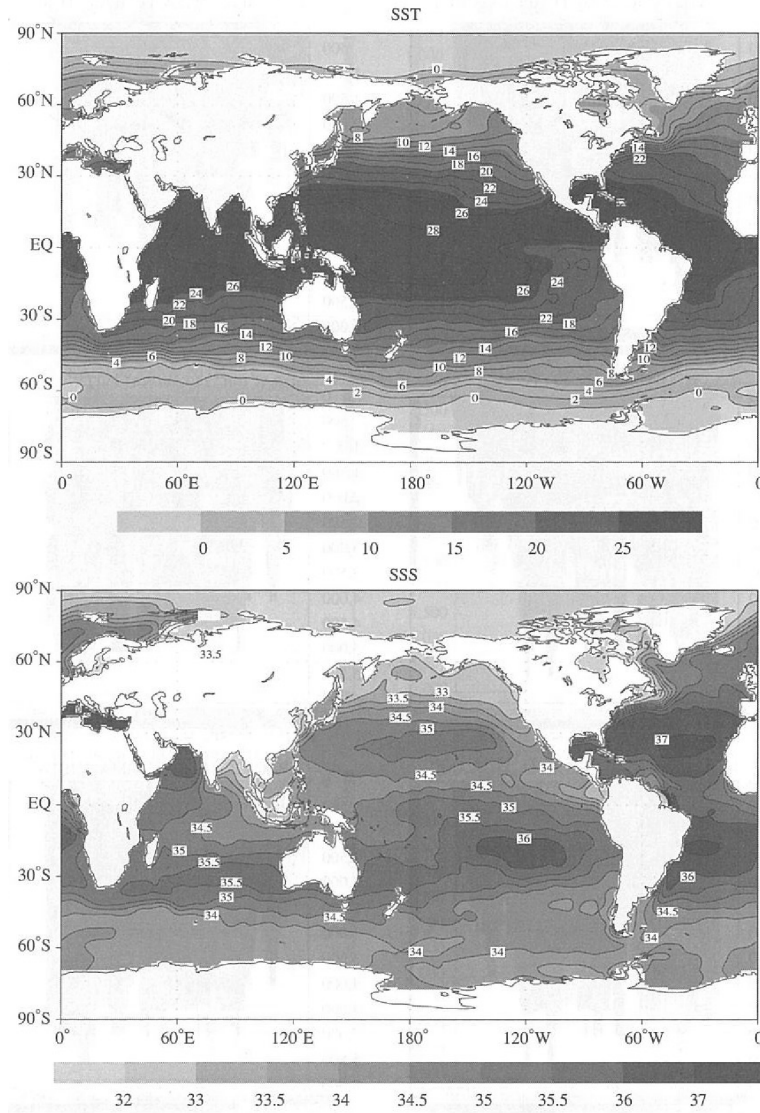


Fig. 1.3. Horizontal distribution of sea surface temperature (SST: upper) and sea surface salinity (SSS: lower).

at the surface where sea water is in contact with air. As shown in the distribution of annually averaged sea surface temperature (SST) (Fig. 1.3 upper), SST is low in high latitudes (minimum temperature is -1.8°C , the freezing temperature for ice formation) and high in low latitudes (maximum is about 30°C). The isotherms are generally zonal because SST is affected by solar radiation which depends on

latitude. Regions where the isotherms are not zonal, usually have special reasons: the existence of heat transportation by strong currents (Kuroshio current along the coast of Japan etc.), cold upwelling regions by equator-ward alongshore winds along the east coast of Pacific and by easterly Trade Wind along the eastern equatorial Pacific etc., and low temperature regions by strong vertical mixing (Kuril island regions and Aleutian Islands etc.).

Heat exchange at sea surface consists of heating by solar radiation (Q_s , ranging 0–400 W/m²), cooling by long wave radiation from the surface (Q_b , Ca –50 W/m²), heat loss by latent heat flux through evaporation of sea water (Q_e), and heat exchange of sensible heat flux caused by temperature difference between sea water and air (Q_h , ± 100 W/m²). Sum of these heat exchanges, ($Q_s + Q_b + Q_e + Q_h$), is net heat flux (Q_{net}). Positive value of Q_{net} means heating of sea water and negative value of Q_{net} means cooling of sea water. In a long term period (more than 1 year), sea water temperature does not change even in the place where the Q_{net} is positive (negative), because the heat increase (decrease) is compensated by horizontal heat transport or vertical diffusion of heat. The latent heat flux, Q_e , and the sensible heat flux, Q_h , are both proportional to wind speed over the sea surface. Cooling by Q_e is enhanced when dry and cold air blows over the warm ocean. Surface heat exchange through Q_h increases with the increase of temperature difference between SST and air temperature. Therefore, cooling is enhanced in winter along the western boundary currents such as the Kuroshio south of Japan, the Gulf Stream south of U.S.A. and Tsushima warm current regions in the Japan Sea.

Distribution of sea surface salinity (SSS) is different from SST as shown in lower panel of Fig. 1.3. High salinity is observed along 20° latitude. Salinity is generally low in high latitudes. When we compare the Pacific Ocean with the Atlantic, salinity is higher in Atlantic Ocean. Some low salinity regions are seen near the coasts. Sea surface salinity distribution is determined mainly by evaporation, precipitation and river discharge. In regions where evaporation (E) exceeds precipitation (P) plus river discharge (R), that is $E - P - R > 0$, salinity tends to be higher. On a long term basis (over a year), SSSs do not change, because the excess salinity is transported to ocean interior by diffusion and currents, and salinity budget is balanced.

In the Labrador Sea in the North Atlantic, high saline water transported from low latitudes is cooled and sinks down to the bottom in winter forming the North Atlantic Deep Water (NADW). On the contrary, in high latitudes of the North Pacific, surface salinity is low because of excess precipitation without deep water formation. Sea-ice forms at freezing temperature. Since the sea ice includes little salt, sea-ice formation releases low temperature, high salinity and high density water called “brine”. Sometimes this brine plays a role in forming deep or intermediate water.

Meridional vertical cross-sections of potential temperature θ , salinity S and potential density σ_θ at 29.5°W for the Atlantic and 159.5°W for the Pacific are shown in Fig. 1.4. θ and S are conservative properties that tend to keep the original values along currents in the ocean interior. Potential temperature θ is less

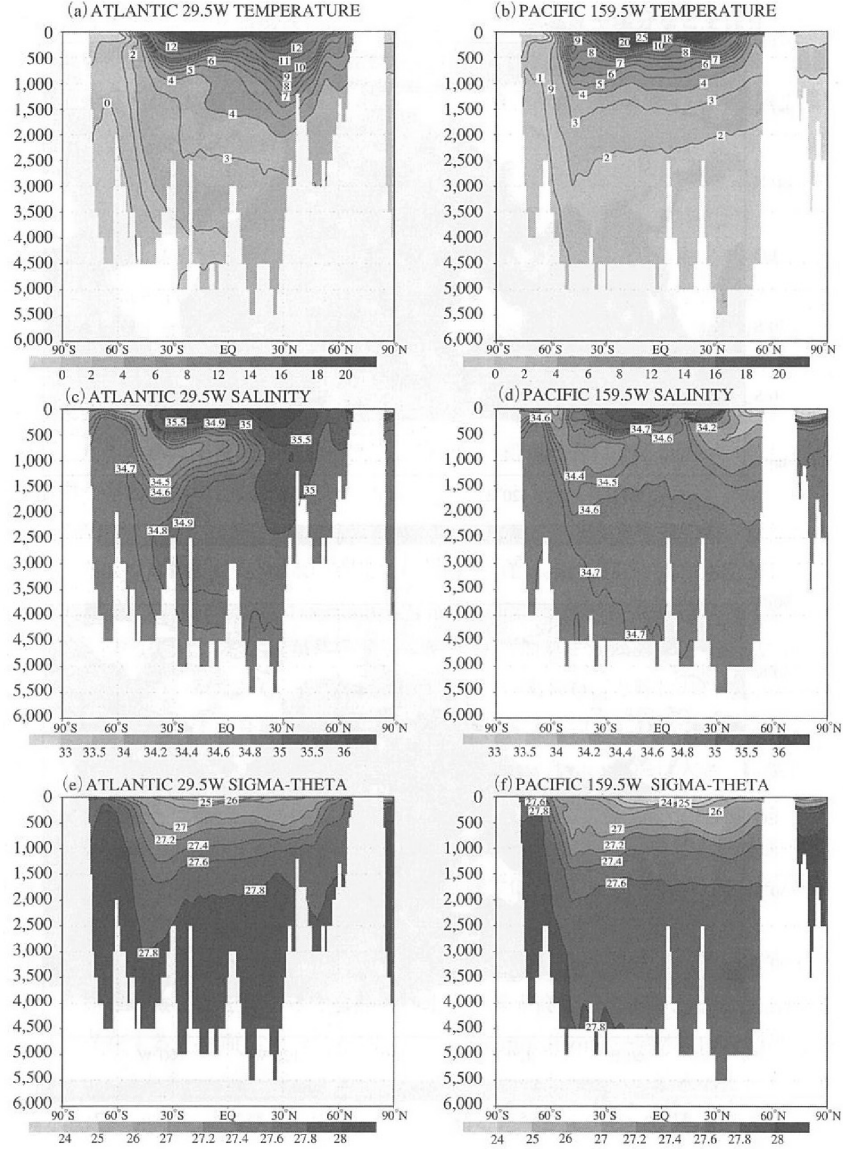


Fig. 1.4. North-south cross section diagram of potential water temperature (upper), salinity (middle) and potential density σ_θ (lower) in the central part of Atlantic (left) and Pacific (right) Ocean.

than 5°C in the ocean deeper than 1000 m; large part of the ocean is thus colder than 5°C. At depths of 500–1500 m, warm and low density water exhibits downward convex shape at around 30° latitude in both hemispheres. These convexes show the existence of subtropical circulation.

