CHAPTER 3

POLLUTION IN COASTAL WATERS AND ESTUARIES

As was mentioned in Chapter 1, one of the main consequences of the growth in population of coastal regions is the sharp increase of the volume of pollutants dumped into the sea. There is quite a variety of substances that can be classified as pollution, including some that only fall into this category when their concentration rises above a certain level.

In addition to the commonly known, and regulated, sewage discharges via marine outfalls or sewage sludge dumping, generally comprising waste products of both domestic and industrial origin, other substances such as the heated seawater used to cool down thermal plants, and the nutrients dragged into coastal waters by river discharges, can also be considered as pollutants, since their effects on the marine environment may agree with those listed in the definition given by Albaigés (1989) – see §1.1 -; cooling water discharges can raise the water temperature by as much as 5° C, significantly affecting marine life, and the nutrient contamination of rivers by fertiliser runoff from intensive farming may lead to a nutrient enrichment of the water column which, in turn, acts as a fertiliser and increases the growth of phytoplankton, leading to algal blooms and eutrophication (Falconer, 1992). Furthermore, and in contrast with those given above, other types of substances do not even have a definite source, but enter the waterbody over a large stretch of the coast, and are almost impossible to control; an example of these could be the sediment loads which enter the water flow from dredging operations.

The present chapter describes the main types of sources through which a pollutant substance may be introduced into coastal waters, presents a classification of pollutants based on their characteristics, and focuses on the four main types of substances whose transport can be relevant near the coast, i.e., microbiological organisms from wastewater discharges, nutrients from river discharges, heated seawater from thermal plant cooling processes, and sediment particles.

3.1 SOURCES OF POLLUTION

Without considering the exceptional cases of marine accidents and disasters, coastal pollution is almost always caused by land-based sources. According to their spatial extent, pollutant sources can be broadly classified into two categories: point sources, and non-point or diffuse sources.

3.1.1 Point sources

These sources release polluting substances at discrete and clearly identifiable locations, they can be measured or quantified, and their impact on the surrounding environment can be somehow evaluated. Under most circumstances, point sources are usually continuous, even though the discharged volume flux can be variable in time. Examples of point sources are effluents from municipal and industrial wastewater treatment plants, and effluents from farm buildings or solid waste deposition sites. In all cases, these discharges result in the release of bacteria and organic matter, nutrients, biological oxygen demand (BOD), and toxic substances.

The pollutants may reach the sea via rivers and estuaries, which carry waste from inland sources, or via direct discharges into the water from coastal sources using marine outfalls. Charlton (1985) classified marine outfalls into four categories, as depicted in figure 3.1:

a) Outfalls where both the pipe and the diffuser section are entirely above the sea bed (fig. 3.1a)

b) Outfalls in which the pipe is buried, and the diffuser section is composed of short risers ending at discharge ports just above the sea bed (fig. 3.1b).

c) Tunnelled outfalls, where the pipe is deeply buried, and in which the diffuser section consists of long risers ending at discharge ports. Charlton (1985) further divides this type of outfalls according to where the risers are connected to the main pipe -risers connected to the tunnel soffit (fig. 3.1c) or to the tunnel invert (fig. 3.1d)- since their behaviour under sea water intrusion is significantly different.

3.1.2 Diffuse sources

The distinguishing feature of non-point sources is that the discharge cannot be related to a specific well defined location; diffuse sources are also characterised by being usually transient, depending on the variable precipitation rate around the year, although some (e.g., atmospheric input) may be considered more or less continuous.

Diffuse sources release pollution originated from weathering processes of minerals, erosion of lands and forests, or from artificial sources related to human activity, such as the use of fertilisers and weed- and pest-control chemicals, the erosion of soils from agricultural farming areas, and erosion and transport in urban developments (Krenkel and Novotny, 1980). Other non-point sources include pollution due to leaching from land disposal of solid wastes, groundwater infiltration, and wet and dry atmospheric fallout.
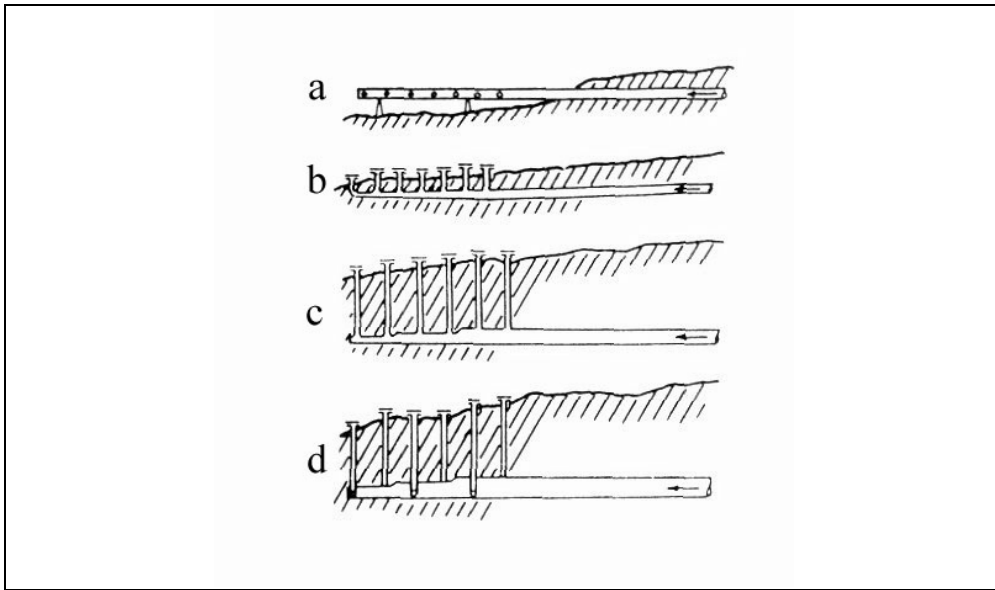
Figure 3.1: The four main types of outfall and diffuser configuration (from Charlton, 1985).

3.2 CLASSIFICATION OF POLLUTANTS

Pollutant substances discharged into the marine environments can be classified in two big groups (WHO, 1982):

a) substances, or groups of substances, that are toxic, durable and accumulative, for which action needs to be taken at the source (Group I)

b) substances, or groups of substances, that are not as toxic as the former (*de per se*, or due to natural dilution processes), and usually affect only limited coastal regions (Group II).

Figure 3.2, from WHO (1982), illustrates in a schematic manner the difference between pollution generated by Group I and Group II substances. In this figure, curve C corresponds to Group I pollutants, and shows a maximum negative effect near or at the discharge point (maximum concentration); the broken line C' is equivalent to C, but for the case in which the substance can accumulate (by sedimentation or bio-accumulation) at some distance from the discharge point. Curves A and B refer to Group II pollution: B is the general case, where the negative effects are greatest at the discharge point, but decrease with time (or distance, or dilution); A corresponds typically to nutrients and organic substances, with negative effects at the injection point, but also with possible positive effects if found in the correct amount in naturally deficient areas. Because Group I substances cannot be legally discharged into the sea, they will not be treated here, and only those belonging to Group II are considered hereafter.

Substances defining water quality are commonly categorised into four groups (Krenkel and Novotny, 1980): chemical, physical, biological and physiological pollutants. A brief description of each type is given below.
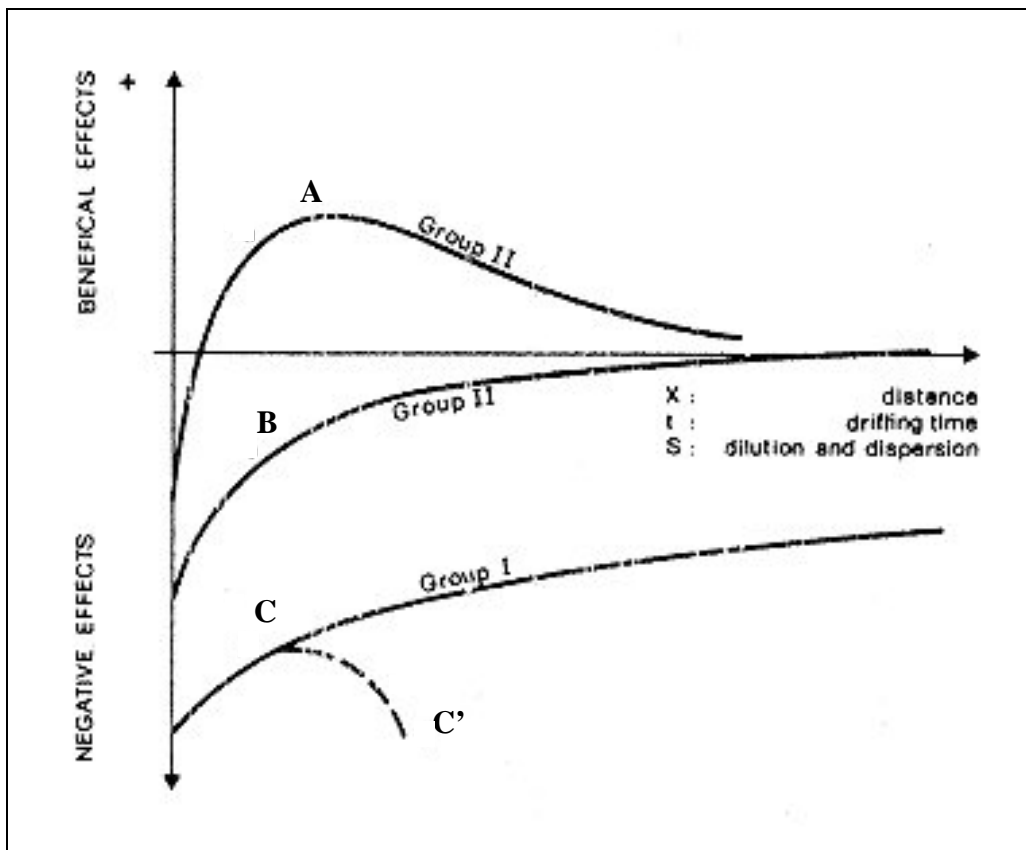
Figure 3.2: Schematic representation of phenomena of marine pollution, caused by the discharge of wastes containing substances of group I and II (WHO, 1982).

3.2.1 Chemical pollutants

Broadly classified into organic and inorganic, where the former are defined as compounds containing a carbon atom.

a) Inorganic pollutants, which include metals (zinc, copper, nickel, tin, titanium, lead, chromium, etc.), either resulting directly from industrial wastes or contained in certain biocides, and some inorganic compounds of phosphorus, that are generally the result of industrial waste discharges. The major effects of inorganic pollutants are changes in the pH of the water and severe toxicity if found in high concentrations; inert insoluble substances may also lead to sludge deposits on the bottom which can inhibit normal benthic biological activity.

b) Organic pollutants, including organic pesticides, crude oils and hydrocarbons, nutrient substances and organosilicon compounds may be included in this group.

Organic phosphorus is a component of a number of biocides, and may be released either as waste from the pesticide production process, or through washout after pesticide application. It can be dangerous due to the natural effect of the pesticide, or in small concentrations may lead to eutrophication, depleting oxygen from the water and lowering dissolved oxygen levels to values unfit for aquatic life. Pesticides other than organophosphoric may additionally cause problems because of accumulation due to their non-biodegradability.

Crude oils -resulting from palm, meat and fruit processing-, mineral oils and hydrocarbons tend to form surface films which may reduce the rate of oxygen transfer. Mineral oils from car washes and garages can be mixed with petrol, detergents and heavy metals (WHO, 1982).

Nutrient substances (phosphates, nitrates, etc.) and, in general, any biodegradable waste presents an effect on oxygen balance due to the immediate oxygen demand and biological oxygen demand (BOD). Even though small concentrations of nutrients are beneficial for marine life, large amounts may lead to a lack of oxygen and eutrophication of the waterbody.

3.2.2. Physical pollutants

These include colour, temperature, foam, suspended solids, turbidity and radioactivity, which may also be associated with chemical pollutants.

The increase of environmental temperature due to the discharge of water used for cooling purposes may affect the potability of water, apart from having significant effects on physical, biological and chemical processes, possibly unbalancing the local ecosystem. The temperature rise increases metabolic activity in the nearby flora and fauna, resulting in a higher BOD and local hypertrophication; on the other hand, it also decreases oxygen solubility in water. The toxic effects of cyanides are also magnified by the temperature rise.

Chemical agents causing foam may have adverse physiological effects on marine life. In addition, surface active agents may cause a reduction in the rate of oxygen gas transfer, reducing the capacity to assimilate waste.

The waste assimilation capacity may also be reduced by an excess of suspended solids in the water body; suspended matter can be organic or inorganic, and may also result in inhibition of photosynthesis and a retardation of benthic activity and production due to the reductions in sunlight intensity.

Finally, radioactive wastes, which proceed mainly from the release of radioactive materials into the sea, but also from the use of radioisotopes in certain industries and for research purposes, may provoke somatic and genetic effects in all living organisms. Radioactivity can be concentrated biologically by direct uptake from the water and passage through food webs, and physically by absorption, ion exchange, coprecipitation, flocculation, and sedimentation.

3.2.3 Biological pollutants

The effects of biological pollutants can be of two kinds: diseases, in which case the pollutant is called primary; and increases in biological growths, caused by so-called secondary biological pollutants.

a)   Primary biological pollutants are mainly domestic and industrial wastewaters that include microbiological agents, such as bacteria, virus, protozoans and helminths, involved in waterborne diseases; a further source of pollution is agricultural runoff. A more detailed description of microbiological pollution is given below (§3.3).

b)   Secondary biological pollutants are substances added to the water that may result in excessive growth of phytoplankton, macrophytes, and fungi, leading to eutrophication.

3.2.4 Physiological pollutants

Following Krenkel and Novotny (1980), substances included in this group are those which produce bad taste and odour in drinking water (e.g., chlorophenols or hydrogen sulphide), yet not necessarily cause adverse physiological effects on either fish or man.

3.3 MICROBIOLOGICAL POLLUTION

One of the consequences of wastewater discharges in the sea is the introduction in this environment of high levels of micro-organisms which may cause communicable diseases. The modes of transmission of pathogens are through ingestion of contaminated water and food, and by exposure to infected water, persons or animals; thus, concentration levels of these organisms are regulated by water quality standards for health-related water use activities, such as bathing, boating, fishing, or water-supply (domestic, municipal and industrial).

The most direct approach to estimate faecal pollution in a waterbody is to quantify the pathogen agents in it. However, this procedure is technically complex and unpractical for routine surveys due to the large number of potentially dangerous micro-organisms that can be found (several types of viruses and bacteria, different protozoa, etc.).

An alternative approach consists in focusing on just a few organisms, called bacteriological indicators, that can be used to detect faecal pollution, independently of their own pathogenicity.

3.3.1 Micro-organism types, effects on human health, and diseases

The most important microbiological organisms found in water, from the point of view of human use and consumption, are those capable of infecting, or transmitting diseases to, humans. These organisms are not native to aquatic systems but, nevertheless, are able to adapt to the hostile environment, and maintain their infectious capabilities for significant periods of time. The main pathogens discharged by sewage effluents include species of bacteria, viruses, protozoa, and parasitic worms:

a)   Bacteria:

They are the most numerous micro-organisms, entering the waterbody through municipal sewage discharge, discharges of sludge, discharge of industrial wastes (dairy products, fruit processing, textile, leather) and animal wastes. Surface water quality is usually evaluated by considering the concentration of non-pathogenic bacteria, such as faecal bacteria, which are used as indicators of the presence of pathogenic micro-organisms (§3.3.2). These are responsible for a number of waterborne diseases, as can be seen in table 3.1.

b)   Viruses:

The sources of viruses include human and animal wastes, together with some industrial wastes. The survival of these micro-organisms in waterbodies has not been studied as much as that for bacteria, but a series of factors seem to indicate that survival rates for viruses is

quite different to bacterial rates. As parasites, viruses cannot reproduce in the absence of living cells, so their survival will depend on the ability to find and infect one; on the other hand, viruses are much more resistant than bacteria to methods commonly used in water treatment plants, and they present a great mutagenic capacity which allows them a relatively easy adaptation to adverse environmental conditions. Table 3.1 also lists some of the diseases transmitted by viruses.

Micro-organism	Disease	Micro-organism	Disease
Vibrio cholerae	Cholera	*Mycobacterium tuberculosis*	Tuberculosis
Salmonella sp.	Salmonellosis	*Bacilus anthracis*	Anthrax
Salmonella typhi	Typhoid fever	*Schistosoma* spp.	Schistosomiasis
Norwalk types	Gastroenteritis	*Shigella* spp.	Bacillary disentery
Hepatitis A virus	Hepatitis A	*Capillaria hepatica*	Hepatic capillariasis
Giardia lamblia	Giardiasis	*Poliovirus*	Poliomielitis

Table 3.1: Examples of diseases transmitted by micro-organisms discharged into waterbodies.

c)   Protozoa:

These are unicellular organisms, more complex than bacteria or viruses in their functional activity. They are complete, self-contained organisms that can be parasite or free-living, pathogenic or non-pathogenic. Under adverse environmental conditions they may form cysts, which must be removed by complete water treatment, since disinfection alone is not sufficient.

d)   Helminths:

Water pollution may result from parasitic worms contained in human or animal wastes, as well as *via* aquatic species of other hosts, such as snails and insects. Modern water treatment methods are very effective in destroying these organisms, so they suppose danger only in untreated waters. Some of the infections transmitted by tape, hook and round worms are also given in table 3.1.

3.3.2 Legislation and standards for water quality. Biological indicators

In order to conserve the beneficial uses of the ocean, each country defines a set of water quality standards and the corresponding legislation to maintain them, although these may vary from country to country, and even between different regions of the same country. The goal of these standards is to safeguard the overall marine ecosystem, the use of waters for bathing and recreational purposes, and shellfish collection.

Different standards are defined for different purposes, and are generally based on the presence of indicator organisms, such as certain types of bacteria. Water quality standards for regions where shellfish are collected are different from standards used for bathing and water contact sports.

The Spanish legislation establishes, within the Real Decreto 734/1988 ("Normas de calidad de las aguas de baño"), a set of microbiological concentration values for bathing waters (table 3.2) and the criteria under which these values must be interpreted: 95% of the samples must not exceed the mandatory values, 90% of the samples must not exceed the guideline values, except those referring to

total and faecal coliforms, and 80% of the samples must not surpass the guideline values given for total and faecal coliforms.

On the other hand, different regulations are set for regions where shellfish are collected. The reason is because these organisms are filter feeders that bio-accumulate waterborne micro-organisms, and the levels of coliform concentration in shellfish can be much higher than in the surrounding waters. In most countries, in addition to water quality tests, samples of shellfish flesh and intervalvular fluid are also analysed. As an example, the Real Decreto 345/1993, which determines the quality standards for shellfish production in Spain, states a series of conditions to be met by shellfish before being put on the market, such as an allowable upper limit of 300 faecal coliforms per 100 gr of flesh and intervalvular fluid, or the absence of salmonella in 25 gr of flesh, amongst others.

Organism	Guideline Value	Mandatory Value
Total coliforms (per 100 ml)	500	10,000
Faecal coliforms (per 100 ml)	100	2,000
Faecal streptococci (per 100 ml)	100	-
Salmonella (per litre)	-	0
Enteroviruses (PFU per 10 l)	-	0

Table 3.2: Spanish bathing water directive. Microbial standards for bathing waters (from Buceta, 1995). PFU stands for "plaque forming units".

Indicator organisms are, generally speaking, those whose presence presumes that contamination has occurred, and suggests the nature and extent of the contaminants (Peavy *et al.*, 1985). According to these authors, the ideal pathogen indicator should *a)* be applicable to all types of water, *b)* always be present when pathogens are present, *c)* always be absent when pathogens are absent, *d)* lend itself to simple and reliable routine quantitative tests, and *e)* not be a pathogen itself, for the safety of laboratory personnel.

Since most of the waterborne pathogens are introduced through faecal contamination, organisms native to the intestinal tracts of humans and warm-blooded animals should be good indicators if they meet the above criteria. Three different groups of faecal bacteria are commonly used to estimate the degree of pathogen pollution in waterbodies:

a)   The total coliform (TC) bacteria group, which comprises "all of the aerobic and facultative anaerobic, gram-negative, non-spore-forming, rod-shaped bacteria which ferment lactose with gas formation within 48 hr at 35°C" (Krenkel and Novotny, 1980). These bacteria have been isolated from both polluted and non-polluted soil samples, as well as from faeces of humans and other warm-blooded animals, and from the gut of cold-blooded animals; the typical example is *Escherichia coli*. This group was widely used in the past, because it was observed that their survival rates and those of pathogenic enteric bacteria were of the same order of magnitude under similar environmental conditions, but has been gradually replaced by tests based on other indicators, due to the high number of coliform sources.

b)   The faecal coliform (FC) bacteria are indicative of organisms from the intestinal tract of humans and other warm-blooded animals (e.g., *Klebsiella*), and can hardly be associated with soils, plants, or insects. Faecal coliforms behave similarly to common enteric pathogens such as *Shigella* and *Salmonella*, and a general correlation has been found between the concentration of FC bacteria and the presence of *Salmonella* (Geldreich, 1978).

c) The faecal streptococci (FS) group includes different varieties of streptococci whose normal habitat is the intestines of humans and warm-blooded animals. However, there is a considerable variation in types and numbers of these bacteria, so it may not be a reliable indicator. Some members of this group can also be found in soils and faeces of freshwater fish.

