Climate Change and Coastal Ecosystems: An Overview

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Abstract: The amounts of trace gases notably CO₂, N₂O and CH₄ have been increasing in the earth’s atmosphere. Increased concentration of these gases leads to a warming of the Earth’s surface and the lower atmosphere. The resulting changes in climate and their impacts such as sea level rise, coastal ecosystems etc, can be estimated without associating the origin of the warming to any one of these gases specifically. In this paper, an attempt has been made to review the processes involved in climate change and its effects on sea level rise and on the coastal ecosystems.

Introduction
The Earth intercepts the Sun’s radiation, which is the energy that drives our weather and climate. About one third of this solar radiation is reflected back into space. The rest is absorbed by different parts of the climate system: the atmosphere, oceans, ice, land and various forms of life. The Earth then sends its energy out into space, or re-radiates, in the form of long wave radiation. Some of this energy is again absorbed and re-emitted through a process known as the Greenhouse Effect. The rest of the energy is lost to space.

There is a delicate long-term balance between the outgoing radiation and incoming solar energy. Any change in the factors that affect this process of incoming and outgoing energy, or change the energy distribution itself, will change our climate. Factors affecting global climate is of two types: Natural factors affecting climate change and Human factors.

Natural Factors Affecting Climate
Over the history of the Earth, the climate has changed. The ice ages and intervening warm periods are examples. Some changes are global in scale, while others have been regional or hemispheric. There are a number of natural factors that contribute to changes in the Earth’s climate over various time scales. It is important to understand these factors when attempting to detect a human influence on climate:

Changes in Solar Output: The amount of energy radiating from the sun is not constant. There is evidence of an 11-year solar cycle (variation in energy output) in the temperature record of the Earth. Longer period changes may occur as well.

Changes in the Earth’s Orbit: Slow variations in the Earth’s orbit around the Sun change where and when solar energy is received on Earth. This affects the amount of energy that is reflected and absorbed. These orbital variations are believed to be a factor in initiating the ice ages.

The Greenhouse Effect: When energy from the sun enters the Earth’s atmosphere, about a third of it is reflected back to space. Of the rest, the atmosphere absorbs some, but most of it is absorbed by the surface of the Earth. The Earth emits energy at longer wavelengths. Some of this energy escapes to space but some of it is absorbed again and re-emitted by clouds and greenhouse gases such as water vapour, carbon dioxide, methane and nitrous oxide. This helps to warm the surface and the troposphere (lowest layer of the atmosphere), keeping it 33°C warmer than it would otherwise be. This is the natural Greenhouse Effect, and vital to life as we know it.
**Aerosols**: These are very fine particles and droplets that are small enough to remain suspended in the atmosphere for considerable period. They both reflect and absorb incoming solar radiation. Changing the type and quantity of aerosols in the atmosphere affects the amount of solar energy reflected or absorbed.

**Interaction between Solar Energy and the Atmosphere**

The sun is a large nuclear reactor, fusing H\(_2\) atoms together and radiates enormous energy into space in the form of electromagnetic waves. These waves have many different wavelengths, important because different wavelengths have different characteristics. Some are visible (light); some warm (infrared); some that causes sunburn (ultraviolet), some cells that encourage cancer in cells (X rays and gamma rays) and other low frequency waves. Most of the sun’s radiation (81%) occurs in the visible and near-infrared wavelengths. Fortunately, not all of this radiation reaches the Earth’s surface, and the atmosphere absorbs much of the energy. But absorption is not a simple process where each gas behaves selectively in absorbing different wavelengths. For example, let us contrast CO\(_2\) and O\(_2\), which absorb energy very differently. CO\(_2\) does not absorb much energy but O\(_2\) does. As solar energy arrives, it filters through the atmosphere (Figure 1).

If the O\(_3\) in the atmosphere decreases, as when an O\(_3\)-hole develops, more UV radiation reaches the Earth’s surface. In fact, 7% more UV radiation now reaches the Earth’s surface than did so 10 years ago. Increased UV radiation is harmful to the reproductive cycle of living organisms and increases the risk of skin cancer and vision disorders such as cataracts (Figure 2B).