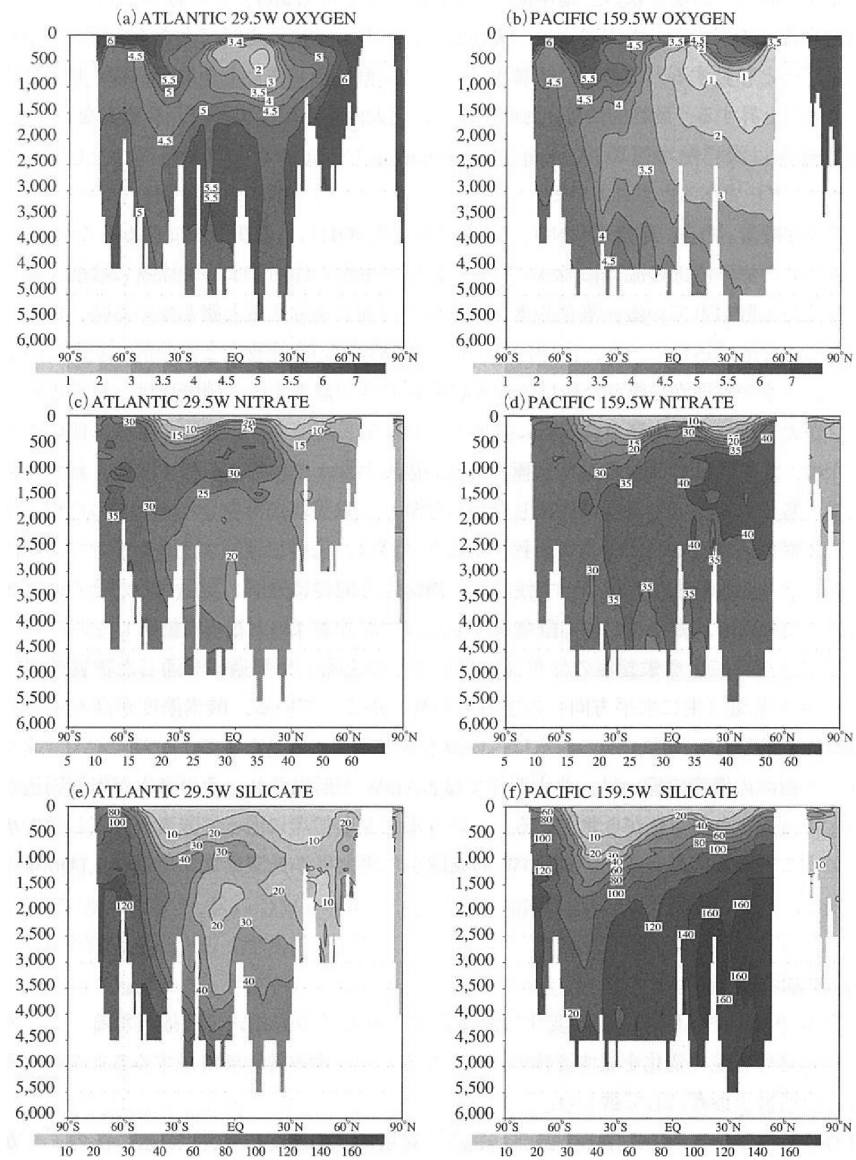


Fig. 1.5. North-south cross section diagram of dissolved oxygen density (upper), nitric acid density (middle) and silicic acid density (lower) in the central part of Atlantic (left) and Pacific (right) Ocean.

Distributions of salinity are more complicated than potential temperature and reflect circulations of intermediate and deep water masses. NADW is produced by sinking of surface water near the coasts of Greenland or Labrador Sea. Along the 3000 m layer in the Atlantic Ocean, NADW extends to the south, that makes the deep layer of Atlantic Ocean warmer, saltier and denser than the deep Pacific. Antarctic Bottom Water (AABW) with relatively low temperature and low salinity sinks along the coast of Antarctica. Origin of deep water in the Pacific Ocean is considered to be the mixture of NADW and AABW produced by upwelling of NADW that is cooled and diluted in surface around Antarctica. In the intermediate depths of 500–1000 m, low salinity waters called intermediate water is observed. This is characterized by the existence of vertical salinity minimum. The low salinity water in the southern hemisphere is called Antarctic Intermediate Water (AAIW), and the one in the North Pacific is called North Pacific Intermediate Water (NPIW). The high temperature and salinity water around 30° in the layer from 500 m to 2000 m in the North Atlantic is called Mediterranean Water.

Figure 1.5 represents vertical cross-sections of dissolved Oxygen (O_2), dissolved nitrate (NO_3^-) and silicate ($Si(OH)_4$). Being different from conservative properties of potential temperature and salinity, concentrations of these properties change along with water movements due to decomposition or biological activities, and are called non-conservative properties. Oxygen in the air rapidly dissolves sea water to nearly saturation at surface when sea water contacts with the overlying air. In the ocean interior, oxygen is consumed by the decomposition of organic substances originally produced from biological activities near the surface euphotic zone. After the departure from surface, concentration of dissolved oxygen thus gradually decreases along the water pathway.

Nitrate is an essential nutrient for production of phytoplankton by photosynthesis near the ocean surface. Dissolved nitrate is utilized during phytoplankton production; nitrate concentration is generally quite low near the surface in low-mid latitudes. Biological wastes and excretory substances aggregate and form particles that sink down to intermediate and deep oceans. In the process of sinking, the particles are decomposed and nutrient is remineralized. Therefore, the concentration of nitrate is relatively high in older intermediate and deeper waters.

Silicate is also used as nutrient by phytoplankton such as diatoms which forms sinking particles of opal. These particles containing silica are also decomposed and silicate is remineralized. Silica decomposition occurs at greater depths than the nitrate decomposition; silica maximum is thus deeper than that of nitrate in the Pacific Ocean.

As described above, distributions of dissolved oxygen and nutrient are determined by a combination of circulation of sea water (mainly for horizontal direction) and circulation of materials (mainly for vertical direction) through biological activities, sinking particles and decomposition of biogenic materials. It can be roughly considered that sea water containing high oxygen and low nutrient spends relatively short time after the contact with air and is thus younger.

In contrast, the water containing low oxygen and high nutrient is relatively old. For the difference in concentrations of oxygen and nutrient between the Atlantic and the Pacific, it can be explained by the ocean deep circulation that NADW produced in the North Atlantic eventually flows into the north Pacific through the Southern Ocean, rising very slowly by vertical diffusion. It takes about 1000 years from formation of NADW to North Pacific Low oxygen regions.

2.3 Introduction of fluid dynamics

Movement of water like ocean currents and changes in temperature and salinity of water can be described by equations in fluid dynamics and thermodynamics. Here the author tries to explain the laws of fluids that represent changes of physical environments in sea, lake and river. For further detailed information, readers are recommended to read relevant books.

Movement of water basically follows Newton's second law ($m \cdot \vec{a} = \vec{F}$, where m is mass of a point particle, \vec{a} acceleration and \vec{F} force working on the point particle). Strictly, water is not a point particle and can change its shape and contact with surrounding area. But let us here regard a infinitely small rectangular cuboid water, of which length are δx , δy and δz in the direction of x , y , z , as a point particle, for simplicity, and consider the forces on the point particle. When the density of the water is ρ , and mass is m ,

$$m = \rho \cdot \delta x \cdot \delta y \cdot \delta z.$$

As forces working on water that will be described from the next paragraph, pressure gradient force works from high to low pressure, Coriolis force works on a moving particle to the right (left) direction perpendicular to the moving direction in the northern (southern) hemisphere on the rotating Earth, and gravity force works in the vertical direction.