Other indicators have been proposed to evaluate the quality of surface waters. Because of the interest in waterborne diseases caused by viruses, and the fact that these organisms are more resistant to water treatment than bacteria, it has been suggested that these may be used as indicators of pathogenic contamination. Coliform bacteria indicators, however, remain superior to virus indicators because of the greater complexity and lesser accuracy of virus counting methods, and the small density of enteric viruses in surface waters (typical ratios of coliform bacteria to viruses are of the order 75,000:1).

Suggested alternative faecal indicators include other types of bacteria (*Pseudomonas aeruginosa*, *Staphilococcus aureus*, etc), faecal sterols such as coprostanol, which is present in human faeces, and bacteriophages (i.e., viruses which infect specific bacteria cells) and enteroviruses (Wood *et al.*, 1993). A detailed review of alternative faecal indicators can be found in Havelaar (1991), for instance.

The ratio FC/FS is a good indicator of whether the sources of bacteria are human or from other warm-blooded animals. In the former case (i.e., sewage), values of FC/FS have been reported to be between 4.0 and 27.9, whereas in the case of surface runoff they have been found to lay between 0.04 and 0.26 (Geldreich *et al.*, 1968).

3.3.3 Organism decay rate

The survival, fate and distribution of bacteria and other organisms in natural waters depend on the particular type of waterbody and associated phenomena that influence the growth, death and other losses of the organisms. Following Thomann and Mueller (1987), these factors are

- Sunlight
- Temperature
- Salinity
- Nutrient deficiencies
- Toxic substances
- Predation
- Settling of the organism population after discharge
- Resuspension of particulates with associated sorbed organisms
- The growth of organisms in the water body (aftergrowth)

and each one may be present in varying degrees depending on the specific case. Additionally, organism concentration will depend on the dilution (mainly initial dilution and dispersion) of the discharge.

The first five environmental factors listed above are included in the general concept of mortality, which refers to the reduction in organism population due to the adverse environmental conditions; predation by other organisms also leads to the same result. On the other hand, the sedimentation and resuspension of organisms and physical processes such as dispersion merely redistribute the existing population, with no reduction in its number. It has been shown that the dominant cause of bacteria inactivation is mortality, due mainly to solar radiation (Wood *et al*., 1993).

3.3.3.1 Mathematical expression for bacteria mortality

Laboratory experiments show that populations of bacteria exhibit a number of mortality patterns -fig. 3.3-, which can be classified as (Mancini, 1978):

a)   First order mortality,

b)   Initial increase of population, followed by first order mortality,

c)   Patterns a) or b), followed by a reduced rate of mortality as time increases.

It is usually accepted that bacteria mortality corresponds to a first-order kinetic process, which can be expressed in the form of an exponential law:

$$\frac{dN}{dt} = -K_d N \tag{3.1}$$

from where the bacteria concentration N_t at time t will be

$$N_t = N_o e^{-K_d t} \tag{3.2}$$

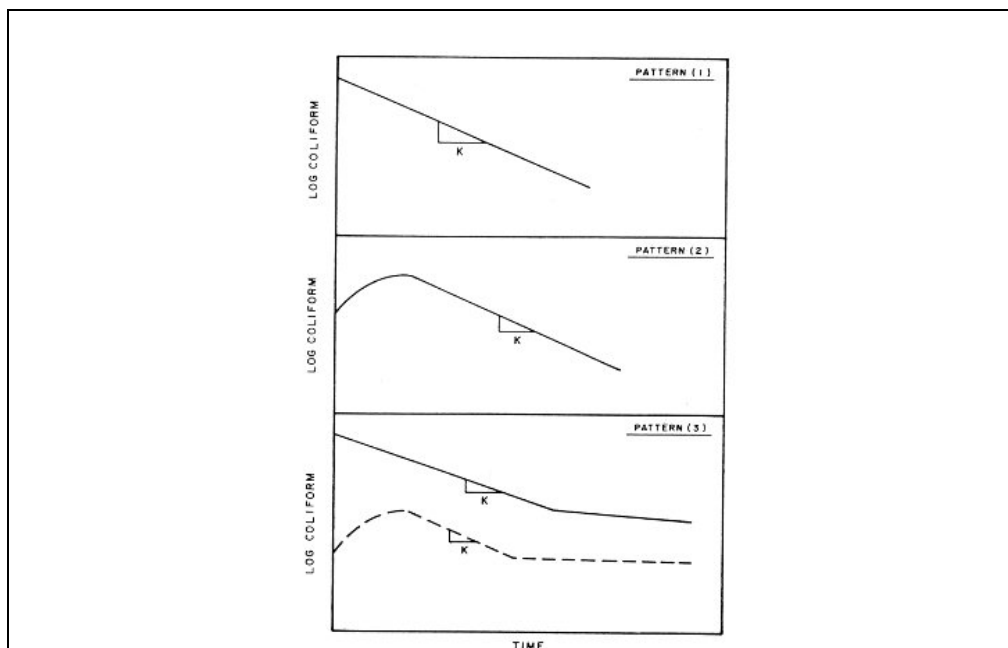where K_d is the mortality rate or decay rate and N_0 is the initial indicator concentration.

Figure 3.3: Different bacteria mortality patterns, from Mancini (1978). The slope of the linear section corresponds to the decay coefficient.

An equivalent parameter used to express the overall decay rate considers the time it takes to obtain a 90% mortality of the original number of bacteria, T_{90}. It is easy to see from equation (3.2) that the relationship between K_d and T_{90} is

$$K_d = \frac{2.3}{T_{90}}$$
(3.3)

As an alternative to the exponential inactivation pattern, Bravo (1985) presented a hyperbolic decay model that agreed better with the data he had obtained off the Spanish Mediterranean coast. The expression he proposed

$$\log N_t = \frac{K_d{}'}{K_d{}'+t} \log N_o$$
(3.4)

where $K_d{}'$ is the hyperbolic decay rate.

3.3.3.2 Estimation of the decay rate

The decay rate varies enormously as a function of the organism type and the environmental conditions; as an example, viruses generally die away at a slower rate than bacteria by about an order of magnitude. Table 3.3 gives some experimental mortality rates for bacteria and viruses, as reported by different authors (from Thomann and Mueller, 1987), from laboratory or field tests. The advantage of laboratory experiments lies in the fact that the influence of each individual factor can be quantified separately, whereas in field experiments this is not possible. On the other hand, data obtained in the laboratory are not always valid when used in the natural environment.

	K_d (day^{-1})	Remarks
Total coliform	0.7 - 84	Both in fresh and seawater, for different temperatures.
Faecal coliform	37 - 110	Sunlighted seawater
Total or faecal coliform	0 - 6.1	Different salinity waters; lower values for dark samples, higher values for sunlighted samples.
Faecal streptococci	18 - 55	Sunlighted seawater
Pathogens (*Salmonella thompson*)	0.5 - 3	Lake Ontario, 18°C
Viruses (Polio type I)	0.05 - 0.26	Marine waters, 4°C - 25°C
Viruses (Enteric)	0.15 - 2.3	Different environment, 0°C - 24°C

Table 3.3: : Some reported decay rates for bacteria and viruses (adapted from Thomann and Mueller, 1987).

Several authors have used this kind of data to derive expressions which yield the value of K_d as a function of different environmental parameters, such as the temperature and the salinity. As far as the author knows, there is no equation for the hyperbolic decay parameter $K_d{}'$, and the use of (3.4) must rely on tabulated experimental values.

In order to fit an algebraic expression to the obtained data, the decay rate K_d can be expressed as

$$K_d = K_1 + K_2 + K_3 - K_4$$
(3.5)

where $K_1 = K_T{}'K_S$ is the term corresponding to the effects of temperature and salinity, K_2 comprises those due to solar radiation, K_3 corresponds to the organism sedimentation minus the resuspension, and K_4 is a

growth coefficient. The latter can be neglected if it is assumed that the nutrient concentration in the sea is usually insufficient for the growth of these organisms.

a) Mancini's (1978) expression; extension by Thomman and Mueller (1987)

Mancini (1978) analysed approximately 100 coliform mortality rates found in the literature, measured in both laboratory and *in situ* experiments, and derived expressions for K_1 and K_2 as functions of temperature T, percentage of sea water (Λ) and average daily surface solar radiation, $\bar{I}_o$:

$$K_1 = 1.07^{(T-20)}[0.8 + 0.006\Lambda]$$
$$K_2 = \frac{\alpha \bar{I}_o}{\gamma_{ex} H}[1 - e^{-\gamma_{ex}H}]$$

(3.6)

where γ_{ex} is the light extinction coefficient, α is a proportionality coefficient, and H an averaging depth (completely mixed layer); the percentage of sea water has been used instead of the salinity because it appears that components in the sea water other than salt could be responsible for part of the coliform mortality. Mancini recognised that equation (3.6) should be used with care, due to the wide scatter observed in the reported mortality rates under similar conditions, to the variety of mortality patterns, and to the absence of definitive data on the influence of sunlight in fresh water systems.

Thomman and Mueller (1987) adopted the above equations for K_1 and K_2, and they introduced the term K_3, parameterised as follows:

$$K_3 = \frac{\upsilon_S}{H}$$

(3.7)

where υ_S is the net loss rate (in m/day) of the particulate bacterial form. However, the estimation of this coefficient is very complex, and is usually ignored.

b) Juanes (1995) equations

Juanes (1995) proposed the following expressions for K_1 and K_2, obtained from laboratory experimental data:

$$K_1 = \alpha 1.040^{(T-20)} 1.012^S$$
$$K_2 = 0.113 I_o(t) e^{-\gamma_{ex}z}$$

(3.8)

where T and γ_{ex} are the same than in the previous expressions, S is the salinity, z is the depth, $I_o(t)$ is the solar surface radiation and $\alpha = 2.553$ is a "darkness" decay rate. After comparing modelled results with data obtained in field tests, Juanes found that this model for K_d worked better in summer than in the winter, and that it showed a greater variation with z than the observed data, probably due to a defective computation of the sunlight extinction in the watercolumn.

c) Expression used in Spanish legal outfall regulations

The Spanish "Instrucción para el proyecto de conducciones de vertidos desde tierra al mar" (1993), which regulates all matters regarding the characteristics and previous treatment of effluents, the projects and engineering of marine outfalls, and sea water quality in general, uses the following equation (from data obtained by Gould and Munro, 1981) to estimate the faecal coliform decay rate in waters with a salinity higher than 30 g/l:

$$T_{90} = \left[\frac{\xi_\alpha}{60} \left(1 - 0.65 C_\%^2 \right) \left(1 - \frac{SS}{800} \right) + 0.02 \cdot 10^{(T-20)/35} \right]^{-1}$$ (3.9)

where T_{90} is obtained in hours if ξ_α is the sun's altitude in degrees, SS is the concentration of solids in suspension, in mg/l, and T is the water temperature in Celsius degrees, and where $C_\%$ is the fraction of cloud-covered sky.

Of all the factors appearing in eqs. (3.6) to (3.9), both the temperature T and the salinity S can be measured directly in the waterbody, and over the whole depth of the water column. On the other hand, the intensity of the solar radiation is not as easy to evaluate and, while T and S may be considered fairly stationary over small time intervals, I_o is highly dependent on a number of parameters which vary significantly throughout the day and seasonally, such as the weather conditions and the solar elevation. Furthermore, micro-organism mortality is not uniform over the whole radiation spectrum, but is clearly wavelength-dependent. In the following section, an approximate estimation of solar radiation intensity at the sea surface, and its decay with underwater depth, is described.

3.3.3.3 Determination of the solar radiation

At any time, the net energy available to the waterbody due to solar radiation is

$$I_o = \Phi_{S0}(1 - A_S) + \Phi_{L0}(1 - A_L) + \Phi_B$$ (3.10)

where Φ_{S0}, Φ_{L0} and Φ_B are the incident shortwave and longwave radiation and the longwave radiative loss, respectively, and A_S, A_L are the shortwave and longwave albedoes or reflectivities.

The shortwave radiative flux (approximately in the range 300-2,500 nm) is partially absorbed in the uppermost 1-2 mm of the waterbody; the absorbed portion is related to the energy associated with wavelengths at the red end of the spectrum, and it also depends on water turbidity (Williams *et al.*, 1981). The remaining fraction (around 50%-60% of the total energy) may penetrate to a greater depth, which depends on the turbidity caused by suspended organic and inorganic material; the effect of turbidity is usually quantified as an extinction coefficient γ. On the other hand, the longwave radiation is completely absorbed in the surface layer (McAlister and McLeish, 1969).

Previous studies on the effects of solar radiation on bacteria mortality have shown that the lethal effects of wavelengths greater than 500 nm are very small (Gameson and Gould, 1974), while Calkins and Barcelo (1982) specify that, even though the energy contained between 280-320 nm is relatively low, it is this band that presents the most bactericidal effects at sea level. Therefore, longwave radiation will be neglected in what follows.

a) Shortwave radiation

The evaluation of the solar radiation is difficult since it depends upon atmospheric scatter, reflection back to space and molecular absorption, even under cloudless conditions. Available formulae attempt to relate the global solar radiation at ground level Φ_{S0}, either to the global shortwave radiation under cloudless skies Φ_{S0}^*, or to the radiation at the top of the atmosphere Φ_∞. Their applicability is in

general geographically restricted, because they are usually derived from data measured at a single station.

The magnitude of the energy reaching sea level is essentially a function of the sun's elevation (latitude, day of the year and time of the day) and of the cloudiness. Numerous experimental results have been presented in the form of tables or maps, and used as a basis to derive empirical expressions (see Ivanoff, 1977, for a review).

One of the most frequently used equations relating solar radiation at ground level to solar elevation ξ_*, in the absence of clouds, is (Holstlag and Van Ulden, 1983),

$$\Phi_{S0}^* = a_1 \sin\xi_* + a_2 \tag{3.11}$$

where a_1 and a_2 are the turbidity coefficients resulting from the total attenuation of extraterrestrial radiation through the atmosphere (absorption and scattering). These coefficients must be estimated from experimental measurements.

The solar altitude is determined from astronomical formulae as

$$\sin\xi_* = \sin\phi_G \sin\delta_* + \cos\phi_G \cos\delta_* \cos H_*$$

$$\delta_* = 0.4093\sin\left(2\pi\frac{J - 79.75}{365}\right) \tag{3.12}$$

$$H_* = \pi - \frac{\pi}{12}\left(t - EC - \frac{\lambda_G}{15}\right)$$

where ϕ_G, λ_G are the latitude and longitude of the location, respectively, δ_* is the sun's declination, J is the Julian day, t is the time of the day, H_* is the corresponding sun's hour angle and EC is the number of hours added to the civilian time due to economical criteria.

When the sky is partially or totally covered by clouds, the usual approach is to correct the clear-sky radiation by a certain factor which depends on the cloud cover. Kasten and Czepelak (1980) proposed the following expression:

$$\Phi_{S0} = \Phi_{S0}^*(1 + c_1 C_\%^{c_2}) \tag{3.13}$$

where c_1 and c_2 are the cloud coefficients, strongly dependent on the climate of the concerned area. The type of cloud also plays an important role in the attenuation of irradiance, as can be seen in table 3.4 after Haurwitz (1948). Lumb (1964) defined up to nine different sky categories -table 3.5-, and found an expression for the irradiance at sea level involving category-dependent coefficients. His results for the North Atlantic Ocean (for $45^\circ < \theta < 65^\circ$) are shown in figure 3.4, taken from Ivanoff (1977).

CLOUD TYPE	Φ_S / Φ_S^*
Cirrostratus and Cirrocumulus	0.65 - 0.85
Altocumulus and Altostratus	0.41 - 0.52
Stratocumulus and Stratus	0.24 - 0.35
Thick Fog	0.17 - 0.19

Table 3.4: Effect of cloud type on the attenuation of irradiance during overcast weather, from Haurwitz (1948).

Values for the turbidity coefficients a_1 and a_2 were calculated by Colomer and Casamitjana (1994) at Lake Banyoles (42° 07' N, 2° 45', 175 m above sea level), not far from Barcelona (41°25'N, 2°10'E), from measurements obtained during two years. They found that experimental data were closely reproduced by equation (3.11) when the coefficients a_1=1118 Wm^{-2} and a_2= -84.75 Wm^{-2} were used. These values are valid for solar radiation under cloud cover $C_\%$ < 0.1, and for solar elevation greater than 5°.

CATEGORY	DESCRIPTION
1	Virtually clear sky, less than 2/8 coverage.
2	Well broken low clouds with little or no medium or high cloud (3/8 - 5/8).
3	6/8 to 8/8 of cirrus (not cirrostratus).
4	Thin layers of altocumulus (6/8 - 8/8).
5	Veil of cirrostratus over whole sky with up to 4/8.
6	7/8 or 8/8 of stratocumulus without rain, with or without some cumulus and little or no medium clouds.
7	Thick altostratus (6/8 - 8/8), with or without layers of stratocumulus beneath some rain.
8	Thick layers of stratus and stratocumulus, overcast including drizzle.
9	Thick layers of nimbostratus, overcast, also include medium clouds, and rain.

Table 3.5: Description of the nine sky conditions determined from observations in the North Atlantic (Lumb, 1964).
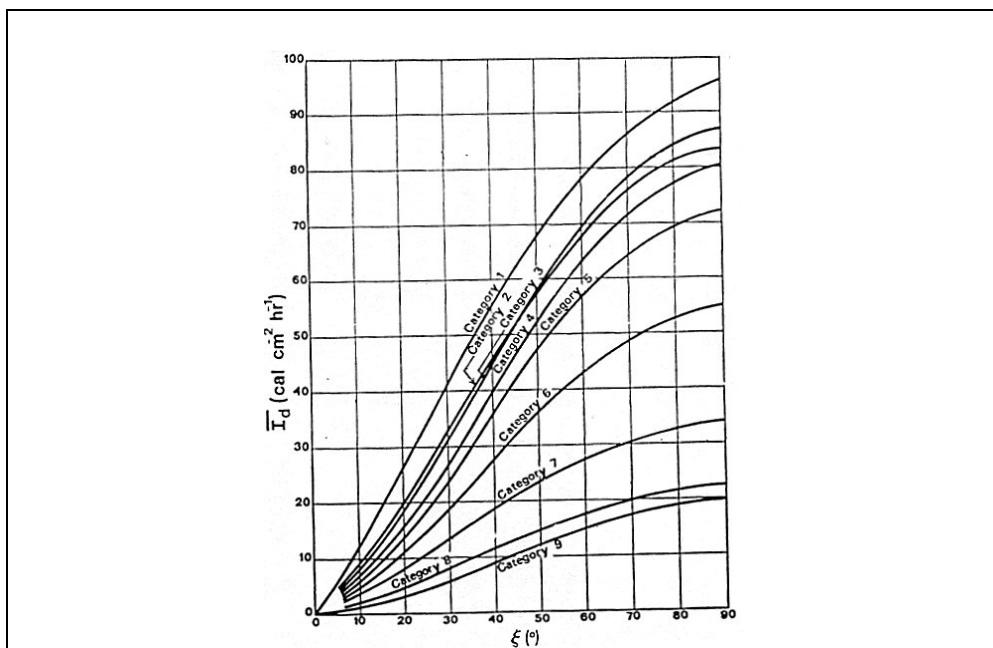

Figure 3.4: Mean solar irradiance at the sea surface (North Atlantic, 45°<θ<65°) as a function of the mean solar elevation, for the nine sky conditions defined in table 3.5 (Ivanoff, 1977).

The same authors compared the measured values of solar radiation under cloudy skies with those given by equation (3.11) to determine the cloud coefficients for equation (3.13). They defined three ranges for $C_\%$, and found c_1 and c_2 for each one. Table 3.6 shows these coefficients for each case, together with their mean value, and the error introduced in calculations when using the mean instead of the proper values.

It can be seen that, by using the mean values of the coefficients, the relative error is small except in the case of heavily clouded sky.

Cloud Cover	c_1	c_2	Error (%)
0.0< <0.1	0.00	0.00	0.3
0.1< <0.3	-0.66	0.93	1.4
0.3< <0.8	-0.76	1.01	5.4
0.8< <1.0	-0.78	1.20	16.0
All $C_\%$ (mean)	-0.71	1.03	0.0

Table 3.6: Values of the cloud coefficients estimated by Colomer and Casamitjana (1994), for Lake Banyoles.

b) Shortwave albedo

The amount of solar radiation that penetrates beneath the water surface is determined by its albedo or reflectivity, that is in general a function of the solar altitude, the state of the surface and the state of the water. However, this latter variable, which takes into account the presence of suspended sediment and organic material, is usually accounted for by an attenuation coefficient.

An algebraic expression for the reflectivity as a function of time of day can be found in Pivovarov (1972), assuming that the surface is reasonably flat:

$$A_S(t) = \frac{a_o}{a_o + \sin\xi_\boxtimes}$$

(3.14)

$$a_o = 0.02 + 0.01\left(0.5 - C_\%\right)\left(1 - \sin\left[\pi\left(\frac{J-81}{183}\right)\right]\right)$$

where $\xi_\boxtimes$, $C_\%$ and J have been defined before.