**Figure 1: Significance of the atmosphere.**

When it encounters CO\(_2\) molecules, the far-infrared wavelengths are absorbed, because CO\(_2\) molecules have the right banding to do so. In fact, CO\(_2\) absorbs all wavelengths longer than about 20 micrometres. But very little solar radiation occurs in such long wavelengths, so CO\(_2\) does not have much solar energy to absorb. The ultraviolet and visible light energy streams freely pass through the CO\(_2\) molecules on its way to the Earth’s surface. In contrast, molecules of O\(_2\) and ozone (O\(_3\)) are bonded differently from CO\(_2\), and are able to absorb all wavelengths at the ultraviolet end of the spectrum (Figure 2A). This shields from the dangerous UV radiation and explains why there is much concern about the O\(_3\) layer.

**Figure 2: The atmosphere and the incoming solar radiation.**

Figure 2 has six graphs showing the absorption of different gases in the atmosphere. We can see that no atmospheric gas absorbs much energy in the visible and near-infrared wavelengths. This is the range in which the sun does most of its radiating, so the bulk of these wavelengths reach the Earth’s surface unimpeded. Overall, the atmosphere absorbs or reflects back into space about half of the solar radiation that tries to enter. The remaining half reaches the Earth’s surface, a third of what remains is reflected back into space. What is left, ~35% of the energy originally received at the top of the atmosphere, is absorbed by the Earth’s surface, warming it.

Earth’s warmed surface acts like any heated surface: it radiates its heat energy into the air. In effect, earth receives light energy, translates it into heat energy and reradiates back into the atmosphere as longer far-infrared wavelengths (Figure 2C). Earth’s re-radiation is important to understand, as it is the key to the greenhouse effect and to global warming.
Radiative Forcing

The ultimate energy source for all weather and climate is radiation from the sun called ‘solar or short-wave radiation’. Averaged globally and annually, about a third of the incoming solar radiation is reflected back to space. Of the remainder the atmosphere absorbs some, but most is absorbed by the land, ocean and ice surfaces. The solar radiation absorbed by the Earth’s surface and atmosphere (which amounts to about 240 W m⁻²) is balanced at the top of the atmosphere by outgoing radiation at infrared wavelengths (Figure 3). Some of the outgoing infrared radiation is trapped by the naturally occurring greenhouse gases (principally water vapour but also CO₂, O₃, CH₄ and N₂O) and by clouds, which keeps the surface and the troposphere about 33°C warmer than it would otherwise be. This is the ‘natural greenhouse effect’. In an unperturbed state, the net incoming solar radiation at the top of the atmosphere averaged over the globe over long periods of time must be balanced by the net outgoing infrared radiation (Figure 3).

![Figure 3: A simplified diagram illustrating the global long-term radiative balance of the atmosphere.](image)

A change in average net radiation at the top of the troposphere (known as the tropopause), because of a change in either solar or infrared radiation, is defined as ‘radiative forcing’. A radiative forcing perturbs the balance between incoming and outgoing radiation. Over time, climate responds to the perturbation to reestablish the radiative balance. A ‘positive radiative forcing’ tends on average to warm the surface while a ‘negative radiative forcing’ tends on average to cool the surface. Radiative forcing averaged globally has been used to compare the potential climatic effect of different climate change mechanisms. For a range of mechanisms, there appears to be a similar relationship between global mean radiative forcing and global mean surface temperature change. However, the applicability of global mean radiative forcing to mechanisms such as changes in O₃ or tropospheric aerosols, which are spatially very inhomogeneous, is unclear. It should be noted that climate variations are believed to occur in the absence of any radiative forcing as a result of the complex interactions between the atmosphere and the oceans and possibly the cryosphere, land surface and the biosphere. The key greenhouse gases with potentially high radiative forcing are discussed below.

Greenhouse Gases

Most greenhouse gases occur naturally. However, modern industry and lifestyles have led to new sources of greenhouse gases, as well as to the emission of entirely new greenhouse gases. Greenhouse gases in the atmosphere are a part of the biogeochemical cycles. They are formed by many processes and then emitted to the atmosphere, where they undergo chemical and/or physical transformations. Finally, they reach the surface again by dry and wet deposition i.e., either by atmospheric turbulence or in precipitation. A dynamic equilibrium exists between the strength of sources and sinks that determines mixing ratios and residence time in the atmosphere. The residence times are rather different, varying from about 12 years for CH₄, 120 years for N₂O to several thousand years for some new man-made greenhouse gases.

Carbon dioxide (CO₂)

Carbon dioxide (Figure 4) comes from the decay of materials, respiration of plant and animal life and the natural and human-induced combustion of materials and fuels. It is removed from the atmosphere through photosynthesis and ocean absorption. CO₂ levels in the atmosphere have increased since the pre-industrial period from about 280 to 356 ppmv (IPCC, 1994).