Pressure $p(x, y, z, t)$ works on area, and is scalar quantity that has no direction. The pressure gradient force in the x -direction is expressed as the difference of pressure force $p \cdot \delta y \cdot \delta z$ (pressure multiplied by area) working on the two areas whose distance is δx .

$$-(p(x + \delta x, y, z) - p(x, y, z)) \cdot \delta y \cdot \delta z,$$

where the negative sign means that the force works from high to low pressure. Since this force is equal to mass multiplied by acceleration $m \cdot Du/Dt$,

$$\rho \cdot \delta x \cdot \delta y \cdot \delta z \cdot \frac{Du}{Dt} = -\left((p(x + \delta x, y, z) - p(x, y, z)) / \delta x\right) \cdot \delta x \cdot \delta y \cdot \delta z$$

where u is the velocity in the x -direction, and Du/Dt is the velocity acceleration in the x -direction. Similar arguments can be applied for y and z direction. Therefore, Newton's second law with the pressure gradient force can be expressed

as,

$$\rho \cdot \frac{D\vec{u}}{Dt} = -\nabla p, \quad \left(\vec{u} = (u, v, w), \quad \nabla p = \left(\frac{\partial p}{\partial x}, \frac{\partial p}{\partial y}, \frac{\partial p}{\partial z} \right) \right).$$

All things on the Earth including human being and sea water rotate 360° in one day. We usually do not notice this rapid rotation in a daily life; though we are moving very fast in the view from space. All the things receive a gravity force toward the center of the Earth which balances with the centrifugal force; they are moving in a uniform rotation with constant rotational speed. For the description of motion of oceanic currents, it is convenient to describe relative motion to the rotating Earth, that is, on the moving frame which rotates exactly with the Earth, because we are observing from surface of the Earth. The resulting equation can be obtained by coordinate transformation of the equation $\rho \cdot D\vec{u}/Dt = -\nabla p$. This procedure means make \vec{u} relative velocity to the Earth.

The equation is expressed as follow

$$\rho \cdot \frac{D\vec{u}}{Dt} = -\nabla p - 2\rho \cdot \vec{\Omega} \times \vec{u} - \rho \cdot \vec{\Omega} \times (\vec{\Omega} \times \vec{r}).$$

Where $\vec{\Omega}$ is the angular speed vector of the Earth's rotation. Please refer to text books on fundamental mechanics for the derivation of this equation. The second term on the right hand side is the Coriolis force and the third term is the centrifugal force. Adding the gravity force and combining the gravity force and the centrifugal force into modified gravity force, the final equation of motion is obtained as

$$\rho \cdot \frac{D\vec{u}}{Dt} = -\nabla p - 2\rho \cdot \vec{\Omega} \times \vec{u} - \rho \cdot \vec{g},$$

where \vec{g} is modified gravity acceleration.

This equation of motion is for a point particle which moves by its velocity \vec{u} whose acceleration is $D\vec{u}/Dt$, and the position of the point particle changes from time to time. We call this type of time derivative D/Dt , as Lagrange derivative, that is the derivative on the moving frame of reference attached to the moving point particle. This Lagrange time derivative is distinguished from the time derivative on the fixed (not moving) frame $(\partial/\partial t)_{x,y,z}$, that we call Euler derivative. Since potential temperature (heat) and salinity of the point particle (water particle) do not change with time under the condition of no exchange of heat and substances with surrounding environment, the time change rate of potential temperature and salinity can be expressed by Lagrange time derivative as follows

$$\frac{D\theta}{Dt} = 0, \quad \frac{DS}{Dt} = 0.$$

(Caution: actually, there exists exchange at surface with air, molecular diffusion and molecular viscosity. We can neglect them in large scale phenomena far from the surface.) This is the equation describing time changes in heat and salinity along with a point particle (fluid particle). In numerical calculations using computer, it is convenient to express the time derivative of velocity acceleration, water potential temperature and salinity on fixed (not moving) frame, that is, expression by the Euler derivative. The equation can be transformed as follows.

When a fluid particle at a time t and a position (x, y, z) drifts to the position $(x + \delta x, y + \delta y, z + \delta z)$ after a infinitely small time δt , the changes of a quantity A (including velocity \vec{u} , water potential temperature θ and salinity S) can be expressed as

$$\delta A = A(x + \delta x, y + \delta y, z + \delta z, t + \delta t) - A(x, y, z, t) = \nabla A \cdot \delta \vec{x} + \partial A / \partial t \cdot \delta t,$$

where $\delta \vec{x} = \vec{u} \cdot \delta t$. When both sides of the equation are divided by δt , the equation is transformed as

$$\frac{DA}{Dt} = \frac{\partial A}{\partial t} + \vec{u} \cdot \nabla A.$$

Using this equation, the Lagrange time derivative can be expressed by the Euler derivative. The first term on the right hand side is the time change rate of a quantity A on a fixed position, and second term on the right hand side expresses the changes by quantity transport from upstream, which is called as advection. The second term is a form of product of quantity and current velocity, and this nonlinearity often makes solution difficult to be obtained in a simple way. The equation of fluid motion is simple itself as shown here; however this simple equation is a basic tool to express variety of fluid phenomena as ocean currents, waves, eddies and turbulence, where nonlinearity often plays roles and makes problems difficult.

Fluids are connecting continuously with each other. For volume conservation, total budget of water which comes to and goes from a small cuboid must be balanced. We can then obtain following continuity equation

$$\nabla \cdot \vec{u} = \partial u / \partial x + \partial v / \partial y + \partial w / \partial z = 0.$$

Another important equation is the equation of state, which describes the relationship among density, water potential temperature, salinity and pressure as

$$\rho = \rho(\theta, S, p)$$

(see Appendix 1 for detailed formulation).

As 7 equations are given for 7 unknown variables, the unknown variables are obtainable at least numerically. It is called as “numerical simulation” to compute currents, temperature and salinity step by step from initial conditions with boundary conditions using computer. The computer simulation is useful to estimate future trend and so on. In the case when the calculation is done experimentally, it is called numerical experiment. Numerical calculation methods for meteorological and oceanographic phenomena have been developed and forecasting in climate and ocean currents is actively performed. However, it is impossible to solve these equations in time and space resolution that can resolve molecular diffusion and molecular viscosity to resolve turbulence, even using most advanced computer. Therefore, simplified equations obtained by approximation are used in numerical simulations. Recently, ecological numerical simulations are performed including biological and chemical process in the numerical simulation.

In the end of this section, let us list 7 equations in a little bit simplified formulation,

$$\frac{Du}{Dt} = fv - \frac{(\partial\rho/\partial x)}{\rho_0} \quad (1)$$

$$\frac{Dv}{Dt} = fu - \frac{(\partial\rho/\partial y)}{\rho_0} \quad (2)$$

$$\frac{Dw}{Dt} = -\frac{(\partial\rho/\partial z)}{\rho_0} - \frac{\rho g}{\rho_0} \quad (3)$$

$$\frac{D\theta}{Dt} = 0 \quad (4)$$

$$\frac{DS}{Dt} = 0 \quad (5)$$

$$\nabla \cdot \vec{u} = 0 \quad (6)$$

$$\rho = \rho(\theta, S, p) \quad (7)$$

where $f = 2\Omega \sin\theta_{\text{LAT}}$. f is called the Coriolis parameter that is determined with rotation rate of the Earth ($\Omega = 2\pi/T$ ($=86400$ s)) and latitude θ_{LAT} . ρ_0 is a standard density of sea water. Approximation that fix density at certain value except terms

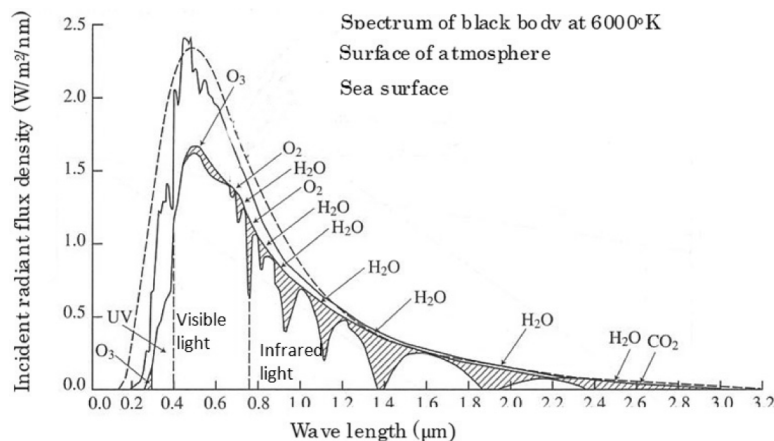


Fig. 1.6. Solar spectrum of black body at 6000°K, surface of atmosphere and sea surface (from Kirk, 1983).

of gravity, second term on right hand side of Eq. (3), is named Boussinesq approximation and is commonly used in numerical calculation in Oceanography. (Ichiro Yasuda)

2.4 Light environment

The solar light reaching the sea surface has a wide spectrum from short wavelength ultraviolet to long wavelength microwaves (Fig. 1.6). Photon energy depends on the wavelength; the shorter the wavelength, the higher is the energy. The photon energy in ultraviolet region is thus higher than that of infrared region. Photon energy between the ultraviolet and infrared regions is suitable to biological light absorbing compounds. Light in the 400–700 nm wavelength is used for photosynthesis and is called photosynthetically available radiation (PAR). In the spectrum of solar light reaching the surface of the Earth, are observed dips in the energy distribution. These dips are due to the absorption of light energy by vapor, oxygen and carbon dioxide. Light absorption by pigments is the initial step in photosynthesis. The light harvesting pigments have conjugated double bonds, and excitation energy of π electron in conjugated double bonds correspond to light energy of PAR. The excitation energy is transferred to other compounds successively through the electron transport chain, where high energy compounds (ATP) and reductants (NADPH) are produced for photosynthesis.