Even though this equation suggests that the albedo increases rapidly at low solar elevations, analyses by Grishchenko (1959) show that values approaching 1.0 are incorrect, and that for solar altitudes smaller than about 10° the albedo decreases rapidly. Additionally, for low values of $\xi_\boxtimes$, the state of the sea becomes important, as can be seen in figure (3.5) -Calathas, 1970-; in this case, the angle of incidence of the solar rays varies with the slope of the waves, and shadow zones and multiple reflections may appear. Furthermore, the albedo is altered by multiple reflections between cloud cover and water surface, as indicated by equation (3.14).

c) Absorption of solar energy by the oceans

The absorption of solar energy by sea water consists mainly of a conversion from radiant energy to heat; the remaining part (about 0.1%) of the absorbed radiant energy is involved in chemical processes such as photosynthesis. Absorption is due to the water itself, to the dissolved salts, to the organic substances in solution and to the suspended matter.

The latter also plays a decisive role in the process of scattering, that is, changing the direction of photons without energy loss; however, since this process increases the photon pathlength, scattering finally leads to an increased absorption and an additional energy loss. In any case, scattering becomes negligible in comparison with absorption at wavelengths greater than 700-800 nm.

The description of the penetration of sunlight into the oceans and its progressive attenuation with depth follows a more or less exponential form, depending upon the wavelength under consideration. As an illustration of this fact, figure 3.6 shows the variation with depth of the downwelling irradiance, as a fraction of the surface value, in the Northwestern Mediterranean Sea (42°14' N, 5°35'E). The relationship between attenuation and depth clearly depends also on the turbidity of the water. In the top hundred meters the attenuation coefficient can vary significantly during the course of the year, mainly because of the changes in photosynthetic production. Furthermore, it is generally assumed that the downward irradiance coefficient is independent of the state of the sky, since the effects of the zenith angle and cloudiness are not fully understood (Ivanoff, 1977).
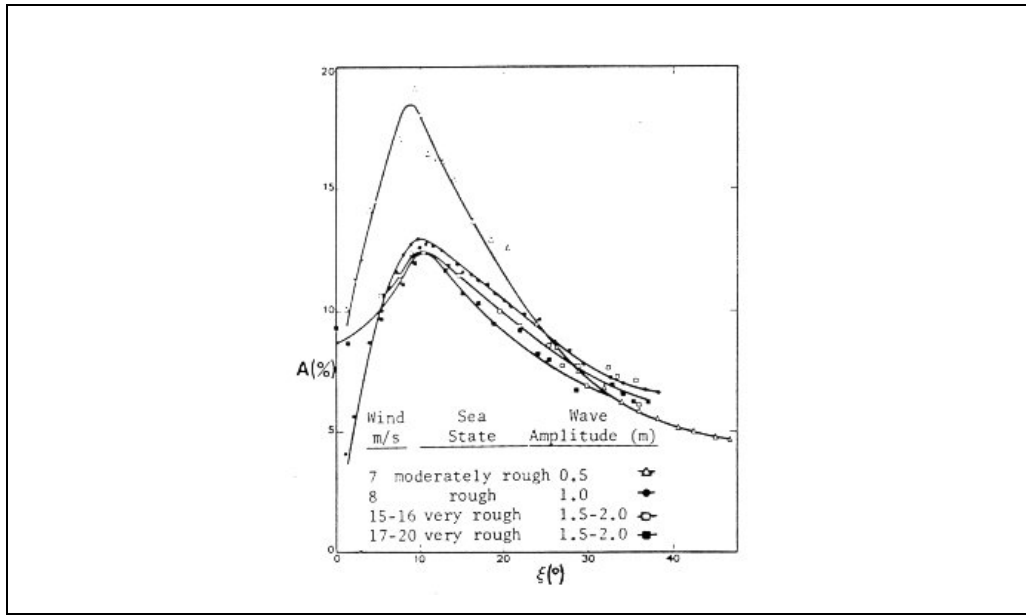

Figure 3.5: Variation of the albedo of the sea with the sun's elevation and roughness of the sea, for clear weather, in the Mediterranean Sea (42° 14' N, 5°35' E), from Calathas (1970).

By using the principle of energy conservation, it can be easily shown that, to a good approximation,

$$\frac{d\Phi_S(z)}{dz} = -\gamma_{ex}(z)\Phi_{S0} \tag{3.15}$$

where $\gamma_{ex}(z)$ is the downward irradiance attenuation coefficient at the depth under consideration. The distribution of shortwave radiation with depth has been parameterised in two different manners. The first is (Henderson-Sellers, 1986)

$$\Phi_S(z) = \Phi_{S0}(1 - A_S)\exp(-z\gamma_{ex,\lambda}(z)) \tag{3.16}$$

while the second is due to Zaneveld and Spinrad (1980):

$$\Phi_S(z) = \Phi_{S0}(1 - A_S)\exp(-\gamma_{ex,\lambda}z)\left[1 - k_1\tan^{-1}(k_2z)\right] \tag{3.17}$$

where $\gamma_{ex,\lambda}$ is now a wavelength-dependent attenuation coefficient and $k_1=0.231$ and $k_2=4$ are empirical constants.

The main problem arising in the calculation of energy absorption by the seawater is the determination of the attenuation coefficient, since it is a wavelength-dependent and extremely local variable. Gameson and Gould (1974) suggested using the extinction coefficient corresponding to the visible range of the spectrum because this is the interval where the mortality effects are greater.
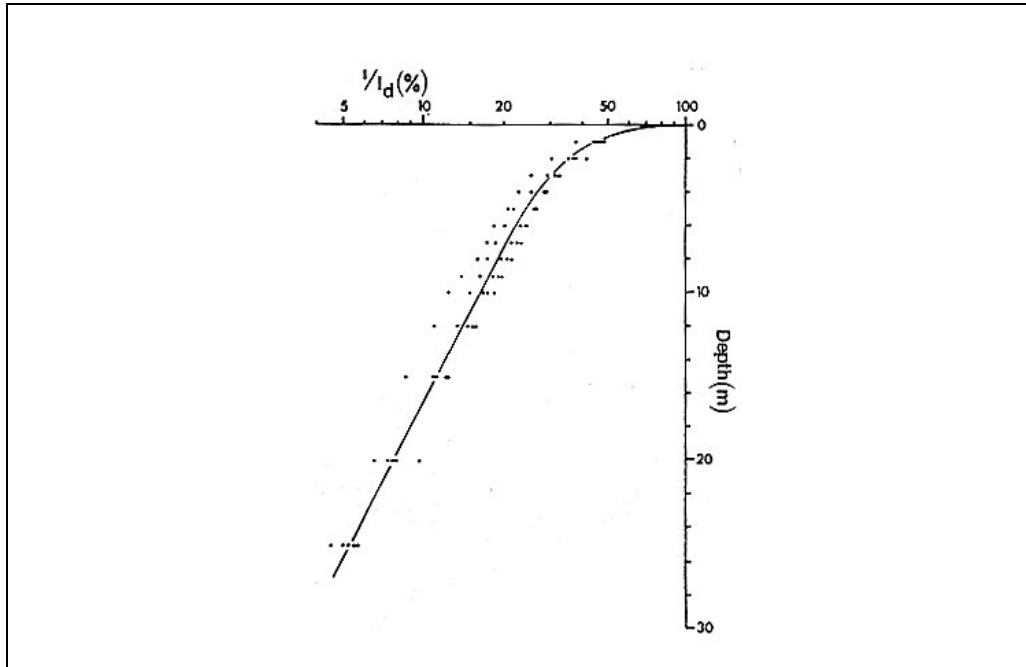

Figure 3.6: Variation with depth of the downwelling irradiance, as a fraction of the surface value, in the Mediterranean Sea (42°14' N, 5°35'E); the curve represents the average of five series of measurements (Ivanoff, 1977).

3.4 NUTRIENTS

The most important constraint on biological production in a waterbody comes from the limited concentration of available nutrients, since the sunlight requirements are satisfied by the penetration of solar rays up to a certain depth, and the oxygen supply is guaranteed by the atmosphere acting as a reservoir. The photosynthetic production of organic matter by phytoplankton in the surface layers of the sea is possible due to the assimilation of organic and inorganic nutrients from the surrounding water. Most of these substances are present at concentrations greatly in excess of the plant's needs, but others, like nitrogen and phosphorus, occur at very low concentration levels, and might be used almost to the point of exhaustion by algae. It is the availability of these nutrients that most frequently controls and limits the rate of organic production in the sea.

The current view in the fields of limnology and oceanology is that the maximum production attainable in pelagic communities at each trophic level is largely determined by the least available nutrient, amongst which nitrogen, phosphorus and silicon are the most outstanding. In contrast with lakes, where production limitations are imposed by phosphorus, the limiting role in marine waters has commonly been attributed to nitrogen, although a rigorous demonstration of the extent of N limitation has not yet been achieved in marine ecosystems (Smith, 1984; Hecky and Kilham, 1988). The critical role of N is suggested by the fact that, in oceans, the flux from river inputs and from aerobic N fixing

is only sufficient to satisfy about 1% of the total photosynthetic flux which is, therefore, very much dependent on *in situ* recycling. Nevertheless, the possible limiting effects of phosphorus in marine waters have also been suggested, based on the elemental composition of particulate oceanic material (Capblanq, 1990), but Ryther and Dunstan (1971) pointed out that about twice the amount of P as can be used by algae is normally present in coastal marine waters, so it is not likely that phosphates do play a relevant role in limiting marine production.

Within the marine environment, differences also arise between coastal and open sea ecosystems, mainly referring to the distribution and variability of nutrient concentrations, because in deeper waters the nutrient remineralisation takes place within the water column, while in shallower coastal regions this process occurs mainly in the sediment layer on the sea floor (Nixon, 1981). This, together with the direct proportionality between water temperature and benthic regeneration rates, leads to differences in the seasonal variation of nutrient concentrations in coastal and open sea waters. In coastal ecosystems, given the right hydrodynamic conditions for vertical mixing in the water column, maximum levels of nutrient concentrations will occur in summer; this may not be true, however, when the waterbody is stratified.

3.4.1 Nutrient Substances

Several chemical elements (at least 19) have been found to be essential for the development of life; the set of these elements, all of which can be found dissolved in water, receive the generic name of nutrients. A formal definition of nutrient elements was given by Parsons (1975), as those which are functionally involved in the processes of living organisms, such as hydrogen, carbon, nitrogen, or phosphorus, amongst others.

The term "nutrients", however, has been applied in oceanology almost exclusively to the chemical elements that limit the growth of aquatic population. Five elements (carbon, hydrogen, oxygen, nitrogen, and phosphorus) are required in large amounts, in molar proportions of 106:181:45:16:1, but the first three are found in abundance in water or dissolved carbon dioxide, and are never long-term limiting factors for aquatic plant growth (Stefan, 1994). On the other hand, the concentration of nitrogen and phosphorus dissolved in natural waters is much lower, and it is usually one of these two elements that provides the limiting factor for growth. Therefore, the word "nutrients", when used in marine sciences, refers mainly to nitrogen and phosphorus and, occasionally, also to silicon (figure 3.7). The first two, usually in the form of nitrates and phosphates, are converted by phytoplankton and higher-order plants into organic plant material through the photosynthesis mechanism, and end up forming the soft tissues of organisms. Silicon, on the other hand, is mainly spent on building the skeletons and shells of diatoms and radiolarians.

3.4.1.1 Nitrogen

Nitrogen is a fundamental element in the development of life, since it is necessary in both the functional and structural aspect of living cells, being a constituent part of proteins and aminoacids.

In natural waterbodies, nitrogen can be found in three different forms, namely dissolved molecular nitrogen (N_2), inorganic substances, and organic nitrogen compounds. Dissolved nitrogen is the most abundant form; nitrogen compounds, organic or not, are only present in small concentrations, amounting scarcely up to a tenth, in weight, of the molecular form.

The principal inorganic forms of nitrogen are nitrate (NO_3^-) and nitrite (NO_2^-) ions, and ammonia (NH_4^+), together with smaller concentrations of other inorganic nitrogen compounds such as nitrous oxide (Riley and Chester, 1971). On the other hand, organic nitrogen compounds can be found

in both dissolved and particulate forms, associated with marine organisms and the product of their metabolism and decay, like urea and proteins; the particulate organic component also includes organic detritus particles and phytoplankton.
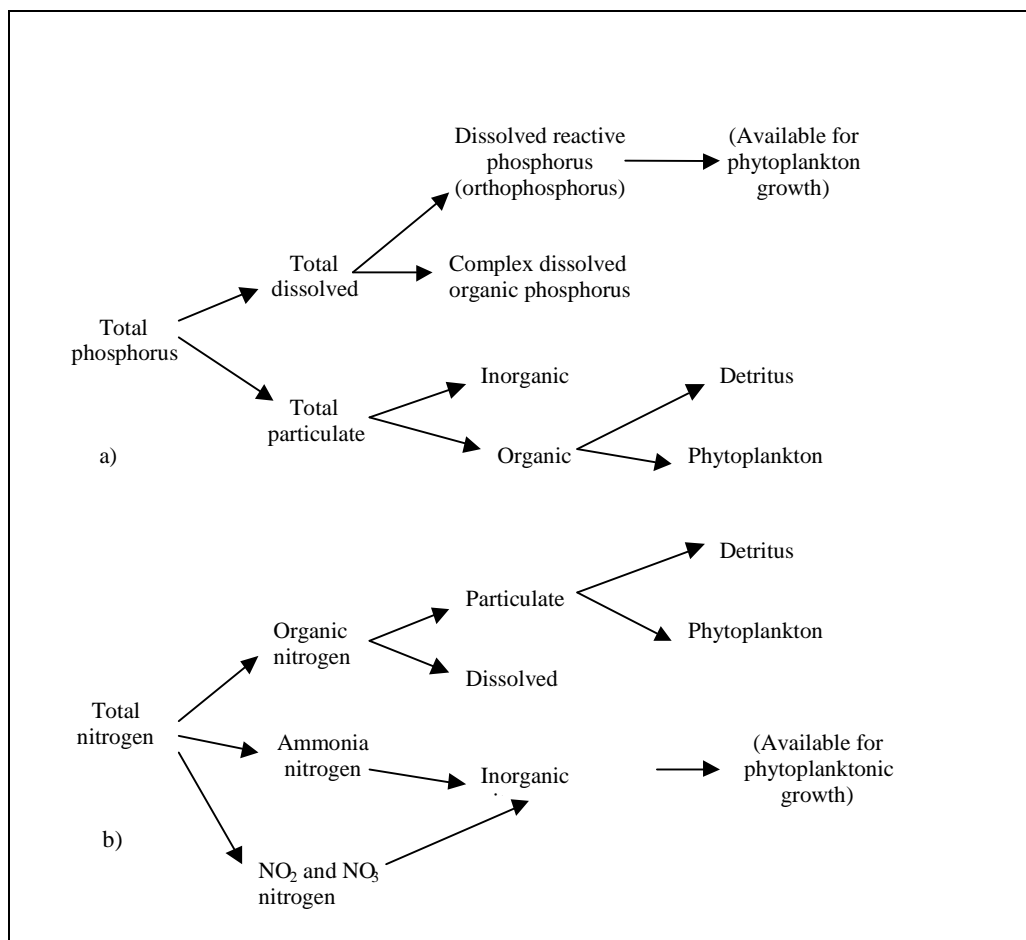

Figure 3.7: Principal components of nutrients: (a) Phosphorus, (b) Nitrogen. From Thomman and Mueller (1987).

3.4.1.2 Phosphorus

The total amount of phosphorus contained in the seawater is accounted for by two main components: dissolved and particulate,.

Dissolved phosphorus can be found either as inorganic or as organic compounds. Inorganic phosphate exists in the sea almost completely in the form of ortophosphate ions (HPO_4^{2-}), although smaller amounts of other ions (PO_4^{3-} and $H_2PO_4^-$) are also present. In addition, poliphosphate ions have been detected in estuarine and coastal waters, as a result of pollution with detergents (Riley and Chester, 1971). Organic phosphorus compounds constitute a variable but significant proportion of the dissolved P in the upper layers of the ocean, and are probably the products of excretion by marine organisms and their decomposition.

Particulate phosphorus forms include inorganic soil runoff phosphorus particulates and organic phosphorus, which in turn comprises detritus and phytoplankton phosphorus and, in general,

the whole range of organic phosphorus compounds involved in the biochemistry of marine organisms and their degradation products. A percentage of the particulate phosphorus is accounted for by possible adsorption of P onto sedimenting particulate matter.

3.4.1.3 Silicon

Silicon is not involved in nutritional processes as nitrogen and phosphorus are, but is used instead to build organism hard parts such as skeletons and shells. Because of this, it is often classified as a different kind of nutrient, and treated separately from N and P.

It can be found in seawater both in dissolved and in particulate form. The former component is probably orthosilicic acid ($Si(OH)_4$), whereas particulate silicon can be found mainly in the form of dead organism hard parts, that slowly dissolve as they sink, and in the form of silicate and aluminosilicate minerals, most of which have been produced by the weathering of rocks on land, and have been transported to the sea by rivers or by wind.

3.4.1.4 Nutrient Sources

The sources of nutrients can be classified into two main types: internal sources and external sources.

a)   Internal sources are usually biochemical processes which take place within the euphotic zone, and the associated nutrient is mainly ammonia or, in some cases, organic nitrogen (urea and aminoacids). Jackson and Williams (1985), and others, have suggested that organic nitrogen and phosphorus may be fixed by phytoplankton in regions where low values of the inorganic form of these nutrients can be found. Thus, it appears that dissolved organic nitrogen and phosphorus can become important sources of nitrogen and phosphorus in the euphotic zone of oligotrophic areas. Upwelling of nutrient-rich deep waters due to vertical mixing or advection can also be included in this group, and in shallow waters the recycling of nutrients from the sediments may be important, even though first they must reach the euphotic layer, where light is available for photosynthesis.

b)   The main external sources of nutrients are municipal and industrial wastes, urban and suburban runoff, agricultural and forest runoff, and atmospheric fallout (figure 3.8). The inputs are nitrates and phosphates from fertilisers washed down by the rain and supplied to the oceans by rivers, which are often the most important sources of nutrients in coastal areas. Additionally, dissolved nitrogen gas in the ocean is supplied by the atmosphere, which becomes an important source of this nutrient. The phosphorus contributed to the seawater from the atmosphere can generally be neglected, even though atmospheric loading can be significant in large lakes and in coastal regions, since the major input mechanism for phosphorus is upwelling from deep waters and discharging from rivers.

The largest source of silicon is also external, since it is supplied to the sea waters mainly by rivers that collect continental runoff, by atmospheric deposition and, in some cases, glacial weathering.

The relative importance of each source depends, in all cases, upon a series of variable factors, such as the existence and erosion of thermoclines, or the existence of short-term pulses of atmospheric dust, for instance; in general, inputs from all sources are variable both in time and space. As an example, table 3.7 gives the measured nutrient inputs from different sources for two locations, for nitrogen and phosphorus.

3.4.2 Nutrient Cycles

Even though each individual nutrient cycle shows specific characteristics, the overall recycling process is very similar for each one of the bio-limiting elements. The general features of nutrient uptake and regeneration are described here, and particular characteristics of nitrogen, phosphorus and silicon cycles will be given below.

		Flux (*) Sargasso Sea	Flux (*) North Pacific Gyre
NITROGEN	Vertical advection	0 - 40	10 – 22
	Eddy diffusion	22 - 210	60 – 730
	Nitrogen fixation	0.5 - 5	0.2 – 2
	Atmospheric input	26 – 54	8 – 26
	TOTAL	$\approx 50 - 310$	$\approx 80 - 800$
PHOSPHORUS	Vertical advection	0 – 4	2.4 – 4.8
	Eddy diffusion	1.7 - 10	5 – 50
	Atmospheric input	0.033 – 0.080	0.005 – 0.012
	TOTAL	$\approx 1.7 - 14$	$\approx 7 - 55$

Table 3.7: Nutrient sources for two different oceanic locations; (*) units are $\mu mol\ m^{-2}\ d^{-1}$. From Chester (1989).

The dissolved form, mainly inorganic, of the nutrients is assimilated by the phytoplankton, on the lower level of the trophic chain, which in turn is grazed upon by zooplanktonic organisms. Part of the nutrient incorporated by the zooplankton is released back into the sea water in a soluble form, while another fraction is excreted in the form of faecal pellets, together with other detrital substances. The phytoplanktonic usage of nutrients, through photosynthesis, requires the presence of light, and is therefore limited to the photic zone (upper layers) of the sea.

The sinking of this zooplankton-generated particulate matter, and the vertical movements of zooplankton and other animals feeding on organisms or detritus, combine to cause a downward displacement of nutrients out of the photic zone. Falling detritus and dead organisms may be used as food by deeper-water animals, and the corresponding nutrients will be excreted as mentioned above; in some cases, microbial communities can settle on the sedimenting matter, and the decay resulting from their activity will contribute to the remineralisation of the nutrients along the water column.

A great majority of the nutrients is recycled in the euphotic zone. In general, approximately a 75% of the nutrients that leave this region are remineralised in the upper 1,000 meters of the water column, in the so-called "particle cycle"; the regenerated nutrients at different depths are then returned to the surface layers by vertical advection and mixing. Another source of nutrients at relatively deep waters is the excreta of animals which live there but feed at the surface waters. Only about 1% of the sinking matter will reach the sea bottom and mix with the sediment, where it will be decomposed mainly by bacteria and benthic fauna.

Several authors (Eppley, 1968; MacIsaac and Dugdale, 1969) have shown that the uptake rate of nutrients by phytoplankton is well described by the Michaelis-Menten equation, when phytoplanktonic growth is limited by the availability of nutrients:

$$V_N = \frac{V_{N,m}[A]}{k_{mN} + [A]}$$

(3.18)

where *[A]* is the nutrient concentration, k_{mN} is the half-saturation constant, and V_m is the maximum uptake rate.
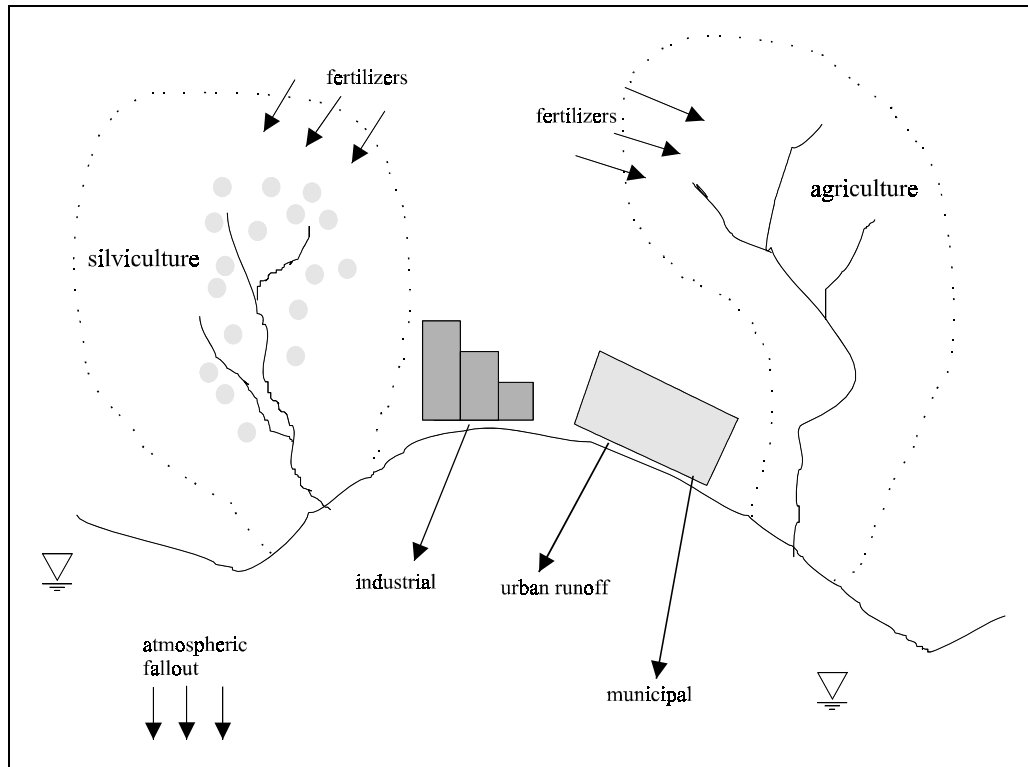

Figure 3.8: General scheme of nutrient sources in oceanic waters (adapted from Thomman and Mueller, 1987).

The Michaelis-Menten half-saturation constant represents the nutrient concentration at which the uptake is one half of the maximum (fig. 3.9). It depends upon a series of factors, such as the form the nutrient takes, the light intensity, phytoplankton size, and temperature; typical values for nutrient uptake are of the order of 10-20 mg/l. In a stationary state, nutrient uptake and phytoplanktonic growth are equivalent, so it can be seen from figure 3.9 that a point is reached from which increasing concentrations of nutrients have no effect on phytoplanktonic growth. Care should be taken when using this equation, because in a given phytoplanktonic community more than one group will be present, and each one will have its own values of k_{mN} and V_m, so an agreement must be reached as to which values to use (Varela, 1991). Nevertheless, MacIsaac and Dugdale (1969) have shown that the concentration of nitrate or ammonia in water, and its rate of uptake, are related through a hyperbolic expression, which would agree mathematically with the Michaelis-Menten equation.

An alternative approach suggests that the rate of growth is related to the nutrient concentration inside the cells, rather than in the environment (Droop, 1973). Mathematically,

$$G = G_\infty \left[1 - \frac{k_Q}{Q_N} \right]$$
(3.19)

where *G* is the growth rate, G_∞ is the growth rate when $Q_N = \infty$, k_Q is the minimum amount of nutrient required for cell growth, and Q_N is the nutrient content within the cell.
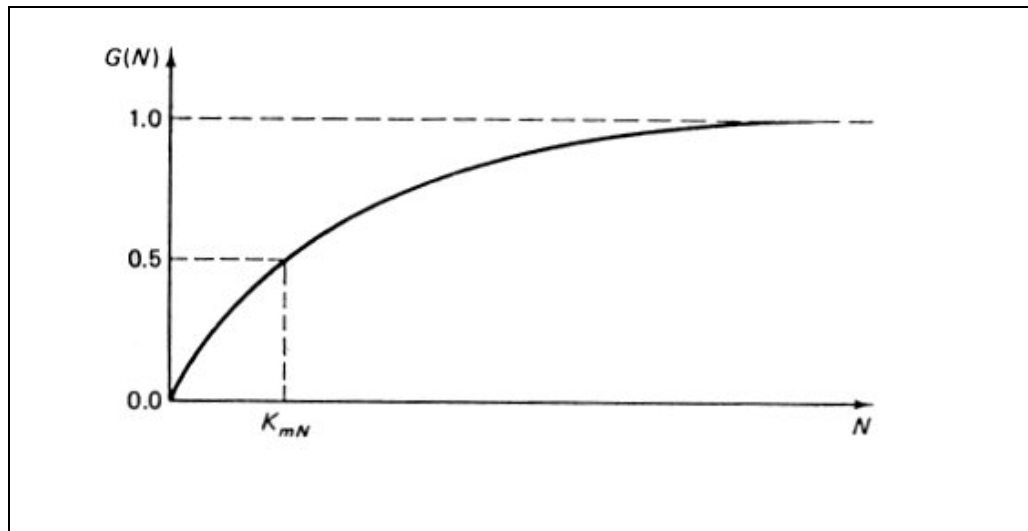
3.23

Figure 3.9: Phytoplanktonic growth rate, G(N), or nutrient uptake rate, depending on nutrients, according to the Michaelis-Menten equation (from Thomman and Mueller, 1987).

It has been said that the concept of limiting nutrients should only be used to refer to new production (i.e., non-recycled nutrients from external sources), since recycling within the marine environment is capable of supplying enough nutrients to maintain the productivity in the surface layers. Studies by Harrison (1980), for instance, revealed that an estimated 80%-90% of the total nutrients used by phytoplankton in the euphotic zone of oceanic waters is derived from regenerative sources. Harrison's work shows that, in oligotrophic environments, micro- and macro-zooplankton excretion can account for up to 90% of phytoplankton nitrogen requirements, and for over 100% of the phosphorus needs. It is estimated that the percentage of each type of production ranges from about 5% of new production in oligotrophic waters of the subtropical gyres to approximately 45% in coastal upwelling regions.

Particular cycle characteristics of the three main nutrients are treated below. The literature concerning nitrogenous nutrients is much more extensive than that referring to phosphorus and silicon, and it is the regeneration processes of N which are best known, due to the variety of chemical forms involved.

3.4.2.1 Nitrogen Cycle

In the sea, the nitrogen cycle cannot be considered a closed system; approximately 10^7 tons of nitrogen are removed yearly from the sea due to deposition of organic nitrogen compounds in sediments (Riley and Chester, 1971), whereas the input of fixed nitrogen from rivers and rainwater occurs at a much greater rate ($8 \cdot 10^7$ tons/year). Due to its traditional role as the limiting element for marine primary production (e.g., Dugdale, 1976), it has been studied in detail, leading to the definition of two new concepts, according to the origin of the nutrient. The nitrogen originated at sources related to physical processes (such as advection, turbulent diffusion, atmospheric transfer, etc.) is termed *new* production, while sources related to biochemical processes supply *regenerated* product. In both types, the main nitrogen form involved is inorganic (nitrate in the former, ammonia in the latter), but organic compounds such as urea or aminoacids may also be included in the regenerated production.

Two characteristics differentiate the nitrogen cycle from the regenerative processes of phosphorus and silicon. Firstly, the complexity of the nitrogen cycle is much greater, because of the large number of different forms of this nutrient, with different oxidation states, which have a significant role in the cycle. Secondly, the main reservoir of nitrogen is in the gaseous form (atmospheric N_2), whereas phosphorus and silicon deposits are sedimentary; this is because nitrogen salts are highly soluble, in contrast with P and Si salts, that present low solubility values.

The concentrations of the various organic and inorganic nitrogen species in the sea are controlled mainly by biological factors, with bacteria, plants and animals playing complementary parts in the various transformations. Physical effects such as the sinking of dead organisms and upwelling, however, tend to bring about a redistribution of these species in the water column, so the instantaneous balance of nitrogen compounds in any waterbody represents a dynamic equilibrium between these various processes. Bacteria probably dominate the regenerative processes in which organic nitrogen compounds are converted into inorganic nitrogen species and, eventually, nitrate, whereas phytoplankton normally synthesise their proteins from nitrate, nitrite and ammonia, forms of nitrogen which are only used by bacteria when organic nitrogen is not available. The only contribution from living animals to the nitrogen cycle is the excretion into the water of ammonia and, to a lesser extent, ammonia precursors such as urea, aminoacids, trimethylamine oxide and peptides.

Riley and Chester (1971) discriminate three main stages in the nitrogen cycle, namely nitrogen fixation, assimilation of fixed nitrogen, and nitrogen regeneration.

The fixation of nitrogen is an endothermic process, requiring a plentiful supply of organic material as an energy source. Some types of blue green algae have been found to fix nitrogen on a large scale in tropical waters, even in the presence of low concentrations of organic matter, apparently obtaining the necessary energy from solar radiation; although certain bacteria, moulds and yeasts share the ability to fix atmospheric nitrogen, it is unlikely that these organisms fix significant amounts of N. It has also been observed that nitrogen fixation is inhibited if an alternative source of inorganic nitrogen is available, which the organisms use in preference to molecular nitrogen (Goering *et al.*, 1966).

Most of the nitrogen required by phytoplankton for the synthesis of their cellular aminoacids is supplied by the ammonia, nitrite and nitrate present in the euphotic layer of the water column, since uptake is a consequence of photosynthesis. Even though all three types of nitrogen can be easily absorbed, phytoplankton shows a clear preference for ammonia (Walsh and Dugdale, 1972), since the reduced form of nitrogen, NH_4^+, seems to be energetically more efficient as a nitrogen source. Expressed in another way, preference for ammonia could be the result of an energy-saving adaptation mechanism by phytoplankton. Experimental results pointing in this direction were obtained by McCarthy *et al.* (1977), who have shown that concentrations of NH_4^+ above 1-2 μM tend to favour the assimilation of ammonia, and to inhibit that of other nitrogen species.

A few classes of phytoplankton seem to be able to satisfy their nitrogen requirements by using dissolved organic nitrogen compounds, such as aminoacids, and some others can do so after attached bacteria have deaminated these compounds. In sewage-polluted waters, an appreciable proportion of phytoplanktonic nitrogen needs may be satisfied by utilisation of urea and uric acid, but in general it is unlikely that significant amounts of N will be derived from the very low concentrations of organic nitrogen compounds present in the open sea. Many species of marine phytoplankton also contribute to the reduction of nitrate to nitrite, by excreting significant amounts of extracellular nitrite when growing in excessively nitrate-rich waters.
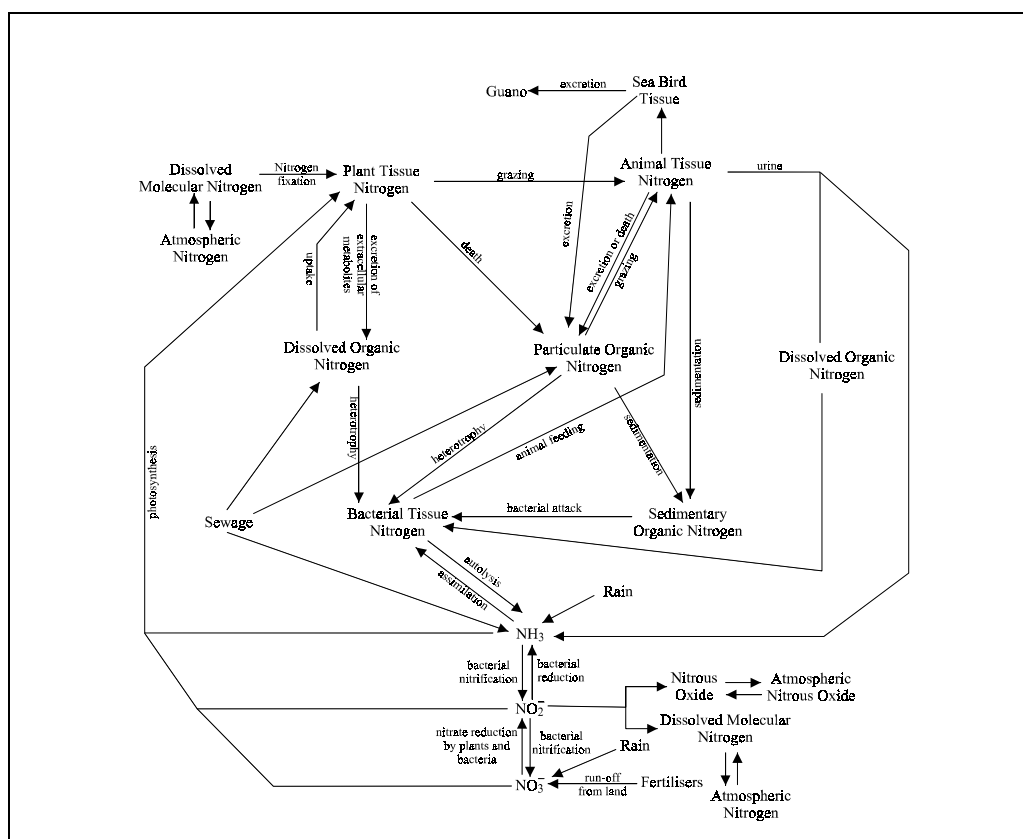
Figure 3.10: Principal transformations involved in the nitrogen cycle, adapted from Riley and Chester (1971).

Although the main nitrogen consumer in the euphotic layer is phytoplankton, some authors (Harrison, 1978; Glibert, 1982) have mentioned that a fraction of the total ammonia may be incorporated by pico- and nano-plankton. In particular, Wheeler and Kirchman (1986) have pointed out that the contribution from heterotrophic bacteria may be important in the depletion of ammonia, and Azam *et al.* (1983) hypothesised that a "microbial loop" existed within pelagic systems, in which organic matter exuded from phytoplankton was utilised by heterotrophic bacteria, which was then grazed upon by protozoa. The role of heterotrophic bacteria as mineralisers of the nitrogenous organic compounds mainly to ammonia is also very important, because the ammonia produced supports the "regenerated primary production", which might be a major part of the total primary production in many cases (Tsirtis, 1995).

In general, different types of heterotrophic bacteria are responsible for at least five processes involving nitrogen forms: deaminization, or breaking of the NH_3 group into aminoacids; nitrification, or conversion from ammonia to nitrites, and then to nitrates; nitrogen assimilation; denitrifiction, which is the opposite process to nitrification; and nitrogen fixation. However, experiments carried out by Carpenter (1983) have shown that the amount of nitrogen fixed on autotrophic and heterotrophic bacteria is small, so this latter depletion mechanism can be usually neglected.

Various species of proetolytic bacteria are responsible for the decomposition into ammonia of soluble and particulate nitrogen compounds of dead organisms and those excreted by plants and animals, and bacterial activity is the main mechanism for nitrification, i.e., the processes by which ammonia is oxidised to nitrite, and then to nitrate. Even though nitrifying bacteria have been found in

all types of ocean waters, bacterial oxidation of ammonia occurs more rapidly in waters with a low oxygen content, independently of the level of organic substrate concentration.

Bacterial action intervenes also in the inverse process, that is, the reduction of nitrate to nitrite, and in denitrification processes by which nitrite or nitrate are reduced to nitrogen or nitrous oxide, respectively. Additionally, the regenerative processes by which the organic nitrogen compounds are reconverted to nitrate via ammonia appear to be mainly due to bacteria. If the organic matter from which bacteria satisfy their energy requirements contains more nitrogen than they really need, the excess is liberated as ammonia (Riley and Chester, 1971).

Even though the phytoplankton and bacteria prefer the inorganic form of nitrogen, all N compounds play a certain role in the marine ecosystem. In natural aerobic waters there is a stepwise transformation from organic nitrogen to ammonia, to nitrite and, finally, to nitrate, as illustrated in figure **3.10**. The differential equations that govern the transformation from one form of nitrogen to another can be expressed as (USEPA, 1987):

$$\frac{dN_4}{dt} = \alpha_1 \rho A - \beta_3 N_4 - \sigma_4 N_4 \tag{3.20}$$

$$\frac{dN_1}{dt} = \beta_3 N_4 - \beta_1 N_1 - F_1 \alpha_1 \mu A + \frac{\sigma_3}{d} \tag{3.21}$$

$$\frac{dN_2}{dt} = \beta_1 N_1 - \beta_2 N_2 \tag{3.22}$$

$$\frac{dN_3}{dt} = \beta_2 N_2 - (1 - F_1) \alpha_1 \mu A \tag{3.23}$$

$$F_1 = \frac{P_N N_1}{P_N N_1 + (1 - P_N) N_3} \tag{3.24}$$

where A is the algal biomass concentration (mg/l), α_1 is the fraction of biomass that is nitrogen, β_1 and β_2 are rate constants of biological oxidation of ammonia nitrogen, and oxidation of nitrite nitrogen, respectively (day^{-1}), β_3 is the rate constant for hydrolysis of organic nitrogen to ammonia (day^{-1}), d is the mean depth of the flow, in feet; F_1 is the fraction of algal nitrogen uptake from the ammonia pool, μ the local specific growth rate of algae (day^{-1}); N_1, N_2, N_3 and N_4 are the concentrations of ammonia, nitrite, nitrate and organic nitrogen (mg/l), respectively, P_N is a preference factor for ammonia, ρ is the algal respiration rate (day^{-1}), and σ_3 and σ_4 are the benthos source rate for ammonia (mg/ft^2-day) and the rate coefficient for organic nitrogen settling (day^{-1}). The preference factor P_N is introduced to account for the fact that phytoplankton seems to incorporate ammonia rather than nitrate when given the choice, as was mentioned above.

3.4.2.2 Phosphorus Cycle

According to Riley and Chester (1971), the distribution of the various forms of phosphorus in the sea is broadly controlled by biological and physical agents that are similar to those which influence the marine chemistry of nitrogen. In a similar manner, it is the inorganic form of P that is used by algae for primary production, but the conversion from organic to inorganic forms should also be considered. Aulenbach and Meisheng (1988) proposed several mechanisms to explain phosphorus removals: rapid removal or adsorption, chemical precipitation involving the transformation of applied soluble phosphate to relatively soluble calcium, aluminium and iron phosphates, biological immobilisation, and plant uptake. The phosphorus cycle is roughly schematised in figure 3.11, and it

must be stressed that it is not a completely closed loop, since part of the phosphorus reaching the sea floor will be converted into phosphate minerals, thus being definitely removed from the water.
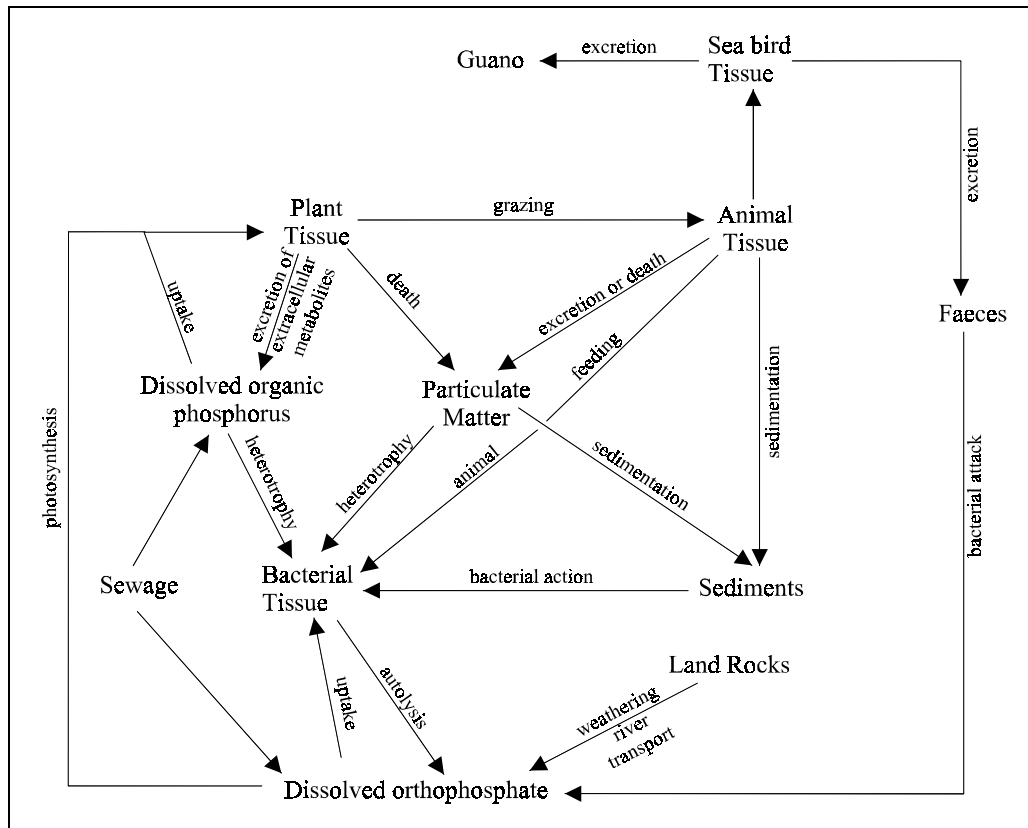

Figure 3.11: General scheme of the phosphorus cycle, adapted from Riley and Chester (1971).