Intensity of light is expressed in terms of photon number or energy by a unit of $\text{mmol photon/m}^2/\text{s}$ or mW/m^2 , respectively. In the former unit, the term photon is usually left out. Light intensity at the sea surface on a clear day ranges from 2000–2500 $\text{mmol/m}^2/\text{s}$ at the solar culmination in the subtropical regions. A portion of solar radiation is reflected on the surface and the rest penetrates into

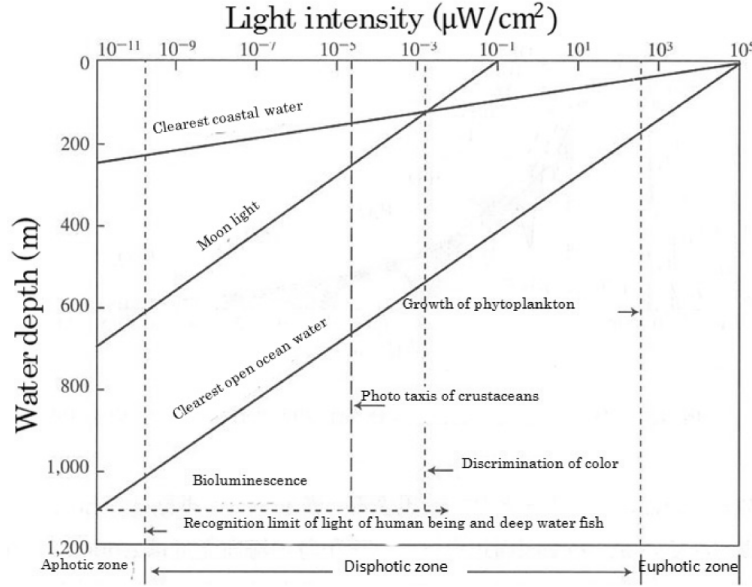


Fig. 1.7. Pattern diagram of transmission of solar light and moon light in most transparent open ocean and coastal water and minimum responsible light intensity of optical receiver (from Parsons *et al.*, 1984).

the ocean interior. The surface reflectance depends on incident angle of the light and surface wave. Reflection is negligibly small when the light enters within 50° from the orthogonal axis to the surface. Downward irradiance decreases exponentially with depth in the water.

$$E_z = E_0 e^{-kz}. \quad (8)$$

Here, E_z and E_0 denote downward irradiance at depth z and at the surface respectively, and k (/m) is called attenuation coefficient or extinction coefficient. Higher the k is, the more rapidly the light diminishes with depth. The multiplication product of k by transparency as measured by a secchi disk is empirically about 1.7. Therefore, once transparency data becomes available, we can calculate k using the above relationship. Figure 1.7 shows examples of light attenuation in water. The horizontal axis is expressed in logarithmic scale. In this plot, k indicates the slope of lines. In the figure, effects of light on biological activities are also shown. The euphotic zone is the zone where sufficient light penetrates for net primary production to occur; where the amount of organic matter assimilated by autotrophs exceeds that consumed by catabolic processes of respiration by autotrophs. In short, the euphotic zone is a region of positive net production. The bottom of the euphotic zone is called the compensation depth, where total organic

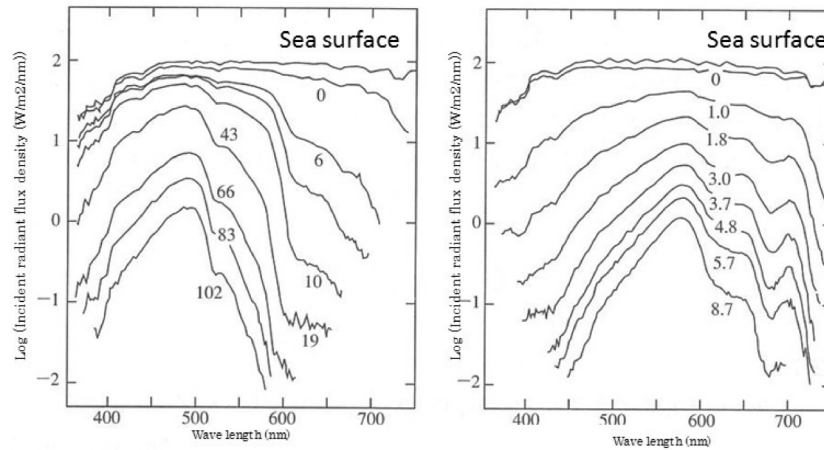


Fig. 1.8. Distribution of downward visible light spectrum in Japan Current (left) and Tokyo Bay (right).

matter production (gross primary production) is equal to loss by respiration. Since intensity of solar radiation changes in the day, the instantaneous compensation depth varies with time. It is located at the surface during night, and reaches its deepest at the solar culmination. Therefore, the term compensation depth is defined on a daily basis, where daily gross primary production equals respiration. The light intensity at the compensation depth is approximately 1% of that at the surface. The compensation depth is shallow in coastal areas and occasionally it is less than several tens centimeters in highly turbid water. In contrast, it is deeper and occasionally reaches 150 m or more in highly transparent open oceans. The disphotic zone is located below the euphotic zone, where no net photosynthesis occurs, but enough light reaches for visual recognition. This zone is also called the twilight zone, and there are many luminescent organisms. The disphotic zone is ecologically important, because biological activity in this zone connects the organic matter production in the euphotic zone and the deep sea. From the view point of light environments, classifications of oceanic vertical zones, namely epipelagic, mesopelagic and bathypelagic zone respectively corresponds to the euphotic, disphotic and aphotic zone where no solar light reaches.

Not only light intensity, but color also changes with water depth. This is because light attenuation depends on wavelength. Red region of PAR diminishes most rapidly with depth followed by violet. Blue green zones become predominant in the PAR spectrum gradually with depth (Fig. 1.8). The changes in light intensity and quality with depth are influenced by the composition and amount of various substances in water including dissolved matter such as salts, organic compounds, particulate matter, that is, plankton, detritus (non living particles originated from organisms), and suspended solids (Fig. 1.9). Total light attenuation

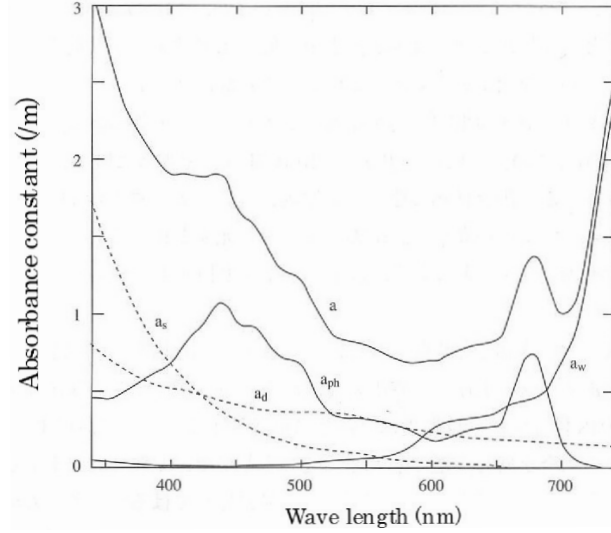


Fig. 1.9. Absorbance constant of whole sea water (a), phytoplankton (a_{ph}), detritus (a_d), dissolved matter (a_s) and sea water (a_w) (Kishino, 1994).

$a(\lambda)$ is expressed by each of these components; water a_w , phytoplankton a_{ph} , detritus a_d and dissolved substances a_s , respectively as follow:

$$a(\lambda) = a_w(\lambda) + a_{ph}(\lambda) + a_d(\lambda) + a_s(\lambda) \quad (9)$$

where, λ denotes wavelength, since the light attenuation is wavelength dependent as mentioned above.

Light attenuation by phytoplankton has distinct absorption peaks at 430 nm and 665 nm. Absorption spectrum of phytoplankton is dependent on the amount and composition of light harvesting pigments including chlorophylls, carotenoids and phycobilin proteins, and varies according to their classes and physiological states. Dissolved organic substances, usually termed as colored dissolved organic matter (CDOM) contain humic substances, and absorb the blue light region. Thus, spectra of downward irradiance in water depend on the amounts and composition of these optically active materials. The attenuation coefficient k is the sum of absorption and reflection by planktons, detritus and dissolved substances. Where amounts of detritus and CDOM are low, light is absorbed mainly by phytoplankton, and blue light region around 480 nm in wavelength penetrates well with depth. In contrast, blue light attenuates with depth due to absorption and scattering of detritus and CDOM and green light region becomes prominent with depth in coastal area. Thus, there is a mirror image relationship between attenuation by substances in water and light environment therein.

(Ken Furuya)

3. CHEMICAL ENVIRONMENT

3.1 Biophile elements

Living organisms require various elements for growth. C, H, O, N, P, Mg, Ca, K, S and Si are required relatively in large amount. In contrast, Fe, Mn, Cu, Zn, B, Mo, V and Co are required in relatively small amounts. These essential elements are named as biophile elements. Relative concentrations of biophile elements between plankton and seawater differ among elements: while carbon and potassium concentrations are higher in seawater, those of nitrogen, phosphorus and silica are higher in plankton body than in sea water. This relative concentration can be regarded as supply and demand ratios of these elements, that is, demands of the elements of which concentrations are lower in seawater exceed their supply and such elements tend to be deficient. Some trace elements such as iron and zinc as well as nitrogen, phosphorus, and silica are in short supply. These elements in short supply play important roles in regulation of biological production in the sea, because these elements often limit primary production. Inorganic salts of nitrogen, phosphorus, and silica, viz NO_3^- , NO_2^- , NH_4^+ , PO_4^{3-} and $\text{Si}(\text{OH})_4$, are called nutrient salts. The most abundant form of nitrogenous nutrients is nitrate. While nitrogen, phosphorus, and potassium are three most important elements in fertilizers used in terrestrial ecosystems for increasing agricultural crop production, potassium exists in excess amount in seawater. Since concentrations of nutrient salts are much higher than trace elements, and are required in higher amounts, nutrient salts are regarded as macro nutrients. Major biophile elements are summarized below:

1) Carbon

Carbon dioxide is dissolved in seawater and concentrations of carbonic ion, bicarbonic ion, dissolved carbon dioxide and its hydrated form are in an equilibrium in seawater (Fig. 1.10). Their total amount in seawater is called total carbonate. Around pH 8, the common pH of seawater, bicarbonate is the dominant chemical form, and the equilibrium shifts toward carbonate dominance at a higher pH, and shifts oppositely at a lower pH. Because of the presence of the equilibrium, seawater dissolves far more amount of total carbonate than expected from the solubility of carbon dioxide. The concentration of total carbonate is about 2 mM in the epipelagic zone. This amount is sufficient for photosynthesis of phytoplankton, and deficiency of carbon dioxide occurs rarely. Overall, concentration of carbon dioxide in the ocean is 50 times higher than that in the air. If 2% of carbon dioxide is released from the ocean into the air, the concentration of carbon dioxide in the air will be doubled. Thus, the ocean is a huge pool of carbon dioxide, and together with its ecosystems plays a crucial role in carbon cycling on this globe.

2) Nitrogen

Nitrogen is an important element for metabolism of organisms, present in protein, nucleic acid and other biomolecules. Major usable forms of nitrogen for organisms are nitrate, nitrite, ammonium, and organic nitrogen such as amino acids and urea. Concentrations of these chemical components in seawater range

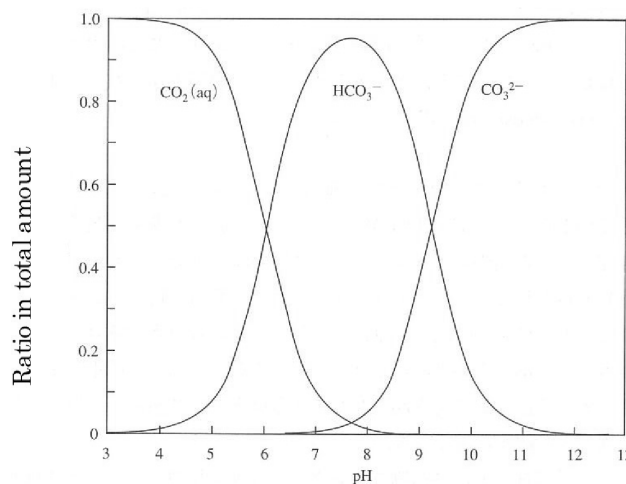


Fig. 1.10. Form of carbonic acid in sea water. Total carbonic acid exists in equilibrium of carbon dioxide (CO_2), carbonic acid (H_2CO_3), bicarbonate ion (HCO_3^-) and carbonate ion (CO_3^{2-}) in sea water.

as follows: nitrate: 0–40 μM , nitrite: 0–5 μM , ammonium: 0–10 μM , amino acids: 0–2 μM , and urea: 0–5 μM . Here the zero concentration indicates an amount below the detection limit, and does not imply 0. In coastal waters which receive sewage discharge, concentration of these nitrogenous compounds may exceed the above ranges. Nitrogen exists in large quantity in seawater as dissolved molecular nitrogen (dinitrogen N_2). Its saturation concentration in seawater depends on salinity and ranges between 370 and 800 μM . Dinitrogen is chemically stable compared with the above nitrogenous compounds, and cannot be directly utilized by organisms other than diazotrophs. Similar to the diazotrophs, some species of blue-green alga (cyanobacteria), eubacteria and archaea can fix dinitrogen to ammonium. These organisms have an advantage over other plankton in areas where nitrogenous nutrients are deficient.

3) Phosphorus

Phosphorus mainly exists in three forms in seawater, namely, dissolved phosphate (PO_4^{3-}), dissolved organic phosphorus, and particulate organic phosphorus. Generally, phytoplankton utilize dissolved inorganic phosphate, and some phytoplankton can use dissolved organic phosphorus. However, these do not directly uptake dissolved organic phosphorus, but have alkaline phosphatase on cell membrane or outside of the cell where dissolved phosphate esters are decomposed, and phosphate is obtained.

4) Silicon

Diatoms, silicoflagellates and radiolarians uptake dissolved silicate and form frustules (cell coverings of diatoms) or cytoskeleton with complex morphological patterns of silica (hydrated silicon dioxide). Concentration of

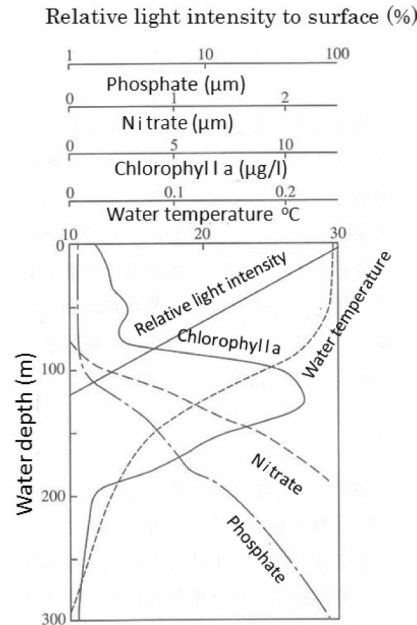


Fig. 1.11. Vertical distribution of water temperature, nutrient salts, light, chlorophyll *a* at latitude 10° north and longitude 154°45' east.

dissolved silicate is high in coastal waters and deep waters in the open ocean, and low in the surface water.

5) Trace metals

Some metals including iron, magnesium, zinc, copper, cobalt, molybdenum and selenium are essential elements for organisms required in extremely small quantities. Some of them can be toxic in high concentration. As a component of heme protein, Iron is important in energy metabolism in cells. Ionic iron is either bivalent or trivalent iron ion, and the latter is a common form in oxidative aquatic environments. Since the solubility of trivalent ionic iron in alkaline water such as seawater is very low, excess iron exists as particles which are not directly utilized by organisms. Therefore dissolved iron concentration is extremely low in the sea, and the iron availability may control biological productivity. Large-scale iron deficiencies occur in the North Pacific subarctic water, Pacific equatorial region and Antarctic Ocean. Major sources of iron supply to the ocean are river discharge, upwelling from deeper layers and dust deposition from the air. In particular, dust deposition is significant in distant oceanic regions from land mass, where other sources are insignificant. Yellow sand from the Chinese desert is an important iron supply for oceanic phytoplankton in the Pacific, while it often causes health issues in the East Asian countries.

Chelating substances in seawater are important in the availability of trace

metals for organisms. When a metal forms a strong complex with a chelating substance, it can be difficult for organisms to absorb the metal. Toxicity of metals can be reduced when they form a complex compared with its free state. While origins and sources of chelating substances are not well understood presently, various biological activities appear to be responsible.