Phosphorus uptake may be due to phytoplankton, or to bacterial population. In the former case, ortho-phosphate is directly assimilated from the sea water for photosynthesis purposes, and absorption and conversion to organic compounds proceeds, even in the absence of light. In the latter case, the large phosphorus requirements are generally satisfied from the organic detritus on which bacteria live, although these microorganisms can assimilate dissolved inorganic P from the water if the amount of phosphorus in organic matter is insufficient. As opposed to nitrogen-based nutrients, bacteria can store more P that they really need, so bacterial uptake of this nutrient will usually exceed its release, acting like a phosphorus sink rather than like a regeneration mechanism.

Although they can mineralise P from dead phytoplankton, the nutrient is often immobilised in bacterial biomass, and remineralised only by the ingestion of bacteria and the subsequent excretion of phosphorus by other organisms (Bloem *et al.*, 1989). Excreted phosphorus may appear in the urine, and unassimilated material is lost in the faecal pellets which contain appreciable amounts of organic phosphates in addition to ortophosphate. Inorganic phosphates from the faecal pellets pass into the sea water, and any undecomposed organic phosphorus compounds are broken down to ortophosphates.

The transformation from one form of phosphorus to another can be described by the following set of equations (USEPA, 1987):

$$\frac{dP_1}{dt} = \alpha_2 \rho A - \beta_4 P_1 - \sigma_5 P_1 \qquad\qquad (3.25)$$

$$\frac{dP_2}{dt} = -\alpha_2 \mu A + \beta_4 P_1 + \frac{\sigma_2}{d} \qquad\qquad (3.26)$$

where α_2 is the phosphorus content of algae (mg/mgA), β_4 is the decay rate of organic phosphorus (day^{-1}), P_1 and P_2 are the concentration (mg/l) of organic and inorganic or dissolved P, respectively, σ_2 is the benthos source rate for dissolved P (mg/ft^2-day), and σ_5 is the organic phosphorus settling rate (day^{-1}).

3.4.2.3 Silicon Cycle

The silicon cycle in marine waters is little known, since its importance as a bio-limiting factor is small, except in very localised areas. Although calcium carbonates are also used by some organisms to form skeletal parts, the availability of calcium and carbon (fifth and eighth most abundant elements in sea water, respectively) discard them as non-limiting elements, and only silicon can therefore be considered to be bio-limiting in this structural sense.

It does seem well established, however, that the ratios between Si and P, and between N and Si are variable, since the general distribution of silicon in the sea differs somewhat from that of phosphates and nitrates. This is due in part to the fact that bacteria are probably not involved directly in the re-solution of silicon from skeletal material, and this has to be done mainly by environmental agents (such as the pressure). The lack of bacterial activity on the sinking silicon leads to its incorporation, either temporarily or permanently, to the sediments, where it may be eroded and re-incorporated into the water column in the dissolved form, at a larger time-scale.

3.4.3 Eutrophication

Depending on the level of primary biological production (eutrophy), waterbodies can generally be classified as oligotrophic, mesotrophic or eutrophic. Characteristics of the former type are low biological productivity, low algal concentration, and high water clarity and transparency. At the other end of the rank, eutrophic lakes and reservoirs present undesirable high production, with high algal concentrations which can eventually become a nuisance, adding colour or taste, or even toxic substances, to the water, and severely affecting its use. The mesotrophic waterbodies lay somewhere in between. The trophic classification (table 3.8) proposed by Nixon (1965) for marine ecosystems differentiated the types of waterbody according to their rate of organic carbon supply:

	Organic Carbon Supply (g C m^{-2}y^{-1})
Oligotrophic	< 100
Mesotrophic	100 – 300
Eutrophic	301 – 500
Hypertrophic	> 500

Table 3.8: Trophic classification of marine waters, acording to the organic C supply.

Eutrophication refers to the process of high primary biological production caused by excessive availability of nutrients in a waterbody, and all its consequences (Stefan, 1994), where primary biological production is taken as the production of aquatic plant mass by photosynthesis in the euphotic layer, i.e., the water layer immediately below the surface, in which the light intensity is sufficient, and the light wave length is appropriate (between 400 and 700 nm), for photosynthesis to take place. A definition of eutrophication was given by Nixon (1965), as an "increase in the rate of supply of organic matter to an ecosystem", thus emphasising the fact that it is a process, not a trophic state.

The eutrophication of coastal waters (and, in general, of any body of water) leads to a series of important ecological and economical hazards. The former may include (Klapper, 1991):

a) High levels of primary productivity due to planktonic plant population.
b) Elevated secondary productivity in food organisms for fish and in fish themselves, when eutrophication is not severe, and a reduction in the proportion of some species of fish.
c) Enhanced elimination of harmful substances due to bioaccumulation and transfer to bottom sediment.
d) Reduced transparency; impaired illumination conditions owing to dense plant stands; planktonic algae may suffer productivity limitations because of self-shading is a possibility.
e) Disturbance of the oxygen budget by daytime - night-time fluctuations. Anaerobic zones may develop, and hydrogen sulphide may also occur in the water and the sediments, leading to increased fish mortality. The living space for fish and their food organisms is drastically reduced.
f) Alteration of the quality and quantity of materials conversion processes. The sediment, under the anoxic conditions of a eutrophic environment, loses its large capacity for phosphate immobilisation, becoming a net exporter of nutrients.
g) Substitution of algae having a useful role in the food chain by largely non-utilisable forms of blue-green algae.

The economical effects include:

a) A threat to the use of water for drinking and fishing.
b) The risk of several diseases (infantile methaemoglobinaemia, nitrosmine-induced cancer and ulceration) arising from consumption of nitrate-rich water.
c) Loss of fish yields and obstruction of fishing activities.
d) Obstruction of navigation.
e) Reduction of reed harvests.
f) Impediments to flow in irrigation and drainage channels.
g) Increased severity of parasitic infestation in tropical bodies of water.
h) Limitations on uses tending to eutrophication such as fish rearing, keeping of waterfowl, etc.

Initially thought to be constrained to inland waters, severe episodes of eutrophication in coastal marine environments have been reported and studied, such as those of the bay of Vilaine (France, Chapelle *et al.* , 1994), the Baltic Sea (Richardson, 1990), the north area of the Adriatic Sea (Chiaudani *et al.*, 1980), or the New York bight (Ryther and Dunstan, 1971). In the particular case of the Vilaine bay, Chapelle *et al.* (1994) point out that dozens of tons of fish died from asphyxia in 1992 as a result of oxygen depletion in bottom waters. Therefore, the capability to predict eutrophication events in any waterbody, and to find solutions to control eutrophication should be a priority objective of all environmental politics.

Even though this phenomenon occurs generally in bounded waterbodies, such as lakes and reservoirs, it can appear in coastal waters under favourable conditions, that is, when the nutrient contribution from external sources is large enough and the rate of water renewal is sufficiently low. In this case, a plankton-based ecosystem, in which the role of microbial and detrital food webs within the nutrient cycles are emphasised, replaces the original benthic ecosystem, due to the effects of overfertilisation and the impact of strong industrial and demographic development of coastal cities. The growth in the rates of nutrient recycling does not correspond to an increase of the capability of the system to fix dissolved nutrients as biomass, and the nutrient surplus is buried in the sediment layer. This unbalanced trophic state gives rise to the proliferation of opportunist species with high growth rates, and phytoplankton blooms, red tide events, and the occurrence of masses of gelatinous aggregates become common (Soler and González, 1994).

Estuaries, coastal and continental shelf regions account for up to one half of the global oceanic primary production (Paerl, 1995). This fact has been commonly attributed to growing levels of direct or indirect anthropogenic nutrient loading in coastal waters, but Paerl (1995), however, has pointed out that from 10% to over 50% of coastal N loading is due to atmospheric wet and dry deposition alone. Atmospheric deposition is a potentially important factor impacting coastal primary production because receiving waters are generally N-limited and hence sensitive to N enrichment.

On the other hand, it is not surprising that an association between riverine nutrient input and eutrophication in coastal waters has been found, since concentrations of nitrogen and phosphorus in many rivers have increased drastically in the last decades, as a consequence mainly of the use of fertilisers and detergents in the watershed. Because the concentration of dissolved silicon has remained approximately constant, the stoichiometric ratios of nutrients, Si:N, N:P and Si:P have changed, possibly having important effects on coastal phytoplankton communities.

Chapelle *et al.* (1994), in their study of eutrophication events in the Vilaine bay, in France, have found that oxygen depletion in the water column, due to phytoplankton respiration and mineralisation of organic matter, was much greater than benthic oxygen consumption; they have also shown that phytoplankton biomass and water hypoxia are further reduced by limiting the nitrogen inputs from discharging rivers than they are when phosphorus loadings are reduced. This could be related to the fact that about twice as much phosphates as can be used by algae is normally present in coastal marine waters, because of the low nitrogen to phosphorus ratio in terrigenous contributions, and to the higher rate of phosphorus regeneration from decomposing organic matter (Ryther and Dunstan, 1971). Both studies conclude that nitrogen is the critical limiting factor to algal growth and eutrophication in coastal marine environments.

3.5 TEMPERATURE

Probably the single most important quality parameter affecting aquatic life is temperature. All biochemical and most physical-chemical reactions are temperature dependent, inasmuch as most reaction rates, whether physical, chemical or biological in nature, are approximately doubled for each 10ºC increase in temperature. Thus, the modelling of temperature in receiving waters is of great importance (Krenkel and Novotny, 1980).

Although it may appear that thermal and wastewater discharges are similar, since they both involve buoyant effluents in a large body of receiving water, they are indeed quite different due to the typical characteristics of the discharges. In general, as pointed out by Fischer *et al.* (1979), discharges

of heated water are larger in quantity, lesser in buoyancy, are usually made in shallow water, and require less dilution. As a consequence, thermal effluents tend to behave as active pollution, modifying the nearshore circulation as a result of their huge momentum.

Thermal pollution in marine environments is generally caused by once-through cooling systems of nearshore large fossil fuel and nuclear power stations, refineries, petrochemical industries, liquefied natural gas facilities, etc. Municipal wastewater effluents, while generally at a temperature level between 10 and 20 ºC, and industrial process water are usually not significant sources of heat. The flow of cooling water from power stations depends on the size of the plant, temperature gradients, efficiency of the heat and its heat exchangers, etc. The discharge is likely to be continuous but seasonal, since more cooling water is required in winter due to the increase of electricity consumption, and larger flows may be necessary in summer when the seawater is warmer.

The excess heat that may change the temperature of the water can affect the marine ecosystem in several ways; if the discharge is in the form of an overflow, as is the most popular and economical case, mixing between the upper and lower layers of the watercolumn is inhibited, minimising oxygen replacement and self-purification in the bottom layers. The smaller amount of dissolved oxygen, the less dilution water available, and a more concentrated organic load in these layers lead to an acceleration of the dissolved organic depletion.

On the other hand, if the discharge is completely mixed with the receiving water, some of the effects mentioned above are eliminated. The rise in temperature, however, still causes a decrease in the ability of water to hold dissolved oxygen, and it also increases the metabolic activity of organisms, the rate of biochemical oxygen demand exertion, and a possible reduction in waste assimilative capacity. As a result, local hypertrophication may appear. This situation is typical of submerged thermal discharges, where the low buoyancy of the effluent induces a vertical fully mixed condition around the diffuser (Kim, 1999).

According to Krenkel and Novotny (1980), the increase in temperature is, in most instances, not large, and probably does not cause biological harm outside the mixing zone. Nevertheless, the possible effects of heat on aquatic life may be:

    a)   Direct death from excessive temperature rise, when it grows beyond thermal death point. This may be particularly serious in tropical areas, where animals and plants have a narrow range of temperature tolerance, and even a rise of a few degrees may cause severe mortality rates.

    b)   Indirect death due to:
       1. less oxygen available,
       2. disruption of the food supply,
       3. decreased resistance to toxic materials; temperature increases have been found to lead to an increase in the rate of the lethal effects of cyanides (WHO, 1982),
       4. decreased resistance to disease,
       5. predation from more tolerant species,
       6. synergism with toxic substances. It is known (WHO, 1982) that several enzyme reactions which take place in organisms are temperature dependent; increase in temperature may result in a significant increase in the uptake and accumulation of certain pollutants.

    c)   Increase in respiration and growth.

d) Competitive replacement by more tolerant species, since the increase in temperature may change some aspects of which the organisms depend, and change behavioural or physiological patterns.

e) Sublethal effects.

All these effects, however, depend on the temperature of the receiving seawater, and also on the efficiency of mixing and dissipation due to currents, tides, and wind. Moreover, as pointed out before, the biota is not only subjected to increased or decreased temperatures, but also to a sudden or "shock" temperature change, due to the varying load factors of most of the sources. Cairns (1955) suggested the effects of these "shock" loadings may be more harmful to marine organisms than continual exposure.

3.5.1 Energy balance within a waterbody

Following Krenkel and Novotny (1980), the energy balance within a waterbody can be expressed as

$$\Phi_S - \Phi_{S\uparrow} + \Phi_L - \Phi_{L\uparrow} - \Phi_B + \Phi_V - \Phi_E - \Phi_H - \Phi_W = \Phi \qquad (3.27)$$

where Φ_s is the shortwave radiation incident to the water surface, $\Phi_{S\uparrow}$ is the reflected shortwave radiation, Φ_L is the incoming longwave radiation from the atmosphere, $\Phi_{L\uparrow}$ is the reflected longwave radiation, Φ_B is the longwave radiation emitted by the body of water, Φ_V is the net energy brought into the body of water in inflow, including precipitation and accounting for outflow, Φ_E is the energy utilised by evaporation, Φ_H is the energy conducted from the body of water as sensible heat, Φ_W is energy carried away by the evaporated water, and Φ is the increase in energy stored in the waterbody.

A short description of the different terms appearing in (3.27) will be now given:

a) The shortwave radiation (Φ_S) is originated directly from the sun, although the energy is scattered by absorption by ozone, scattering by dry air, absorption scattering by particulates, and absorption and scattering by water vapour. It varies with latitude, time of day, season, and cloud clover, as was stated in §3.3.3.3.

b) The longwave atmospheric radiation (Φ_L) depends primarily on air temperature and humidity, and increases with air moisture.

c) The reflected shortwave ($\Phi_{S\uparrow}$) and longwave ($\Phi_{L\uparrow}$) are estimated as a function of the incident radiation, although $\Phi_{L\uparrow}$ can be assumed to be a constant. Krenkel and Novotny (1980) suggest

$$\frac{\Phi_{S\uparrow}}{\Phi_S} = a_s \xi_\mathrm{¤}^{b_S} \qquad\qquad \frac{\Phi_{L\uparrow}}{\Phi_L} = a_L \xi_\mathrm{¤}^{b_L} \qquad (3.28)$$

where $a_{S,L}$ and $b_{S,L}$ are constants that depend on the cloud cover.

d) Longwave radiation may be sent back to the atmosphere by the water (Φ_B), that radiates almost as a perfect blackbody, so the Stefan-Boltzmann law can be applied to its calculation

$$\Phi_B = 0.97\sigma(T_0 + 273)^4 \qquad (3.29)$$

where σ is the Stefan-Boltzmann constant ($1.171\cdot10^{-7}$ cal/cm^{2}deg^{4}day).

e) The energy used by evaporation is a significant term in the energy budget, since each mass unit of water evaporated carries its latent heat of vaporisation. Krenkel and Novotny (1980) propose a general type of evaporation equation

$$E = (a + bU_W(z))(p_{wvp} - p_{avp}) \qquad (3.30)$$

where a and b are empirical coefficients, $U_W(z)$ is the wind speed at a certain elevation z, p_{avp} is the air vapour pressure, p_{wvp} is the saturation vapour pressure of water determined from water surface temperature, and $E = \Phi_E/\rho_{ew}L_{heat}$, with ρ_{ew} the density of evaporated water, and L_{heat} the latent heat of vaporisation.

f) The rate of energy leaving or entering the water body by conduction (Φ_H) due to temperature differences between the waterbody and the air is equal to the product of a heat transfer coefficient and the temperature differential. It may be computed indirectly by means of the Bowen ratio,

$$R_B = \frac{\Phi_H}{\Phi_E} = \frac{0.61\,p_a\,(T_0 - T_{ar})}{1000(p_{wvp} - p_{avp})} \qquad (3.31)$$

where p_a is the atmospheric pressure, and T_{ar} and T_0 are the temperature of the air and water surface, respectively.

g) The energy (Φ_W) carried away by the evaporated water, which is at a higher temperature than the surface water, is relatively small, and can be obtained from

$$\Phi_W = \rho_{ew}c_p V_E(T_{ew} - T_r) \qquad (3.32)$$

where c_p is the specific heat of water, V_E is the volume of evaporated water, T_{ew} is the temperature of evaporated water, and T_r is the base or reference temperature.

h) The net energy contained in water entering and leaving the waterbody (Φ_V) can be expressed as

$$\Phi_v = \frac{\left[c_{p_{si}}\rho_{si}V_{si}(T_{si} - T_r) - c_{p_{so}}\rho_{so}V_{so}(T_{so} - T_r) + c_{p_p}\rho_p V_p(T_p - T_r)\right]}{A} \qquad (3.33)$$

with V the volume of inflowing or outflowing water in the receiving body, T is water temperature, and A is the average surface area; the subscripts stand for surface inflow (*si*), surface outflow (*so*), precipitation (*p*) and base or reference level (*r*).

i) Finally, the increase in energy stored in the waterbody (Φ) can be found from properly averaged field measurements of temperature, and the following equation

$$\Phi = \frac{\left[c_p\rho_1 V_1(T_1 - T_r) - c_p\rho_2 V_2(T_2 - T_r)\right]}{At} \qquad (3.34)$$

where ρ_1 and ρ_2 are the density of the water at temperature T_1 and T_2, respectively, V_1 and V_2 are the volume of the waterbody at the beginning and the end of the measuring period, T_r is the base temperature, A is the average surface area during the period, and t is the length of the period, in days.

Using these concepts, it is possible to predict heat dissipation. Heat transfer in a flowing waterbody is usually a combination of advective motion and radiation transfer, together with emission, scattering and absorption of radiation energy. The equations governing this transport are the continuity and momentum equations, which have been discussed previously, plus an energy or heat balance equation.

3.5.2 Heat balance equation

The transport of heat within a waterbody is governed by a convection-diffusion type equation, that may be written as

$$\frac{\partial T}{\partial t} + u_j \frac{\partial T}{\partial x_j} = \frac{\partial}{\partial x_j}\left(K_{ij}^T \frac{\partial T}{\partial x_i} \right) + DS_{eff} + DS_S \tag{3.35}$$

where K_{ij}^T are diffusion coefficients, and DS_{eff} and DS_S are source terms related to effluent and shortwave radiation heat input. By neglecting the radiation term, which is not of importance in this discussion, only the heated effluent discharge remains as a source of energy; to simplify the analysis, we will further assume that no advection nor turbulent transport exist.