3.2 Chemical environment in open ocean

Chemical environment of the open ocean can be characterized by low concentrations of nutrient salts near the surface. Vertical distributions of nitrate and phosphate in a region where stratification develops in the upper water column throughout the year, shows that both nutrient salts are exhausted near the surface (Fig. 1.11). Concentrations of nutrients increase rapidly with depth in the vicinity of thermocline, that is, below the surface mixed layer. Nutrients are consumed actively by phytoplankton near the surface, but the thermocline prevents upward transport of nutrient salts from the deeper layer. The layer where concentrations of nutrients rapidly increase with depth is the nutricline, which is often located at the depth where the light intensity is 0.1–1% of that at the surface. Below the nutricline concentrations of nutrient salts increase with depth and reach stable concentrations in deeper layer than 1 km (Fig. 1.4). Below the euphotic zone, regeneration of nutrients from organic matter through mineralization by heterotrophic organisms such as zooplankton and bacteria exceeds the consumption by phytoplankton due to light deficiency leading to accumulation of nutrients. Mineralization of organic matter also occurs in the euphotic zone, but the consumption by phytoplankton exceeds the mineralization.

3.3 Chemical environment in coastal waters and estuary

Coastal water is the area between the land and open ocean. Here, we adopt the definition by E. Haeckel who defined coastal water as an area between the coast line and 200 m depth. The 200 m depth corresponds to the outer boundary of the continental shelf. Enclosed and semi enclosed coastal areas such as bays and inland seas are surrounded by land mass. Water exchange between the inside and outside of an enclosed area depends on the topography and depth, and is generally limited. Tokyo Bay, Ise Bay, and Seto Inland Sea are typical enclosed bays.

The coastal waters are under influences of material supply from land, in particular, in the vicinity of highly populated big cities and industrialized areas, where anthropogenic pollutants and nutrients are released as sewage effluent and agricultural run-off carrying fertilizers. Various substances are transported from forests, arable lands and cities into coastal waters, including inorganic substances (nutrient salts, heavy metals, sand, etc.), organic substances (living drainages such as dietary oil and food waste, human excretion, manmade harmful organisms such as PCB and dioxin, plastics, etc.), radioactive wastes and so on. In an extreme rhetoric, as a land-side end of water circulation, coastal water may receive all residuals and wastes from land. Stable toxic substances such as heavy

metals tend to accumulate within organisms through trophic chains. This is called biological concentration. Using of transported substances from land as resources, productivity of marine organisms is generally high in coastal areas. Coastal waters are also influenced by material supply from bottom. Because of its shallow depth, material accumulated on the bottom can easily reach near the surface by re-suspension caused by physical turbulence, such as wind-induced vertical mixing.

Eutrophication is an increase in concentrations of nutrient salts such as nitrogen and phosphorus from an oligotrophic condition (low concentrations) to an eutrophic condition (high concentrations) resulting in high primary productivity of the ecosystem. Eutrophication occurs very slowly in deep lakes as a natural process by sedimentation of sand, soil and plant litters and other biogenic materials. However, recent human-related eutrophication poses serious environmental problems. Eutrophication of some coastal areas, in particular in enclosed and semi-enclosed areas has been accelerated from the mid 20th century in developed countries. As a result, high standing stocks of phytoplankton persist in eutrophic waters, and red tides occur frequently. Typical symptoms of the progress of eutrophication are lowering of transparency (high turbidity), and increase in Chemical Oxygen Demand (COD). The reduced transparency implies attenuation of light by high abundance of plankton and detritus. COD defined as the amount of oxygen required for oxidization of organic substances in water, is the most common measure of eutrophication. At an early phase of eutrophication, both the pelagic and benthic fish catch may increase due to increased biomass of food plankton. However, along with further progress of eutrophication, serious impacts on both biota and chemical environments become evident.

Eutrophication influences not only plankton biomass but also the species composition, because the change in chemical environment poses selective impacts on organisms. For example, anthropogenic eutrophication generally results in increased availability of nitrate, ammonium and phosphate, but not silicate. Therefore, such conditions favor species other than diatoms which require silicate as an essential element. In fact, phytoplankton assemblages in eutrophic waters are often characterized by dominance of dinoflagellates and raphidophytes in western Japan. Such a change in species composition of phytoplankton induces changes in zooplankton assemblages through trophic interaction, and these shifts in plankton composition thus spread to the whole local ecosystem.

Formation of hypoxia and anoxia is a typical negative environmental impact of eutrophication. Accumulation of organic matters such as dead remains of organisms, feces and other detritus proceeds with eutrophication. Organic matters form soft sediments called sludge that are actively decomposed by bacteria and other heterotrophs. Dissolved oxygen is consumed during the process of decomposition. If the consumption of oxygen exceeds supply, oxygen content of water is exhausted to a certain level that affects the biological activity. This state is called hypoxia, and an ultimate condition of hypoxia is anoxia. In summer when thermocline is developed and the water column becomes stratified, seawater between the near-surface and bottom does not mix much, and supply of oxygen

to the bottom is extremely limited. Consequently, anoxic condition prevails near the bottom, under which aerobic organisms cannot live, and are replaced by organisms that can live under reductive conditions. Under anoxia, sulfate reducing bacteria consume organic materials and sulfate as an electron acceptor of electron transport activity. By this, hydrogen sulfide, a toxic substance is produced. In the inner Tokyo Bay and Mikawa Bay in late summer, when prolonged wind blows towards the offshore, a wind-induced upwelling occurs near the coast, and anoxic water is transported from the bottom to surface. Organisms such as clams exposed to this upwelled anoxic water die. The resulting dead remains cause further anoxigenic consumption of sulfate, and thus a negative feedback proceeds. Hydrogen sulfite in the anoxic water is oxidized by dissolved oxygen to colloidal sulfur near the surface, which makes the sea surface greenish blue in color. This phenomenon is called the blue tide. This eerie color of blue tide is the worse consequence of eutrophication on fisheries and coastal ecosystems.

3.4 Chemical environment of inland waters

Water pollution and eutrophication are serious issues in inland water as well. Pollution is caused by contamination of hazardous substances from the air, industrial drainage and sewage. In rivers, amount of organic matter is conventionally monitored by Biochemical Oxygen Demand (BOD), while COD is used as the index of water quality for lakes and ponds. BOD is defined as the amount of oxygen required for decomposition of organic substances by microorganisms within water. Thus, the different indices have been used for flowing (lotic) and stagnant (lentic) water systems.

4. DYNAMICS IN EUPHOTIC ZONE

4.1 Surface mixed layer

Stirring by surface wind or convection by surface cooling develop a surface mixed layer characterized by vertical uniformity of temperature and so on. As will be described in next subsection (4.2), mixed layer depth h and mixed layer temperature T have a strong impact on biological productivity.

The time change rate of temperature (Lagrangian time derivative) DT/Dt in mixed layer is formulated by heat exchange at surface Q (W/m^2) with air and by entrainment of cold subsurface water when the mixed layer deepens. Here we use the notations of entrainment velocity w_e , entrained water temperature T_d , from just beneath the mixed layer to mixed layer, and mixed layer temperature T . Temperature difference between the mixed layer and the entrained water is denoted as $\Delta T = T - T_d$. Change in total amount of heat is expressed as

$$\rho \cdot c \cdot [(h + \delta h) \cdot (T + \delta T) - h \cdot T] = Q \cdot \delta t + \rho \cdot c \cdot \delta h \cdot T_d. \quad (10)$$

Left hand side means change in total heat amount of the mixed layer when the

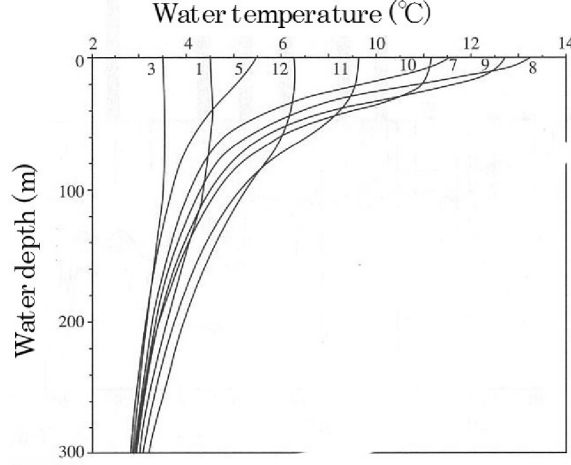


Fig. 1.12. Seasonal fluctuation of thermocline at North Pacific (40.5°N, 145.5°E).

depth and the temperature increase to $h + \delta h$ and $T + \delta T$ after infinitely small time δt . Right hand side is the sum of heat from the overlying atmosphere through surface net heat flux and heat by entrainment water, of which thickness is $\delta h = w_e \cdot \delta t$ where w_e is entrainment velocity. Change in total heat amount and the sum of heat exchange should be balanced. If we neglect $\delta h \cdot \delta T$, because it is quadric product of small quantity, time change of the mixed layer temperature can be formulated as

$$\rho \cdot c \cdot \delta T / \delta t = \rho c DT / Dt = Q - \rho \cdot c \cdot w_e \cdot (T - T_d). \quad (11)$$

Surface mixed layer develops more rapidly under stronger surface wind and stronger cooling conditions when the depth becomes deeper ($w_e > 0$). In warm and heating seasons ($Q > 0$), generally, stratification progress from surface; and formation of seasonal thermocline and mixed layer depth decreases. The layer becomes deep when effects of wind stirring exceed the stratification by heating.

4.2 Surface mixed layer depth and biological production

Surface mixed layer in mid latitudes is cooled and deepened by cooling and strong wind from autumn to winter, becomes shallower in spring that is heating period, and has the highest temperature in summer. This is a typical seasonal change in mid latitudes (Fig. 1.12). Mixed layer deepening can be observed in several events like typhoon. When clear and warm weather continues after strong vertical mixing by typhoon, the mixed layer becomes shallower. As described in Subsection 2.4, phytoplankton can be transported to deep layers where they cannot accept enough amount of light for their growth, when mixed layer is

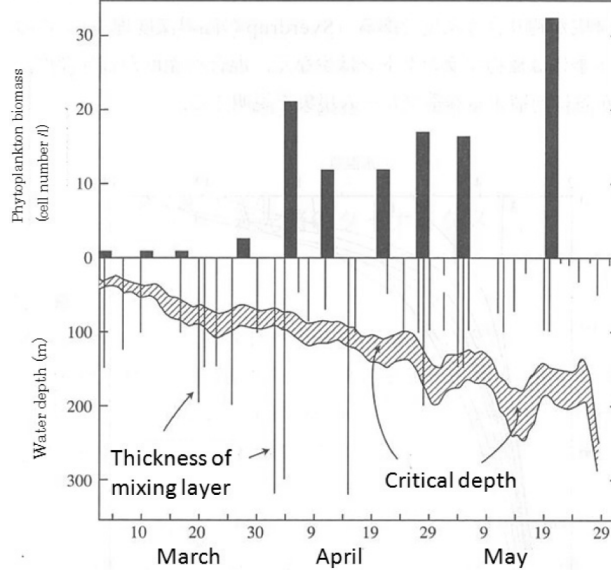


Fig. 1.13. Increase of phytoplankton biomass with formation of mixing layer in Norwegian Sea in spring.

deeper than compensation depth. The maximum depth of distribution of phytoplankton, at which positive population growth of phytoplankton in the mixed layer can be maintained is called critical depth. Critical depth is the depth where rates of respiration and photosynthesis are balanced when we integrate the amount along the depth in mixed layer (Sverdrup's critical depth theory). This theory can explain the phenomena that phytoplankton biomass is smaller in winter season when the mixed layer is deeper than critical depth and that phytoplankton blooms rapidly in early spring when the mixed layer becomes shallower than the critical depth (spring bloom).

Using Eq. (8) in Subsection 2.4, energy for production by photosynthesis at a depth z can be expressed as $E_z = E_0 e^{-kz}$ (k is light attenuation coefficient). When we express the amount of respiration converted to light energy as I_c . Total amount of photosynthesis (left side) and respiration in water column from surface to critical depth are expressed as follows:

$$\int_0^{D_{cr}} E_0 e^{-kz} dz = E_0 \left[1 - e^{-k \cdot D_{cr}} \right] / k = I_c \cdot D_{cr}. \quad (12)$$

Here, we assume that rate of respiration does not depend on light intensity and is vertically uniform.

When mixed layer becomes shallower by heating or critical depth becomes deeper by increasing solar radiation and elongation of day length with time from

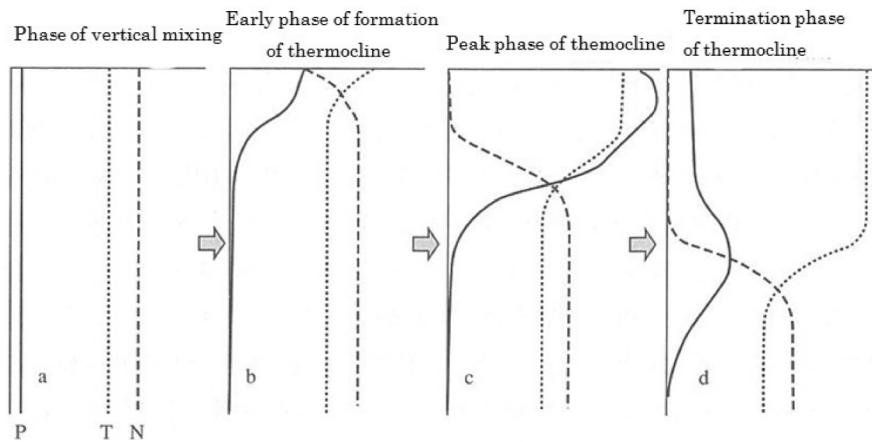


Fig. 1.14. Pattern diagram of time series changes of nutrient salts and phytoplankton biomass with formation of thermocline.

winter to spring, growth of phytoplankton is enhanced near the surface, and mass propagation called spring bloom starts (Fig. 1.13). The changes of mixed layer depth and critical depths are actually complicated and influenced by weather conditions such as passing of low atmospheric pressure and continuous clear weather. Mixed layer depth in Fig. 1.13 largely fluctuates from beginning to mid April. Standing stock of phytoplankton increases when mixed layer depth becomes shallower than critical depth. Figure 1.13 is modified from the original article on the concept of critical depth (Sverdrup, 1953).

Nutrients are other important factors of biological production with the change of mixed layer depth. Water is well mixed vertically, nutrients are supplied from deeper layer, and photosynthesis is limited by light intensity. Ample nutrient are supplied near surface in winter (Fig. 1.14a). In spring when mixed layer becomes shallow, phytoplankton grows actively under suitable light condition using available nutrients (Fig. 1.14b) thereby resulting in spring bloom (Fig. 1.14c). In many cases, the bloom is composed of diatoms. After that, when seasonal thermocline is developed in surface layer by heating, nutrient supply from bottom layer decrease and nutrient in mixed layer are exhausted by phytoplankton consumption. Phytoplankton density shows a vertical peak at the depth where light intensity that decreases towards bottom and concentration of nutrients decreases towards the surface are balanced (Fig. 1.14d). The surface mixed layer depth varies largely during the year in mid and high latitudes. Stratification is well developed and vertical mixing does not reach to the bottom throughout the year in low latitudes. Nutrients are exhausted throughout the year and there are peaks of chlorophyll concentration around the bottom of euphotic zone (Subsurface chlorophyll maximum, or Deep chlorophyll maximum).

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5. MATERIAL CYCLING

5.1 *Concept of primary production*

Organic matter produced by plants through photosynthesis is distributed to various heterotrophic organisms through food chain. Finally, the organic matter is mineralized to inorganic matter by decomposers which are also heterotrophs, and is re-used by plants. The whole process is called material cycling, and occurs throughout the biosphere. Carbon, hydrogen, oxygen and nitrogen compose the main structure of the cycling as major elements and other biophile elements circulate along with major elements. Driving force of the circulation is the light energy from the sun. Thus, production of organic matter by photosynthesis is the starting point of material cycling, thus called the primary production. Primary production is not exactly equal to photosynthesis, but includes the entire processes of organic matter production in which various biophile elements are incorporated into organic compounds originated from carbohydrate produced by photosynthesis. In photosynthesis, glucose is synthesized from carbon dioxide through reduction and chemical energy released during the light reaction of photosynthesis where molecular oxygen is produced as a byproduct of degradation of water. Primary production is limited to the euphotic zone where sufficient intensity of solar light can penetrate. A rough indication of the lower limit is 0.1–1% of surface light intensity. Major primary producers in open ocean are phytoplankton, while macro algae and sea-grass (angiosperms in sea such as *Zosteraceae*) are important in shallow waters. Photosynthetic bacteria use hydrogen sulfite, thiosulfate or organic substances as reducing power of carbon assimilation under anerobic conditions, and do not produce molecular oxygen. Primary production by oxygenic photosynthesis is overwhelmingly large in ocean. Only in reductive environments where sufficient light reaches but dissolved oxygen is very low, photosynthetic bacteria are the primary producer.

Driving force of primary production is not only photosynthesis, but also includes chemosynthesis. In chemosynthesis energies obtained by oxidization of inorganic substances such as hydrogen sulfite, sulfur, and ammonium are used for carbon assimilation instead of light energy. A typical example can be found in hydrothermal vents where no light reaches but reducing substances are supplied from vents. Thus, chemosynthesis depends on supply of reducing substances. It is important locally, but in terms of amounts of organic matter produced, chemosynthesis is minor compared to photosynthesis.