Under these conditions, the energy input due to a heated water discharge can be expressed as (Thomman and Mueller, 1987)

$$W_H = \rho_{eff} c_p Q T_{eff} \tag{3.36}$$

where ρ_{eff} is the effluent water density, T_{eff} is the temperature of the effluent, and Q is the effluent flow. Because the final water temperature T will depend on the balance of heat input and output, its time variation (due only to the effluent) will be

$$\frac{dT}{dt} = \frac{\Delta H}{\rho_{eff} c_p h_{mix}} \tag{3.37}$$

where h_{mix} is the depth over which the net heat input (ΔH) is vertically well-mixed. Edinger *et al.* (1974) have shown that the net heat input can be estimated as

$$\Delta H = K(T_e - T) \tag{3.38}$$

where K is an overall heat exchange coefficient and T_e is the equilibrium temperature, i.e., the temperature that a body of water would reach if all meteorological conditions were constant in time. Equation (3.37) can then be rewritten as

$$\frac{dT}{dt} = \frac{K(T_e - T)}{\rho_{eff} c_p h_{mix}} \tag{3.39}$$

so that, if K and T_e are known or can be estimated, the rate of heat dissipation in a body of water can be calculated.
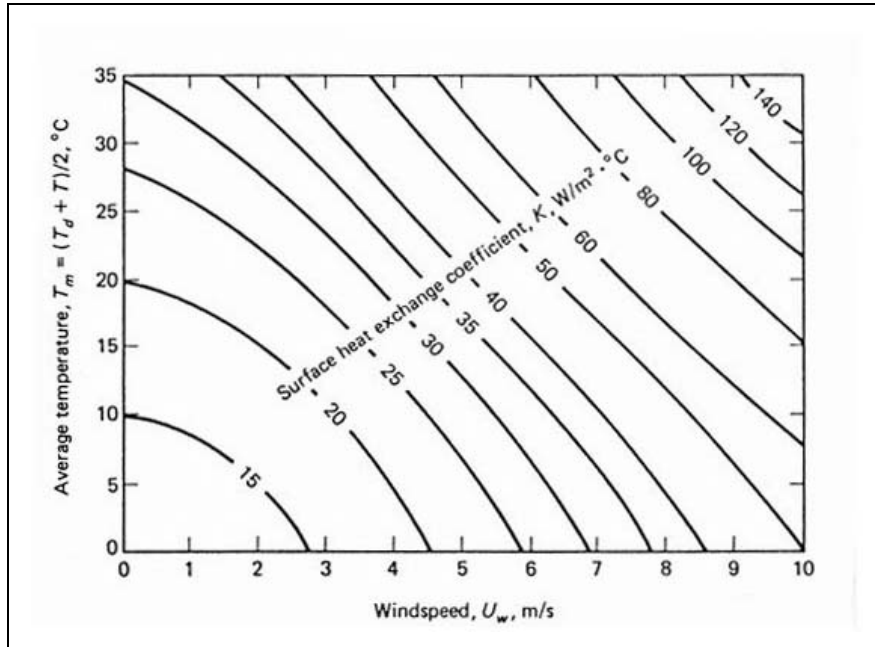

Figure 3.12: Exchange coefficient as a function of wind speed and temperature T_m (from Thomman and Mueller, 1987)

The exchange coefficient K is a complicated function of the water temperature, wind speed, and other meteorological parameters such as the saturated vapour pressure. Edinger *et al.* (1974) give the following expression for this coefficient (in W/m²°C),

$$K = 4.5 + 0.05T + \beta\, f(U_W) + 0.47\, f(U_W) \tag{3.40}$$

where the wind function $f(U_W)$ is in the form

$$f(U_W) = 9.2 + 0.46 U_W^2 \tag{3.41}$$

and the coefficient β is given by

$$\beta = 0.35 + 0.015\, T_m + 0.0012\, T_m^2 \tag{3.42}$$

where
$$T_m^2 = \frac{T + T_d}{2} \tag{3.43}$$

and T_d is the dew point temperature. The relationship between K, wind speed and temperature T_m is shown in figure 3.12.

3.6 SEDIMENTS

Transport of suspended sand is of great importance in marine engineering; in certain cases, such as those resulting from dredging operations, the volume of the transported sediment can be so large that it may pose a threat to the local environment, therefore being considered a pollutant.

The most relevant mechanisms of sediment transport are those related to the structure of the flow, such as currents, vortices, waves and turbulence. In addition, the downward motion due to the weight of the sediment particles is also important. Whereas currents and vortices with horizontal axes transport sediment grains at large distances, the effect of surface waves is mainly to re-entrain deposited sediment into the flow, although they also induce a suspension-maintaining mechanism (Bhattacharya, 1971), and they affect the settling velocity, due to the introduction of non-linearity in the drag force.

3.6.1 Equation of motion

Nielsen (1979, 1992) proposed using a perturbation method to analyse the motion of suspended sediment, by presuming that the velocity of any particle in a flow is given by

$$\mathbf{u_B} = \mathbf{u} + w_S\left[1 + \varepsilon U_{r1} + \varepsilon^2 U_{r1} + ...\right] \tag{3.44}$$

where $\boldsymbol{u}$ is the flow velocity, and ε is the relative magnitude of the fluid accelerations,

$$\varepsilon = \frac{1}{g}\left|\frac{d\mathbf{u}}{dt}\right| \tag{3.45}$$

The perturbation approach is justified by the fact that the fluid accelerations, assuming a wave field superimposed on a quasi-steady current, will be mainly wave-induced, and theey are generally very small compared to the acceleration of gravity (Nielsen, 1992), so that ε is adequately small. In addition, the perturbation method is useful because it reproduces the sediment trapping in vortices – the most important suspension-maintaining mechanism- with the zero-order approximation.

To derive a solution in the perturbation form, the suspended sediment motion equation must be considered. For a spherical particle of diameter d_s and relative density s, moving under the effects of gravity and pressure gradients and drag from the surrounding fluid, this equation can be expressed as (Nielsen, 1992; Horikawa, 1988)

$$\rho s\frac{\pi}{6}d_s^3\frac{du_B}{dt} = \rho s\frac{\pi}{6}d_s^3 g - \frac{\pi}{6}d_s^3\nabla p + \frac{1}{2}\rho\frac{\pi}{4}d_s^2 C'_D\left|u - u_B\right|(u - u_B) +$$
$$+ \rho\frac{\pi}{6}d_s^3 C_M\frac{d_s(u - u_B)}{dt} \tag{3.46}$$

where C'_D is a variable drag coefficient, defined as

$$\frac{C'_D}{C_D} = \left(\frac{|\mathbf{u_B} - \mathbf{u}|}{w_S} \right)^{-\gamma} \tag{3.47}$$

with γ varying between 1, for small particles following the Stokes law (equation (3.118)), and 0 for large particles which settle under fully developed turbulent conditions, and w_S is the sediment settling velocity. The last term in the motion equation corresponds to the fluid pressure on the added hydrodynamic mass, and the coefficient C_M can be assumed to be close to 0.5 (Nielsen, 1992). Equation (3.46) is a simplification of the general equation, which includes the so-called Basset term accounting for changes in the fluid drag due to changes in the flow structure around the particle, and other effects due to the rotation on the particles.

By introducing the relative sediment particle velocity, $\mathbf{u}_r = \mathbf{u_B} - \mathbf{u}$, rearranging terms, and eliminating C'_D using equation (3.47), the following expression is obtained:

$$\frac{d\mathbf{u}_r}{dt} + \alpha \mathbf{g} \left(\frac{|\mathbf{u}_r|}{w_S} \right)^{1-\gamma} \frac{\mathbf{u}_r}{w_S} + \mathbf{u}_r \cdot \nabla \mathbf{u} = \alpha \mathbf{g} - \alpha \frac{d\mathbf{u}}{dt} \tag{3.48}$$

with
$$\alpha = \frac{s-1}{s+C_M} \tag{3.49}$$

Nielsen (1992) points out that the nature of the solutions to the above equations strongly depends on the term $\mathbf{u}_r \cdot \nabla \mathbf{u}$, which corresponds to the structure of the small scale flow. In a pure wave motion ($w_S T_w / L_w \ll 1$), this term is negligible, reflecting the small effect of waves on particle settling rate; in other cases, $\mathbf{u}_r \cdot \nabla \mathbf{u}$ may become very significant, providing sediment trapping mechanisms.

Since this section intends only to offer a general description of sediment transport, further details on the solutions of equation (3.48) for particular cases are not presented, and the reader is referred to Nielsen (1992), Horikawa (1988) or Van Rijn (1993) for an extensive exposure of solutions for different types of particles moving in different kinds of flows.

3.6.2 Bottom-shear stress

Because sediment transport processes are dominant in the near-bed region, it is important to know the magnitude and direction of the velocities and the shear-stresses near the bottom. The latter variable determines whether or not there is grain transport, since it governs the onset of sediment motion, whereas the former is the main transport mechanism.

The total bed shear stress may be seen as consisting of two contributions, the form drag and the skin drag. The former is generated by the pressure difference between the upstream and downstream sides of bed forms, and does not directly affect the stability of individual surface sediment particles (Nielsen, 1992). The skin friction τ'_b, on the other hand, does. On flat beds, it is the skin friction that plays a relevant role on the initiation of sediment transport, whereas both friction types become important when the bed is rippled.

The usual approach to the description of bottom-shear stresses, that is also followed here, considers first the cases in which only currents or waves are present, and then describes the general case of co-existing currents and waves as a combination of the former.

3.6.2.1 Currents

The current-related time-averaged bed-shear stress is defined as

$$\tau_{b,c} = \rho g \frac{\bar{u}^2}{C_h^2} = \frac{1}{8} \rho f_c \bar{u}^2 \qquad (3.50)$$

where $\bar{u}$ is the depth-averaged current velocity, C_h is the Chezy coefficient, and f_c is a Darcy-Weisbach friction factor. Although C_h depends on the type of flow, a widely used expression in coastal hydrodynamic models is

$$C_h = 18 \log\left(\frac{12h}{k_s}\right) \qquad (3.51)$$

where k_s is the Nikuradse or equivalent sand grain roughness (Nikuradse, 1933), usually taken as proportional to a characteristic particle diameter. Using equation (3.51), the friction factor becomes

$$f_c = \frac{8g}{C_h^2} = 0.24\left[\log\left(\frac{12h}{k_s}\right)\right]^{-2} \qquad (3.52)$$

3.6.2.2 Waves

The bottom-shear stress related to wave propagation, averaged over half a wave period, is defined as (Van Rijn, 1993)

$$\tau_{b,w} = \frac{1}{4} \rho f_w \hat{U}_\delta^2 \qquad (3.53)$$

where f_w is a wave-related friction coefficient, and $\hat{U}_\delta$ is the peak velocity at the top of the wave boundary layer, i.e., "the thin layer forming the transition layer between the bed and the upper layer of irrotational oscillatory flow" (Van Rijn, 1993). This velocity can be calculated, under the linear wave theory approximation, as

$$\hat{U}_\delta = \omega \hat{A}_\delta = \frac{\pi H_w}{T_w \sinh(kh)} \qquad (3.54)$$

with

$$\hat{A}_\delta = \frac{H_w}{2\sinh(kh)} \qquad (3.55)$$

the peak value of the orbital excursion at the top of the wave boundary layer. A commonly used expression for the friction factor in rough turbulent flow is that proposed by Swart (1974),

$$f_w = \exp\left[-5.997 + 5.213\left(\frac{\hat{A}_\delta}{k_s}\right)^{-0.194}\right] \qquad (3.56)$$

being $f_{w,max} = 0.3$ for $\dfrac{\hat{A}}{k_s} \leq 1.57$. Based on different experimental data sets, Nielsen (1992) proposed adjusting the coefficients in (3.56) to yield

$$f_w = \exp\left[-6.3 + 5.5\left(\frac{\hat{A}_\delta}{k_s}\right)^{-0.2}\right] \qquad (3.57)$$

3.6.2.3 Currents and waves

Several researchers (e.g., Visser, 1986; Bakker and Van Dorn, 1978; Nap and Van Kampen, 1988) have performed experiments that reveal the distinct influence of waves on current velocity profiles. Based on their data, it appears that wave-induced vortices in the wave boundary layer reduce the current velocities near the bed, the reduction factor depending on the relative magnitude of the current and wave field, and on the angle between waves and currents.

A straightforward estimate of the bed-shear stress in the presence of waves and currents was proposed by Bijker (1986) to be simply

$$\left|\tau_{b,cw}\right| = \tau_{b,c} + \left|\tau_{b,w}\right| \qquad (3.58)$$

where the current-related and wave-related shear stresses may be computed from equations (3.50) and (3.53), respectively. This expression, however, ignores the current velocity reduction due to the wave-current interaction.

A reduction factor was included in the calculation of $\tau_{b,cw}$ by Van Rijn (1989). He assumed a double-logarithmic profile for the current in the presence of waves, one logarithm nearest to the bed, affected by bed-form roughness (k_s), and one affected by an apparent roughness (k_a) outside the near-bed mixing layer (of thickness $\delta \approx 3\delta_w$), and reached the following expression for the combined shear stress:

$$\left|\tau_{b,cw}\right| = \alpha_r \tau_{b,c} + \left|\tau_{b,w}\right| \qquad (3.59)$$

where again $\tau_{b,c}$ and $\tau_{b,w}$ are given by equations (3.50) and (3.53), and α_r is a bed-shear stress reduction factor, defined as

$$\alpha_r = \left[\frac{\ln(30\delta/k_a)}{\ln(30\delta/k_s)}\right]^2 \left[\frac{-1 + \ln(30h/k_s)}{-1 + \ln(30h/k_a)}\right]^2 \qquad (3.60)$$

with the ratio between bed-form and apparent roughness given by

$$\frac{k_a}{k_s} = \exp(\gamma \frac{\hat{U}_\delta}{\bar{u}})$$ (3.61)

and where γ depends on the angle δ_{wc} between waves and currents; an estimate of this coefficient, supplied by Van Rijn (1993), is

$$\gamma = 0.8 + \delta_{wc} - 0.3\delta_{wc}^2$$ (3.62)

3.6.3 Modes of sediment transport

Although in natural conditions there is no sharp division between one and the other, sediment transport modes are usually classified into two types, bed-load and suspended load, which are described below. Between both, Horikawa (1990) includes an intermediate mode, and he also subdivides the suspended sediment movement depending on whether or not the sediment particles are suspended near ripples, and whether or not they are trapped by vortices; finally, he adds a fourth transport mode, the sheet flow transport.

The way in which sediment grains are transported depends mainly on the magnitude of the bed-shear velocity. When the shear velocity just exceeds the critical value for motion initiation, the particles will start to roll and/or slide; for larger values of the bed-shear velocity, the particles will move along the bed in more or less regular jumps, or saltations; for increasing shear stresses, when the shear velocity exceeds the particle settling velocity, the sediment particles will be lifted up to levels at which the upward turbulent forces will be larger than, or comparable to, the submerged weight of the particles, and these will go into suspension.

3.6.3.1 Bed-load transport

This is the mode in which the sediment particles are transported in continuous contact with the bed. In fact, Bagnold (1956) defined bed-load as that part of the total load which is supported by intergranular forces, and bed-load transport as that in which the successive contacts of the particles with the bed are strictly limited by the effect of gravity. Particle displacements take place either by rolling over the bed, sliding, or saltating. Mathematically, Bagnold (1956) proposed

$$\rho(s-1)g\int_z^\infty C_B(z_1)dz_1 = \sigma_e(z)$$ (3.63)

where C_B is the bed-load concentration, and σ_e is the dispersive stress which supports the bed-load travelling above the level z. Solving this equation gave him an estimate of C_B as a function of the dispersive stress,

$$C_B = \frac{-1}{\rho(s-1)g}\frac{d\sigma_e}{dz}$$ (3.64)

and he defined the dispersive stress (or effective normal stress) in terms of the Bagnold number,

$$Bg = \frac{sd_{50}^2 \sqrt{\lambda_S} \dfrac{du}{dz}}{\nu} \tag{3.65}$$

as

$$\sigma_e = \begin{cases} \dfrac{1}{25}\rho s\left(\lambda_S d_{50} \dfrac{du}{dz}\right)^2 & Bg > 450 \\[4mm] 1.3(1+\lambda_S)\left(1+\dfrac{\lambda_S}{2}\right)\rho v \dfrac{du}{dz} & Bg < 40 \end{cases} \tag{3.66}$$

in which λ_S is the relative distance between the surface of sediment particles

$$\lambda_S = \frac{1}{\left(\dfrac{C_{max}}{C}\right)^{1/3} - 1} \tag{3.67}$$

and C is the volumetric concentration, and C_{max} is approximately 0.65.

a) Initiation of motion

The initiation of particle motion under waves and steady currents will occur when the instantaneous fluid force on the sediment grain becomes larger than the instantaneous resisting force due to the submerged particle weight and the friction coefficient. The driving force is strongly related to the local near-bed flow velocity and, therefore, to the shape of the bed.

A simple measure of the ratio between the disturbing and balancing forces acting on a particle under waves is given by the mobility number,

$$\Psi = \frac{(A\omega)^2}{(s-1)gd_{50}} \tag{3.68}$$

where A is the typical semi-excursion and d_{50} is the median particle diameter (Van Rijn, 1993). A different measure of the forces balance, however, was suggested by Shields (1936), who defined the following parameter which bears his name:

$$\theta = \frac{\tau_{b,cw}}{\rho(s-1)gd_{50}} = \frac{u_*^2}{(s-1)gd_{50}} \tag{3.69}$$

where $\tau_{b,cw}$ is the bottom shear stress, u_* is the bottom shear velocity, and use has been made of the relationship between both:

$$\tau_{b,cw} = \rho u_*^2 \tag{3.70}$$

The Shields parameter corresponding to the skin friction,

$$\theta' = \frac{\tau'_{b,wc}}{\rho(s-1)gd_{50}} \tag{3.71}$$

is frequently used to predict initiation of motion and magnitude of sediment concentrations. For flat beds, the form friction vanishes, and $\theta = \theta'$.

The value of θ' at which the sediment starts moving is called critical Shields parameter, θ_{cr}, and is a function of the sediment size and density, the density and viscosity of the fluid, and the flow structure; according to Nielsen (1992, figure 3.13) in most practical cases a value of 0.05 for the critical Shields parameter seems justified. The effective shear stress which corresponds to θ_{cr} following equation (3.71) is the critical shear stress ($\tau_{b,cr}$) for the particular sediment, and bottom shear stresses larger than $\tau_{b,cr}$ will set the particles in motion. The critical depth-averaged velocity for inititiation of sediment motion can be estimated using the Chezy equation:

$$\bar{u}_{cr} = 5.75 u_{*cr} \log\left(\frac{12h}{k_s}\right) = 5.75[(s-1)gd_{50}]^{0.5} \theta_{cr}^{0.5} \log\left(\frac{12h}{k_s}\right) \tag{3.72}$$

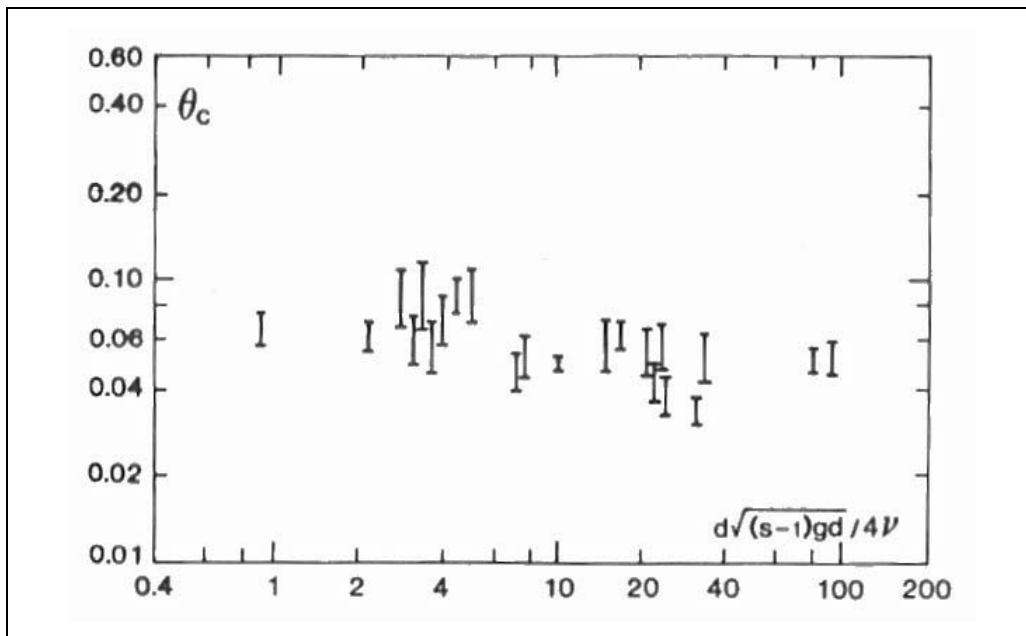where k_s ($\propto d_{90}$) is the effective bed roughness for a flat bed.

Figure 3.13: The critical Shields parameter as a function of sediment characteristics (from Nielsen, 1992).

Under purely oscillatory flows, a time-averaged Shields parameter, analogous to (3.71), is defined, using the time-averaged bed-shear stress related to waves given in equation (3.53). For combined wave and currents, the critical Shield parameter may be obtained by considering the total bed-shear stress as a combination of the current-related and the wave-related stresses (see equation (3.59)).