Primary production and respiration functions as consumption and production of carbon dioxide, respectively. As defined in Subsection 2.4, difference between gross primary production P_g and net primary production P_n is respiration R of phytoplankton.

$$P_g = P_n + R_p \quad (13)$$

Net primary production is net amount of organic matter that can be utilized

Table 1.3. Net primary production in a year (Field *et al.*, 1998).

	Net primary production (PgC)	
	Ocean	Land
Season		
April–June	10.9	15.7
July–September	13.0	18.0
October–December	12.3	11.5
January–March	11.3	11.2
Sea area		
Oligotrophic sea	11.0	
Mesotrophic sea	27.3	
Eutrophic sea	9.1	
Seaweed	1.0	
Total	48.5	56.4

by organisms in the ecosystem. This definition is very clear, but in practice, net primary production is difficult to measure. Let's have a brief look at its measurement procedure. First, seawater sample containing phytoplankton is introduced into a number of transparent bottles. Some bottles (light bottles) are exposed to light, and the others (dark bottles) are kept in dark place. Both groups of bottles are incubated for some period. In the light bottles organic matters are produced by primary production (P_g) and a portion of the products are consumed by algal respiration (R_p). Thus the light bottles denote net primary production (P_n). In the dark bottles, photosynthesis does not occur and only respiration (R_p) takes place. Thus, gross production (P_g) can be derived from the above equation. However, by this process we cannot obtain algal respiration, because numerous small heterotrophic plankton including bacteria and zooplankton are present in the water sample along with phytoplankton, and cannot be eliminated from the bottles. Thus, the respiration measured in the dark bottles denotes the sum of respiration by phytoplankton (R_p) and that by heterotrophic organisms (R_h). The sum is called community respiration (R_c), and the above equation should be modified as follow:

$$P_g = P_n + R_c \quad (14)$$

Thus, in planktonic ecosystems, it is impossible to measure net primary production directly, the net production is estimated with some assumptions.

5.2 Primary production in the ocean

Summary of global net primary production is shown in Table 1.3 (Field *et al.*, 1998). Annual net primary production of the ocean is 48.5 PgC (48.5 Gt, 1 Pg =

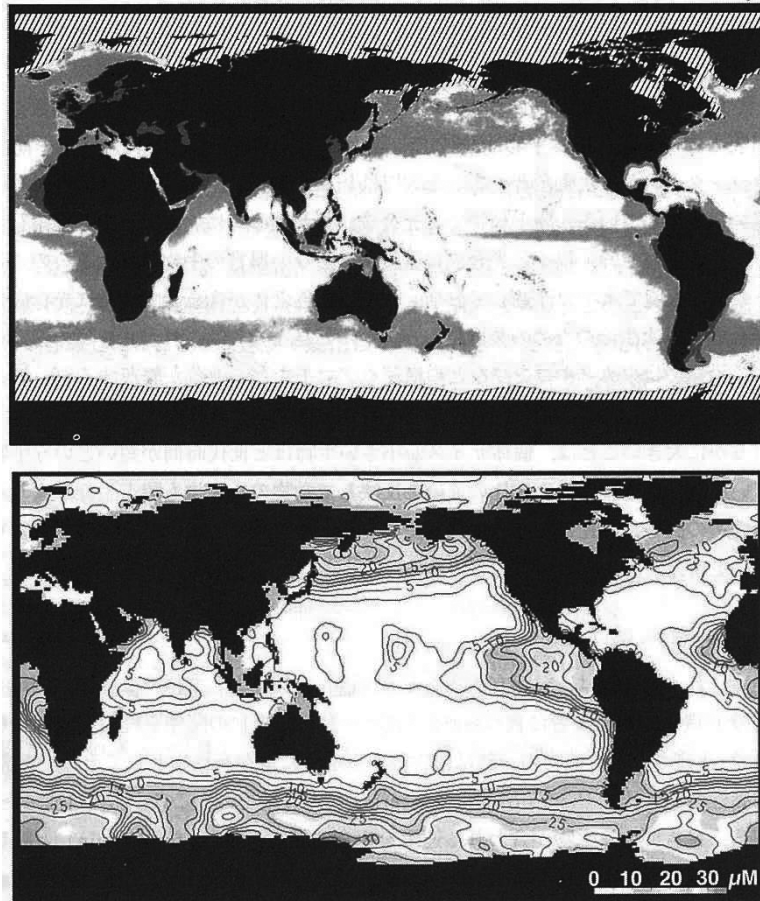


Fig. 1.15. Distribution of annual primary production (upper) and nitrate at 80 m depth (lower). Primary production is high in grey zone in upper map. No data in shaded parts.

10^{15} g). This is 85% of total terrestrial production. Thus, primary production in both regimes is roughly in the same magnitude. For comparison, 7 Pg of carbon dioxide is released into the air annually by human activity. Production by macroalgae is 1 Pg, occupying only 2% of total marine production. The rest is produced by phytoplankton. Primary production by phytoplankton varies largely among areas (Fig. 1.15, upper). While equatorial area is slightly more productive than neighboring subtropical areas, productions of tropical and subtropical areas are generally low. Especially, production of the area between latitude 30° north and south is markedly low. In contrast, there are high production areas in the subarctic and polar zones. This latitudinal distribution pattern is different compared with the production on land. This distribution pattern indicates that water

Table 1.4. Net production and biomass of primary producer in ocean (Whittaker, 1975). Weights are expressed as dry weight.

Category	Primary producer			
	Area (10 ⁶ km ²)	Production (P) (m ² /year)	Biomass (B) (kg/m ²)	P:B ratio (/year)
Open ocean	332.0	125	0.003	42
Upwelling zone	0.4	500	0.02	25
Continental shelf	22.6	360	0.01	36
Seaweed bed and coral reef	0.6	2,500	2	1.3
Bay	1.4	1,500	1	1.5
Whole sea area	361	152	0.02	15.2
Whole land area	149	773	12.3	0.063

temperature which regulates metabolic rates is not a major controlling factor of primary production. Let us compare the global distribution of primary production with that of solar light intensity and nutrient salts. Annual total solar radiation at the surface decrease from the equator to higher latitudes, and this distribution does not match the primary production in Fig. 1.14a. On the contrary, the distribution of nitrate at bottom of euphotic zone (80 m depth layer) that indicated upward supply of nitrate well agrees with the distribution of primary production. Thus we can understand that major factor controlling global primary production is the supply of nutrient salts to the euphotic zone.

From the fact that the annual primary production is similar in magnitude in the ocean and on land, can we regard that rate of material cycling, that is, turnover rate is similar in these environments? The answer is “no”. Primary production in itself does not indicate turnover rate. It depends on both production and biomass. Table 1.4 summarizes net primary productions and algal biomass in 5 marine ecosystems: open ocean, upwelling, continental shelf, sea grass bed and coral reef, and estuary. The ratio of production to biomass (P:B ratio) is an indicator of annual number of times of replacement of organic matter, that is, a number of generation. The mean P:B ratio in the ocean is 15.2, and is 241 times larger than that of lands. Especially, the rates are high in the open ocean, upwelling and continental shelf, where phytoplankton are the primary producers. This difference originates from a fundamental difference in primary producers between sea and land. Comparing the structure of a terrestrial plant, a large part of its structure such as stem, branch and root which serve as support of leaves against force of gravity and supply of water do not directly contribute to photosynthesis. In contrast, phytoplankton, floating in water, do not require such supporting parts, a major portion of cellular materials is involved in photosynthesis. This is the major reason for the big difference in P:B ratio. The fact that unicellular phytoplankton have far huge P:B ratio fits to a general rule in biology that species whose body size is small has shorter life span. The mean P:B ratio of aquatic

animals is also larger than that of terrestrial animals, reflecting the difference in primary producers. The high P:B ratio of planktonic ecosystem implies that the system produces large amount of organic matter from relatively small biomass. This high turnover capacity is a characteristic nature of the aquatic ecosystem.

FURTHER COMMENT

Twentieth century is the first time when human being recognized that aquatic environments are disturbed by human activities, as exemplified by pollution by harmful and toxic substances, negative impacts of eutrophication with which substances are not harmful and toxic in itself in normal concentration. We read newspaper articles on environmentally negative issues almost everyday. The issues used to be local, but recently many global issues become to be critical. Our environments and their ecosystems are facing critical points (Millennium Ecosystem Assessment; <http://www.millenniumassessment.org/en/index.aspx>). The most serious problem of these phenomena is that those degradation processes proceed in high speeds which human being and organisms on the Earth have not experienced. Can aquatic ecosystem adapt to such rapid changes?

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