A different criterion for initiation of motion, although similar to Shields's, was proposed by Amos *et al.* (1988), setting

$$\frac{\tau_{b,c} + \tau_{b,w}}{\rho(s-1)gd} = 0.04 \tag{3.73}$$

where they estimated the current related shear stress as a function of the mean current velocity one meter above the bed ($\overline{u_{100}}$):

$$\tau_{b,c} = \frac{1}{2}\rho 0.003\overline{u}_{100}^{2}$$ (3.74)

b) Bed-load transport rate

The bed-load transport rate can be defined as the product of particle concentration, particle velocity and bed-load layer thickness (Van Rijn, 1993; Nielsen, 1992):

$$Q_B = C_B u_B \delta_B = N_B V_B u_B$$ (3.75)

where C_B is the volumetric sediment concentration in the immobile bed, N_B is the number of moving particles per unit area, and V_B is the particle volume.
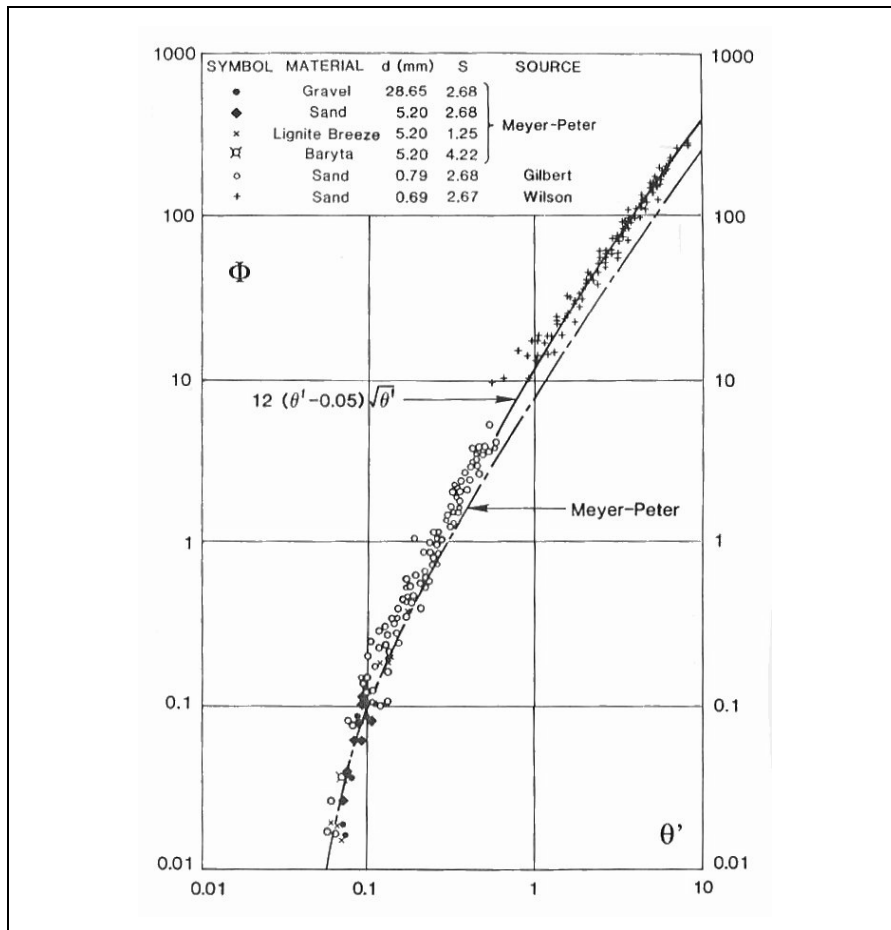

Figure 3.14: Total sediment transport rate over flat beds under steady flows (from Nielsen, 1992).

Several authors have proposed different expressions for the bed-load transport rate, either using experimental data or making different assumptions. Meyer-Peter and Mueller (1948), for

instance, analysed an extensive set of experimental data of bed-load transport in steady flows, and found that it could be expressed as

$$\Phi = \frac{Q_B}{d_m\sqrt{(s-1)gd_m}} = a_1\left(\mu\theta - \theta_{cr}\right)^{1.5}$$

(3.76)

where Φ is a dimensionless flow, d_m is the mean particle diameter, and $\mu = (C_h/C_h')^{1.5}$ is a bed-form factor, with C_h the overall Chezy coefficient, given in (3.51), and $C_h'=18\log(12h/d_{90})$. The proportionality constant a_1 is set equal to 8, but it appears that for large stresses ($\theta>1$) a value of 12 is more appropriate (figure 3.14). A similar equation is given in Nielsen (1992), which shows good results for larger stresses:

$$\Phi = 12\left(\theta - \theta_{cr}\right)\sqrt{\theta}$$

(3.77)

By making some assumption on δ_B (such as relating it to some particle diameter, d_{50} or d_{90}), it is possible to obtain the sediment velocity by using the definition of Q_B and equations (3.76) or (3.77).

On the other hand, Van Rijn (1984a, 1989) assumed that the motion of the bed-load particles is dominated by saltations, and solved the equations of motion for an individual bed-load particle in a unidirectional flow

$$m\ddot{x} - F_L\left(\frac{\dot{z}}{|\mathbf{u_r}|}\right) - F_D\left(\frac{u-\dot{x}}{|\mathbf{u_r}|}\right) = 0$$

$$m\ddot{z} - F_L\left(\frac{u-\dot{x}}{|\mathbf{u_r}|}\right) - F_D\left(\frac{\dot{z}}{|\mathbf{u_r}|}\right) + F_G = 0$$

(3.78)

in which m is the particle mass, $\mathbf{u_r}$ is the particle velocity relative to the flow (of components $u-\dot{x}$ and $\dot{z}$), and F_L, F_D and F_G are the lift force, the drag force, and the submerged particle weight, respectively. After obtaining the saltation characteristics (particle velocity, saltation height, and saltation length), he assumed that the vertical dimension δ_B in equation (3.75) was equivalent to the saltation height, and used measured bed-load transport rates to estimate the bed-load concentrations C_B. He found that

$$\frac{C_B}{C_{max}} = 0.18\frac{T_*}{D_*}$$

(3.79)

where T and D_* are the excess bed-shear stress parameter and the particle parameter as defined in equations (3.80) and (3.81), respectively:

$$T_* = \frac{\tau_b - \tau_{b,cr}}{\tau_{b,cr}}$$

(3.80)

$$D_* = d_{50}\left[\frac{(s-1)g}{v^2}\right]^{1/3}$$

(3.81)

and C_{max} is the maximum volumetric concentration (= 0.65). In the presence of bed forms, the bed-shear stress in equation (3.80) must be substituted by the effective shear stress $\tau_b' = \mu \tau_b$, with the form factor μ defined after equation (3.76).

Using the expressions he found for the saltation characteristics,

$$\frac{\delta_B}{d_{50}} = 0.3 D_*^{0.7} T_*^{0.5} \qquad \text{(saltation height)} \tag{3.82}$$

$$\frac{u_B}{u_*} = 10 - 7\left(\frac{\theta_{cr}}{\theta}\right)^{0.5} \qquad \text{(particle velocity)} \tag{3.83}$$

and equation (3.79), he reached

$$Q_B = \begin{cases} 0.053(s-1)^{0.5} g^{0.5} d_{50}^{1.5} D_*^{-0.3} T_*^{2.1} & T_* < 3 \\[2ex] 0.1(s-1)^{0.5} g^{0.5} d_{50}^{1.5} D_*^{-0.3} T_*^{1.5} & T_* \geq 3 \end{cases} \tag{3.84}$$

Other expressions for bed-load transport rate can be found in Van Rijn (1993) and Nielsen (1992), including equations for particular cases such as transport at low shear-stresses, transport on different types of steeply sloping beds, and transport under different types of waves. Van Rijn (1993) also presents a number of stochastic solutions for Q_B, although he highlights the fact that they are not easily applicable, due to the statistics involved, and they are not more accurate than deterministic solutions for uniform flows.

3.6.3.2 Suspended transport

In this transport mode, sediment particles are moved around without bed contact. The suspended load may contain wash load, which is generally defined as the portion of the suspended load which is governed by the upstream supply rate and not by the composition and characteristics of the bed material (Van Rijn, 1993).

a) Initiation of suspension

Once the motion of the sediment grain has been initiated, an increase of the bottom-shear velocity will lead to the sediment particle moving by saltation, as long as the shear velocity is smaller than the fall velocity. When the shear velocity becomes larger, the particles go into suspension.

Bagnold (1966) observed that a particle only remains in suspension when the turbulent eddies have dominant vertical components which exceed the particle fall velocity. Van Rijn (1984b) noted that several studies on turbulence in boundary layer flow have revealed that the maximum value of the vertical turbulence intensity w' is of the same order as the bed-shear velocity; based on these studies, the critical bed-shear velocity for initiation of sediment motion is

$$\frac{u_{*,cr}}{w_S} = 1 \tag{3.85}$$

from where a critical Shields parameter for sediment suspension may be deduced,

$$\theta_{cr,s} = \frac{w_S^2}{(s-1)gd_{50}} \tag{3.86}$$

A different criterion for initiation of suspension was proposed by Engelund (1965),

$$\frac{u_{*,cr}}{w_S} = 0.25 \tag{3.87}$$

although Van Rijn (1984b) warned that it was based on a crude stability analysis. In the same paper, Van Rijn reports results from experiments undertaken in Delft Hydraulics Laboratory in 1982 from which additional initiation criteria were obtained:

$$\frac{u_{*,cr}}{w_S} = \begin{cases} \dfrac{4}{D_*} & 1 < D_* \leq 10 \\[3mm] 0.4 & D_* > 10 \end{cases} \tag{3.88}$$

b) Suspended load transport rate

The depth-averaged suspended-load transport rate is defined as the product of velocity and concentration, integrated from the edge of the bed-load layer ($z=\delta_B$) to the water surface ($z=h$) –Van Rijn, 1993-:

$$Q_S = \int_{\delta_B}^{h} C(z)u(z)dz = C_B \bar{u}hF \tag{3.89}$$

where $F = \int_{\delta_B}^{h} \dfrac{u(z)}{\bar{u}} \dfrac{C(z)}{C_B} d\left(\dfrac{z}{h}\right)$ is a dimensionless shape factor (e.g., Dyer and Soulsby, 1988). The computation of Q_S requires information on the vertical velocity and concentration profiles, together with a known concentration at a reference level.

The sediment concentration distribution over the water depth can be estimated using different analytical models, which are generally classified into diffusion models, energy models, and stochastic models.

In the former, a time-averaged balance is assumed between the downward transport of sediment particles due to gravity, and the upward transport by the turbulent velocity fluctuations:

$$Cw_S - \overline{c'w'} = 0 \tag{3.90}$$

and Fick's law (equation 2.6) is adopted to express $\overline{c'w'}$ as a multiple of the vertical concentration gradient, so that (3.90) can be written as

$$Cw_S + \varepsilon_S \frac{\partial C}{\partial z} = 0 \tag{3.91}$$

in which ε_S is a mixing coefficient at height z above the bed.

This mixing coefficient is usually assumed to be proportional to the fluid mixing coefficient (eddy viscosity):

$$\varepsilon_S = \beta \phi v_T \qquad (3.92)$$

where β is a factor that describes the difference in the diffusion of a fluid "particle" and a sediment particle, and ϕ accounts for the influence of sediment particles on the turbulence structure of the fluid.

Nielsen (1992), however, has noted that the diffusion model is inadequate to describe suspended sediment distributions, since data from experiments by Coleman (1970) and several other researchers (e.g., McFetridge and Nielsen, 1985) imply that both the magnitude of ε_S as its distribution strongly depend on the settling velocity and, therefore, on the particle size. For such a model to be acceptable, ε_S should be the same for all particles; nevertheless, the diffusion model has been widely used over the years.

Several expressions are available to obtain ε_S, generally as a function of u_*, h and z. Keeping in mind Nielsen's objections to the diffusion model, the most satisfactory seem to be the parabolic and parabolic-constant profiles, the former because of the physics underlying its derivation (linear shear stress distribution and logarithmic velocity profile), and the latter due to its agreement with experimental data:

$$\text{parabolic} \qquad \varepsilon_S = \beta \phi \kappa u_* h \frac{z}{h}\left(1 - \frac{z}{h}\right)$$

$$(3.93)$$

$$\text{parabolic} - \text{constant} \qquad \varepsilon_S = \begin{cases} \beta \phi \kappa u_* h \dfrac{z}{h}\left(1 - \dfrac{z}{h}\right) & \dfrac{z}{h} < 0.5 \\[3mm] 0.25\beta \phi \kappa u_* h & \dfrac{z}{h} \geq 0.5 \end{cases}$$

Substituting the expressions for ε_S in equation (3.91) and integrating, yields the concentration profiles; using the parabolic and parabolic-constant expressions given before, the concentration distributions become

$$\text{parabolic} \qquad \frac{C}{C_a} = \left(\frac{h-z}{z}\frac{a}{h-a}\right)^{w_{S0}/(\beta \kappa u_*)}$$

$$(3.94)$$

$$\text{parabolic} - \text{constant} \qquad \frac{C}{C_a} = \begin{cases} \left(\dfrac{h-z}{z}\dfrac{a}{h-a}\right)^{w_{S0}/(\beta \kappa u_*)} & \dfrac{z}{h} < 0.5 \\[4mm] \left(\dfrac{a}{h-a}\right)^{w_{S0}/(\beta \kappa u_*)} e^{-4\left(w_{S0}/(\beta \kappa u_*)\right)\left(z/h - 0.5\right)} & \dfrac{z}{h} \geq 0.5 \end{cases}$$

for the case of low concentrations ($w_S = w_{S0}$ and $\phi = 1$). The variable C_a represents the bottom boundary condition (i.e., the concentration at a height $z = a$), whereas $w_{S0}/\beta \kappa u_* = Z$ is called the suspension number, with w_{S0} the sediment settling velocity in clear water (§3.6.4.2). When bed forms

are present, their influence must be accounted for through the shear velocity; experimental results (Ikeda, 1980) indicate that the mixing of sediment particles is more intensive in the presence of bed forms, leading to more uniform density profiles than those measured over flat beds.

The reference concentration C_a is generally assumed to be equal to the bed-load concentration (C_B), and the reference level is taken to be the edge of the bed-load layer, ($z = a = \delta_B$). Values for C_B and δ_B over flat beds are given by equations (3.79) and (3.82), respectively, but other expressions may also be used. Engelund and Fredsøe (1976), for example, assumed $\delta_B = 2d_{50}$, and found that

$$C_B = 0.65\left(\frac{\lambda}{1+\lambda}\right)^3 \qquad (3.95)$$

where

$$\lambda = 4.3\left(\frac{\theta - \theta_{cr} - 0.26p}{\theta}\right) \qquad (3.96)$$

and

$$p = \left(1 + \left(\frac{0.26}{\theta - \theta_{cr}}\right)^4\right)^{-0.25} \qquad (3.97)$$

where θ and θ_{cr} are the particle mobility parameter and the critical mobility parameter, respectively, defined before. Zyserman and Fredsøe (1994) used the same definition for δ_B, and came up with a different expression for C_B:

$$C_B = \frac{0.331(\theta' - \theta_{cr})^{1.75}}{1 + 0.72(\theta' - \theta_{cr})^{1.75}} \qquad (3.98)$$

Other expressions for plane beds have been given by Einstein (1950) and Smith and McLean (1977). In the presence of bed forms, the flow becomes non-uniform, and the suspension process becomes more complicated. The suspension of particles occurs on the upsloping part of the dune, being maximum at the top of the slope; over the form troughs, the suspended sediment is transported further upwards by turbulence mixing produced by the shear of the vortex generated in the troughs. Nevertheless, Van Rijn (1984b) proposed an effective reference concentration at the top of the dunes, in the form

$$C_a = 0.015\frac{d_{50}}{a}\frac{T_*^{1.5}}{D_*^{0.3}} \qquad (3.99)$$

in which a (= k_s) is the reference level above the mean bed.

Therefore, using the given expressions for $C(z)$ and C_B, and making an assumption concerning the shape of the velocity profile (usually taken as logarithmic), the suspended sediment transport rate can be calculated; the methods given by Einstein (1950), Van Rijn (1984b) and Bijker (1971) are presented.

Einstein (1950) used a logarithmic velocity profile and a parabolic distribution of the mixing coefficient to express the suspended sediment transport rate as

$$Q_S = 11.6u_*C_a a\left[I_2 + I_1\ln\left(30.2\frac{eh}{d_{65}}\right)\right] \qquad (3.100)$$

3.49

in which I_1 and I_2 are defined below, and e is a correction factor.

$$I_1 = 0.216 \frac{(a/h)^{Z-1}}{(1-a/h)^Z} \int_{a/h}^{1} \left(\frac{1-z/h}{z/h} \right)^Z d\left(\frac{z}{h} \right)$$ (3.101)

$$I_2 = 0.216 \frac{(a/h)^{Z-1}}{(1-a/h)^Z} \int_{a/h}^{1} \left(\frac{1-z/h}{z/h} \right)^Z \ln\left(\frac{z}{h} \right) d\left(\frac{z}{h} \right)$$ (3.102)

A different method, proposed by Van Rijn (1984b), uses the second part of equation (3.89) to calculate Q_S. He assumed a logarithmic velocity profile, and a parabolic-constant mixing coefficient distribution, and he obtained the following expression for F:

$$F = \frac{u_*}{\kappa \overline{u}} \left(\frac{a}{h-a} \right)^Z \left(\int \left(\frac{h-z}{z} \right)^{Z'} \ln\left(\frac{z}{z_0} \right) d\left(\frac{z}{h} \right) + \int_{0.5}^{1} e^{-4Z'(z/h-0.5)} \ln\left(\frac{z}{z_0} \right) d\left(\frac{z}{h} \right) \right)$$ (3.103)

in which Z' is the suspension number corrected by a stratification parameter that accounts for turbulence damping and hindered settling effects due to water stratification:

$$Z' = Z + \psi = Z + 2.5 \left(\frac{w_S}{u_*} \right)^{0.8} \left(\frac{C_a}{C_{max}} \right)^{0.4}$$ (3.104)

where $C_{max} = 0.65$ (Van Rijn, 1984b).

Finally, the method by Bijker (1971), based on similar assumptions as Einstein's, yields

$$Q_S = 1.83 Q_B \left[I_2 + I_1 \ln\left(\frac{33h}{k_s} \right) \right]$$ (3.105)

3.6.3.3 Additional transport modes

An intermediate transport type between bed load and suspended load is mentioned by Horikawa (1988). In this mode, clouds of suspended sediment form over a rippled bed, and there is simultaneous motion of bed-load sediment, which may be entrained into the flow, and suspended-load sediment, which may settle.

A fourth mode is also taken into account by Horikawa (1988), called sheet flow mode, which appears when the shear stress on the bed exceeds a certain limit. In this situation, any sediment ripples disappear, replaced by a thin moving layer of sediment in high concentration, above the bed.

3.6.3.4 Total sediment transport rate

The total load transport rate is determined by adding both the bed-load and suspended sediment contributions

$$Q_T = Q_B + Q_S$$ (3.106)

With this relationship, it is possible to used measured values of total sediment concentration to determine the suspended sediment concentration very close to the bed, although it is still necessary to estimate the bed-load concentration using some of the expressions given above.

Figure 3.15, from Nielsen (1992), based on experimental data from Horikawa *et al.* (1982), shows the measured values of total, relative sediment concentrations, together with the estimates of the bed-load and suspended sediment concentration. Above a certain level ($z = 5$ mm, in this case), the total concentration is due only to suspended sediment, and its behaviour becomes nearly exponential with height above the bed. Below this level, the concentration gradients are considerably greater, and the bed-load contribution increases.
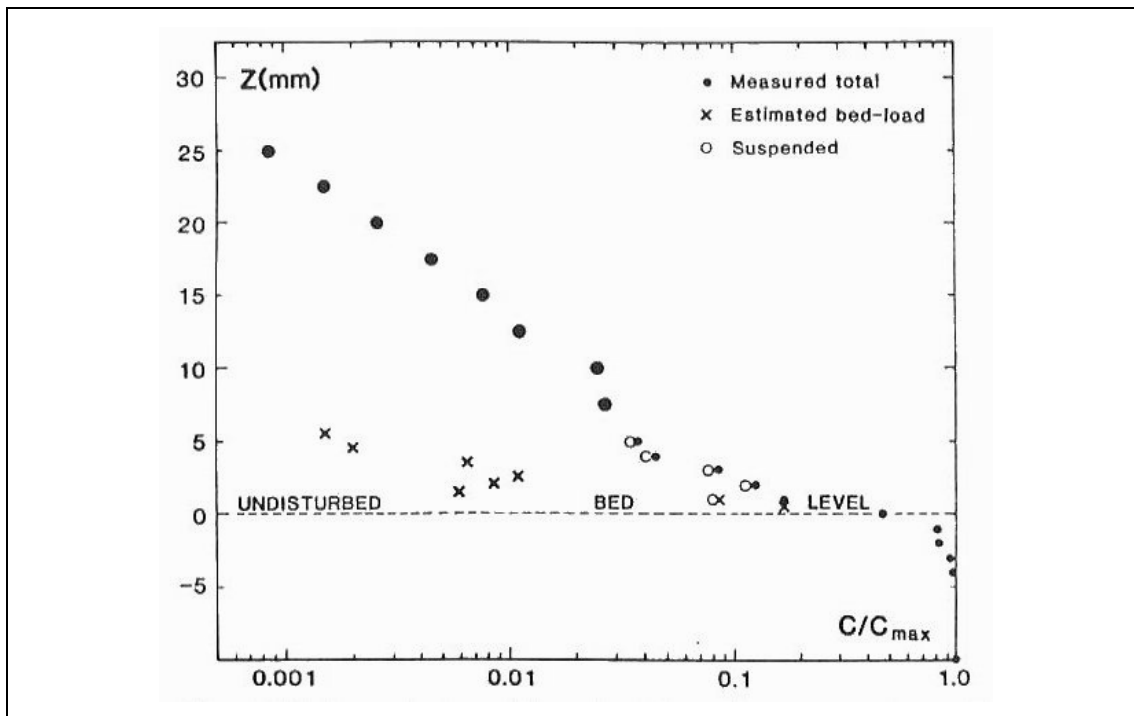

Figure 3.15: Measured values of total, relative sediment concentration, with estimated bed-load and suspended sediment concentrations (from Nielsen, 1992).

The ratio of suspended and total load sediment transport can be written as

$$\frac{Q_S}{Q_T} = \frac{1}{\dfrac{Q_B}{Q_S} + 1}$$

(3.107)

Horikawa (1988) presents a set of tables with over 30 representative formulae for sediment transport rates, after classifying them according to the direction of the sediment movement (longshore, cross-shore and transport under the coexistence of waves and currents) and the driving agent (bottom fluid velocity, bottom shear stress, and wave energy or height). He remarks, however, that the derivation of reliable expressions to describe sediment transport is hampered because of the complexities introduced by the combination of oscillatory and quasi-uniform flow, and he underlines the fact that the equations he presents are to be used only within the limits of their applicability.

3.6.4 Sediment entrainment and deposition

In the case of uniform and steady flows, it is possible to assume equilibrium between bed shear stresses and near-bed sediment concentrations, and relate the sediment transport rates directly to the shear stresses. When the flow is non-uniform or unsteady, this assumption is not realistic, since a decrease in bed-shear stress may coincide with an increase in near-bed concentration due to the settling of suspended sediment. To model sediment transport in this type of flows, it is necessary to adopt a different approach: instead of attempting to describe the mechanisms which distribute the sediment through the water column, sediment entrainment into the flow and sediment deposition are considered as independent processes.

The sediment deposition rate through a level z_d is defined as $d_r(x,y,z,t) = C(x,y,z_d,t)w_S$, where w_S is the sediment settling velocity; the entrainment rate through a level z_d is defined in terms of a pickup function $E_r(x,y,t)$. The rate of bed level change ($\partial z_B/\partial t$) is then related to d_r and E_r, and to the horizontal sedimennt fluxes q_x and q_y through the conservation of sediment equation

$$n\frac{\partial z_B}{\partial t} = d_r - E_r = -\int_0^h \left(\frac{\partial c}{\partial t} + \frac{\partial q_x}{\partial x} + \frac{\partial q_y}{\partial y} \right) dz \qquad (3.108)$$

in which the volume of solids contained in a unit volume of bed material is denoted by n.

3.6.4.1 Pickup functions

The pickup function is a non-negative function with the dimension of a sediment flux. According to Van Rijn (1993), it is usually defined in terms of the number of particles picked up from the bed per unit area and time, i.e.,

$$N_p = \eta n P_S = \frac{\eta}{\alpha_1 d_s^2} P_S \qquad (3.109)$$

in which η is the fraction of susceptible particles per unit area exposed to the flow, n is the number of particles at rest per unit area, d_s is the particle diameter, α_1 is a shape constant, and P_S is the number of pickups per grain per unit time.

Several authors have proposed equations to quantify the pickup rate E_r for different types of flows, and using different approaches. So, Yalin (1977), assumed a proportionality between the pickup timescale and the ratio d/u_*, and suggested using

$$E_r = \alpha \rho_S u_* P \qquad (3.110)$$

where α is a coefficient, ρ_S is the sediment density, and P is the pickup probability, which is a function of the Shields parameter. A similarly stochastic approach was followed by Einstein (1950), who proposed

$$E_r = \alpha \rho_S \left[(s-1)g d_{50} \right]^{0.5} P \qquad (3.111)$$

Experimental data has been used by Fernández-Luque (1974), Nagakawa and Tsujimoto (1980), and Van Rijn (1984). The former author proposed the following expression

$$E_r = \alpha \rho_S \left[(s-1)g d_{50} \right]^{0.5} \left[\theta - \theta_{cr} \right]^{1.5} \tag{3.112}$$

with α = 0.02, based on experiments undertaken within the range $0.05 < \theta < 0.11$ and $0.9mm < d_{50} <$ 1.8mm. Nagakawa and Tsujimoto, on the other hand, analysed data obtained from experiments in the range $0.03 < \theta < 0.2$ and $3mm < d_{50} < 13.5mm$, and concluded that E could be expressed as

$$E_r = \alpha \rho_S \left[(s-1)d_s \right]^{0.5} \left[1 - \frac{0.035}{\theta} \right]^3 \theta \tag{3.113}$$

where α is again equal to 0.02. Finally, Van Rijn (1984b) used experimental data for particles in the range $0.13mm < d_{50} < 1.5mm$, and steady flows such that $0.06 < \theta < 1.0$ to obtain

$$E_r = 0.00033 \rho_S \left[(s-1)g d_{50} \right]^{0.5} D_*^{0.3} T^{1.5} \tag{3.114}$$

For unsteady flows, Nielsen (1992) presents two options to calculate E_r. The first is to infer the pickup function from measured near-bed sediment concentration measurements (C_S), by assuming that the time-averaged pickup function (i.e., the entrainment rate) is equal to the deposition rate ($w_S C_S$), if the process is stationary. The second option is to adapt the steady-flow pickup function, which requires a knowledge about the time-dependence of the Shields parameter. Nielsen (1992) uses Van Rijn's (1984) formula and derives the pickup function for the particular cases of sine waves and irregular waves over a flat bed, and for waves over vortex ripples.

3.6.4.2 Settling velocity

The sediment settling velocity w_{S0} is defined as the terminal velocity of a single sediment particle which falls through an extended, resting fluid (Nielsen, 1992). If the fluid accelerations are not negligible compared to the acceleration of the gravity, this definition must be modified to include the effects of the fluid velocity variations.

In the case of a fluid at rest, the settling velocity is attained when the fluid drag on the particle exactly balances the force of gravity (Nielsen, 1992),

$$F = \frac{1}{2} \rho A C_D w_{S0}^2 = \frac{1}{2} \rho \frac{\pi}{4} d_s^2 C_D w_{S0}^2 = \rho(s-1)\frac{\pi}{6} d_s^3 g \tag{3.115}$$

for a spherical particle with relative density s and diameter d_s, and where ρ is the density of the fluid, and A the area of the projection of the particle upon a plane normal to the flow direction. The settling velocity can then be expressed as

$$w_{S0} = \sqrt{\frac{4(s-1)g d_s}{3C_D}} \tag{3.116}$$

The drag coefficient C_D is a function of the particle Reynolds number $Re = w_{S0}d_s/\nu$ and the particle shape. For spherical particles and very low Reynolds numbers, C_D corresponds to the linear drag force, or Stokes law,

$$F = 3\pi\rho w_{S0}d_s \nu \tag{3.117}$$

and is

$$C_D = \frac{24}{w_{S0}d_s/\nu} = \frac{24}{Re} \tag{3.118}$$

Figure 3.16 illustrates the dependence of C_D on the Reynolds number; the linear asymptote corresponds to the linear drag or Stokes law. Substituting (3.118) into (3.116) yields a simple expression for the fall velocity of very small spherical particles ($Re < 1$),

$$w_{S0} = \frac{(s-1)gd_s^2}{18\nu} \tag{3.119}$$

Fredsøe and Deigaard (1992) proposed an alternative expression for the drag coefficient as a function of Re,

$$C_D = 1.4 + \frac{36}{Re} \tag{3.120}$$

from which a different equation for w_{S0} may be obtained.
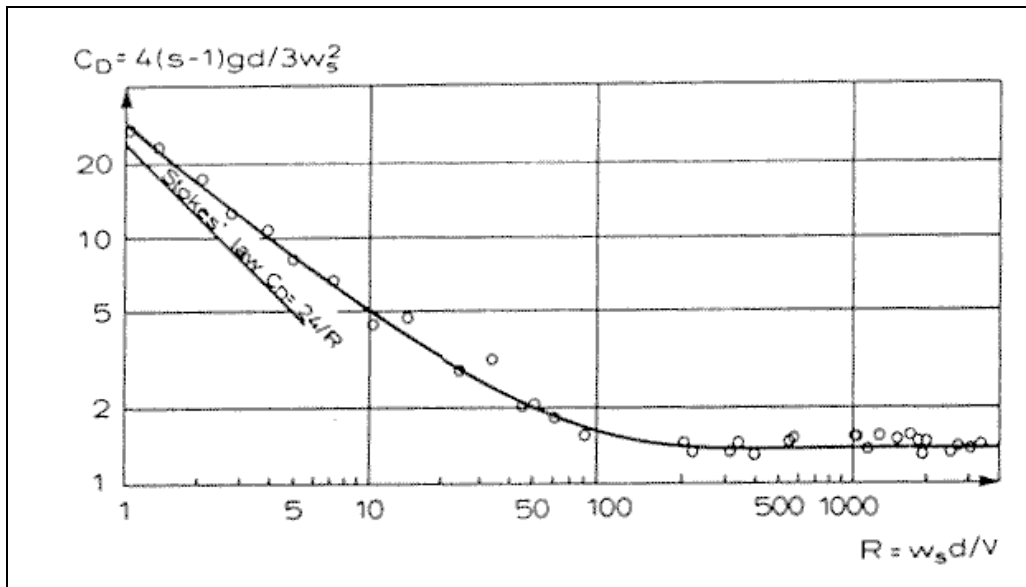

Figure 3.16: Drag coefficients for spherical particles (from Nielsen, 1992).

An empirical formula for settling velocity in a resting fluid was obtained by Gibbs *et al.* (1971), after using glass spheres of different sizes ($0.0063 < d_s$ (cm) < 1.0) and densities. They found that

$$w_{S0} = \frac{-3v + \sqrt{9v^2 + gd_s^2(s-1)(0.003869 + 0.02480d_s)}}{0.011607 + 0.07440d_s} \qquad (3.121)$$

if d_s was measured in cm, and the water viscosity in cm^2/s, yielding a settling velocity in cm/s. Figure 3.17 shows the agreement between this expression and experimental data, and the comparison of equation (3.121) with other equations for the settling velocity of sediments.
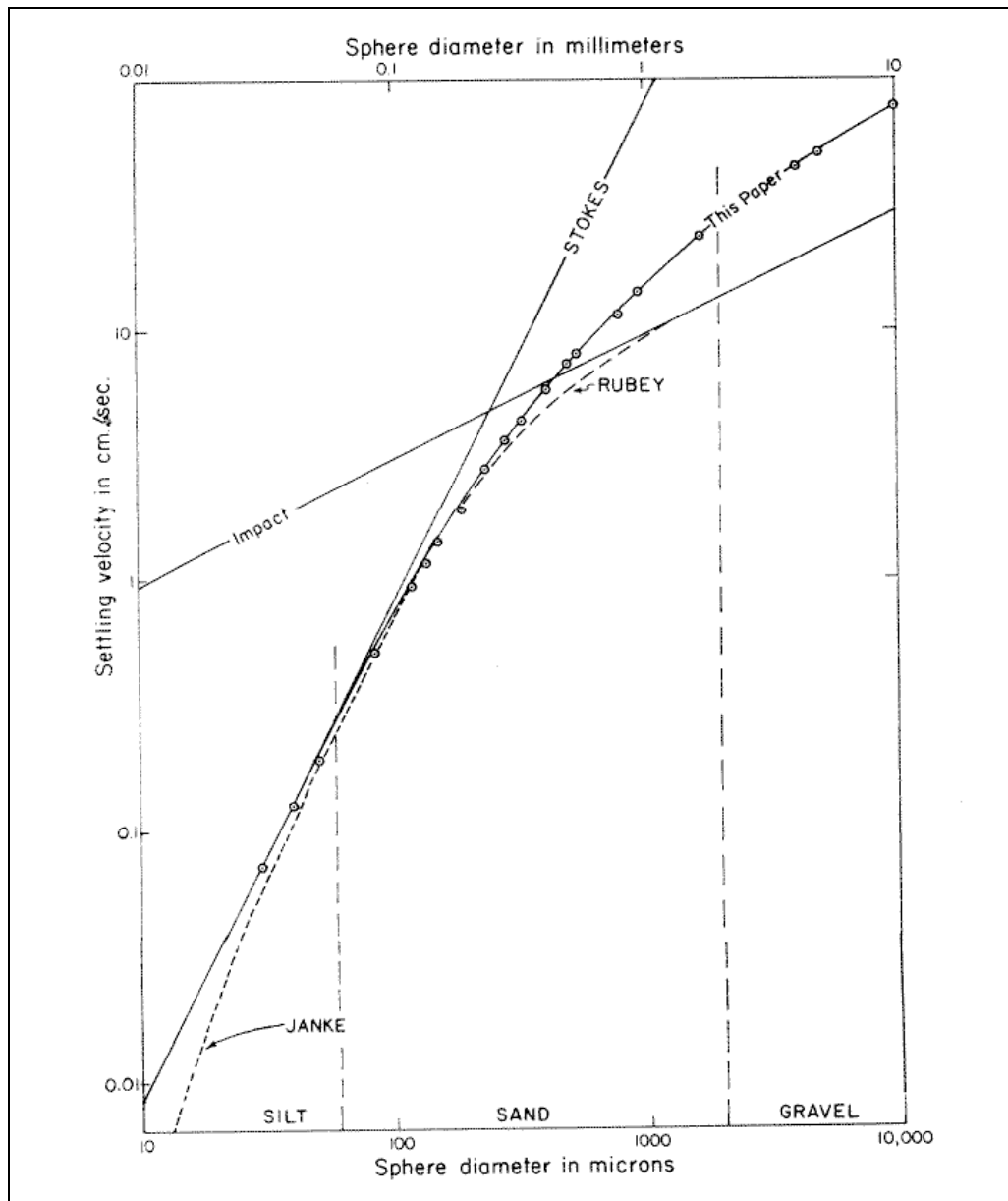

Figure 3.17: Settling velocity versus sphere diameter, for different proposed equations (from Gibbs *et al.*, 1971).

Van Rijn (1984b) proposed an equation for the settling velocity of suspended sand particles in the range of 100-1000 μm:

$$w_{S0} = 10 \frac{\nu}{D_s} \left\{ \left[1 + \frac{0.01(s-1)gD_s^3}{\nu^2} \right]^{0.5} - 1 \right\} \tag{3.122}$$

where D_s is a representative particle diameter, defined in Van Rijn (1984b), which can be considerably smaller than d_{50}. Van Rijn also offers a simple equation for particles larger than about 1000 μm:

$$w_{S0} = 1.1 \left[(s-1)gD_s \right]^{0.5} \tag{3.123}$$

whereas for particles smaller than about 100 μm he used equation (3.119).

Equations (3.116), (3.119) and (3.121) must be taken as upper limits for the velocity as function of size for real particles, because their more or less angular shape increases the drag coefficient C_D, thus reducing w_{S0}. Additionally, the settling velocity may be affected by the sediment concentration in the water; if the sediment forms a thick homogenous suspension, the fall velocity will be somewhat smaller, because the downward motion of each particle will generate an upward compensating flow in the neighbourhood which will delay the downward motion of nearby particles. On the other hand, small particles in a dense cloud of sediments may fall in the wake of larger particles, therefore presenting larger than expected settling velocities.

For non-spherical particles, the effects due to the irregular shape are largest for relatively large particles ($d_s > 300$μm). Van Rijn (1993) supplies a set of equations to obtain the terminal fall velocity for irregularly shaped particles:

$$
\begin{aligned}
w_{S0} &= \frac{(s-1)gd_s^2}{18\nu} & 1 < d_s \leq 100\mu m \\
w_{S0} &= \frac{10\nu}{d_s} \left[\left(1 + \frac{0.01(s-1)gd_s^3}{\nu^2} \right)^{0.5} - 1 \right] & 100 < d_s \leq 1000\mu m \\
w_{S0} &= 1.1 \left[(s-1)gd_s \right]^{0.5} & d_s \leq 1000\mu m
\end{aligned}
\tag{3.124}
$$

Richardson and Zaki (1954) corrected the settling velocity w_{S0} in order to include the effects of the volume sediment concentration C_s, and they proposed

$$w_S = w_{S0}(1 - C_s)^n \tag{3.125}$$

where the exponent n is given by

$$
\begin{cases}
n = 4.45 \text{Re}^{-0.03} & 0.2 < \text{Re} < 1 \\
n = 4.45 \text{Re}^{-0.10} & 1 < \text{Re} < 500 \\
n = 2.39 & 500 < \text{Re}
\end{cases}
\tag{3.126}
$$

Since usually c_s is much smaller than 1, equation (3.125) can be re-written as

$$w_S' = w_{S0}(1 - C_s n) \tag{3.127}$$

Equations (3.125) and (3.127), however, appear to overestimate the settling velocity by a 20% to 30% for small concentrations, as revealed by experiments undertaken by Oliver (1961) and

McNown and Lin (1952); Oliver proposed a different corrective equation which yields better results over the full range of concentrations:

$$w_S' = w_{S0}(1 - 2.15c)(1 - 0.75C_S^{0.33})$$
(3.128)

The variation in the settling velocity of a single particle given by (3.127) is illustrated in figure 3.18, for different particle Reynolds numbers.
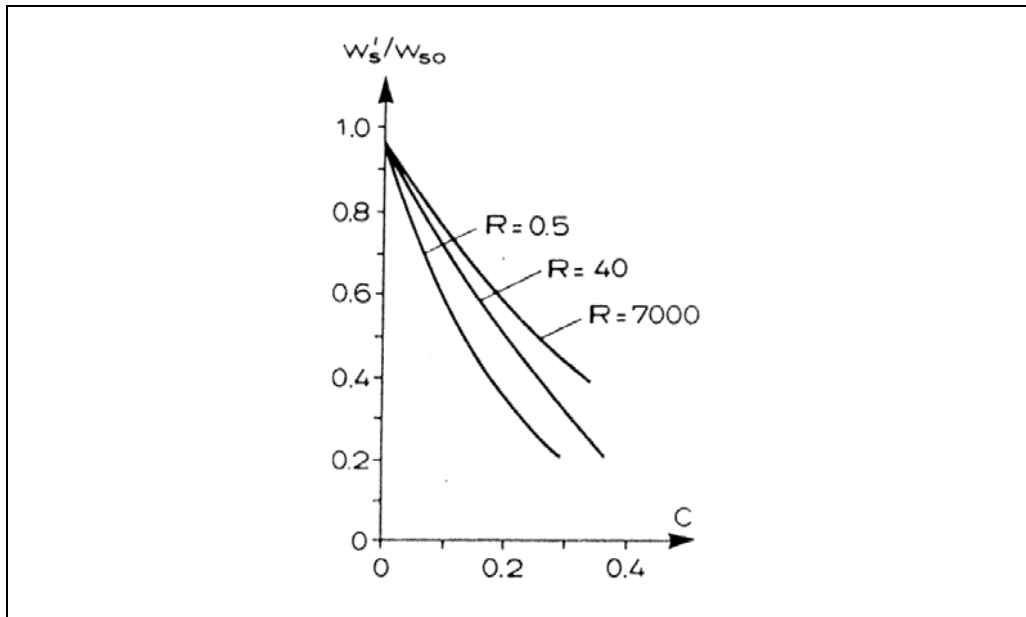

Figure 3.18: Reduction in settling velocity at different concentrations, for various Reynolds numbers (from Fredsøe and Deigaard, 1992).

Additional corrections to the fall velocity of a particle in still water must be introduced to account for the effect of the flow structures on the settling particle. In particular, Hwang (1985) and Nielsen (1984) have shown that, in oscillating flows, the major mechanism governing the fall velocity reduction is the non-linearity of the fluid drag. Nielsen (1994) concluded that, for typical near-bed coastal flows, the delay induced by the oscillations was only $0.002 w_{S0}$. Several authors have studied the effects of turbulence on the fall velocity: Murray (1970) observed that isotropic turbulence effects due to the drag non-linearity could considerably reduce w_{S0}; Jobson and Sayre (1970), on the other hand, found that turbulent motions may slightly increase the particle fall velocity. Another mechanism that may affect the settling velocity is intensive eddy production close to the bed, inducing upward motions which may reduce w_{S0} until the eddies dissolve at higher levels.

3.7 CONCLUSIONS

The quality of coastal waters may be affected by several factors. Generally speaking, the most important are those related to human activity, since they involve the continuous introduction of a very large volume of polluting substances into the environment over an extended period of time. However, the water quality in a coastal environment is not determined by mankind only, but nature itself plays a

role through the introduction of nutrients *via* riverine discharges and by sediment transport along the coast.

On the one hand, direct degradation of the coastal environment is achieved through the discharge into the sea of domestic and industrial wastewater, containing products that might be harmful to marine organisms and to man itself, either *via* the trophic chain, or *via* direct contact with a polluted waterbody. The main source of concern in this aspect are pathogenic organisms discharged into the sea through marine outfalls.

Another anthropogenic threat to the delicate balance of marine ecosystems is the discharge of heated seawater, resulting from the cooling processes required by energy plants and specific industries located on or near the coast. The excess temperature input, although limited to the area closest to the outfall, may cause a number of harmful effects on local marine population, either directly or not.

On the other hand, the riverine discharge of freshwater loaded with nutrients from agricultural runoff, and the transport by ambient currents of large volumes of sediments resulting from dredging processes, also play an important role in determining the quality of coastal seawater.