![Figure 4: Molecular structure of carbon dioxide (CO₂).](image)

Rising with the smoke of burning fossil fuels and cleared forests, CO₂ lasts up to a 100 years in the atmosphere. The most abundant heat-absorbing gas after water vapour, it accounts for roughly half the man-made share of the global warming problem.
The radiative forcing due to this increase is 1.56 W m\textsuperscript{-2} (Figure 5). Evidence that the observed increase in atmospheric CO\textsubscript{2} concentration due to anthropogenic activity comes from the following facts:

- The long-term rise in atmospheric CO\textsubscript{2} closely follows the increase in anthropogenic CO\textsubscript{2} emissions.
- Although CO\textsubscript{2} is well mixed in the atmosphere, concentrations are slightly higher in the Northern Hemisphere due to higher emissions. The increase in the inter-hemispheric gradient is growing in parallel with CO\textsubscript{2} emissions.
- Fossil fuel and biospheric carbon have a lower ratio of the carbon isotope \(^{13}\text{C}\) to the isotope \(^{12}\text{C}\). Fossil fuels contain no \(^{14}\text{C}\) because of their age. Decreases since pre-industrial times in the \(^{13}\text{C} : ^{12}\text{C}\) isotope ratio and in \(^{14}\text{C}\) are fully consistent with the addition of fossil fuel and biospheric carbon by human activity.

**Carbon Cycle** Atmospheric CO\textsubscript{2} provides a link between biological, physical and anthropogenic processes. Carbon is exchanged between the atmosphere, the oceans, and the terrestrial biosphere and more slowly with sediments and sedimentary rock. Our understanding of the C cycle has improved our knowledge about the removal of CO\textsubscript{2} from the atmosphere, which is distributed between the sinks in the ocean and on land, and in the preliminary quantification of feedbacks. Considerable quantitative uncertainty remains regarding the processes which contribute to the sinks on land and in the ocean. The main anthropogenic sources of CO\textsubscript{2} are the burning of fossil fuels and land use changes. Land use changes cause both release and uptake of CO\textsubscript{2}. On an average, CO\textsubscript{2} will be released to the atmosphere, if the original ecosystem stored more C than the modified ecosystem, which replaces it. Deforestation acts as a CO\textsubscript{2} source. The latest estimate of net CO\textsubscript{2} release due to global land use changes 1.1 ± 1.2 GtC yr\textsuperscript{-1} (IPCC, 1994). The oceans are a large sink of anthropogenic CO\textsubscript{2}. At present, net global oceanic uptake cannot be measured directly: it is estimated using models which describe the exchanges between the surface and deep ocean and the atmosphere.

CO\textsubscript{2} concentration measurements in ice cores have shown that the concentration in the atmosphere has been nearly constant for the last 1000 years until the beginning of industrialization: it can be assumed that the system has been in a steady state in which large amounts of CO\textsubscript{2} have been exchanged between the atmosphere, but a balance has been kept. The anthropogenic input of CO\textsubscript{2} into the atmosphere acts as a perturbation of this steady state.

After the sediments, the ocean is by far the largest carbon pool: about 50 times more carbon compared to the biosphere. The oceanic biosphere contains comparatively few amounts of carbon (>30 Gt C), the biggest part can be found in inorganic compounds (about 37,500 Gt C). In addition, there are also the dissolved carbon and the particulate carbon (about 1000 Gt C). The ocean can be roughly divided into the following layers. There is a well-mixed surface layer (50-100 m depth), which shows permanent interaction with the atmosphere. Below this layer is the steep temperature gradient called permanent thermocline (up to about 1000 m depth), followed by the intermediate water and deep sea. In the deep ocean, the interactions occur on time scales very different from those of the processes in the upper layers. There are three processes, which take part in the storage of carbon in the ocean:

- The chemical balance: CO\textsubscript{2} is dissolved in the surface layer and reacts with water to bicarbonate and carbonate;
- The biological pump: fixation of carbon by marine phytoplankton; and
- The physical pump: sinking of cool surface water with higher CO\textsubscript{2} content and upwelling of deep seawater with a lower CO\textsubscript{2} concentration.

The oceanic uptake of excess CO\textsubscript{2} proceeds by:
1. Transfer of CO\textsubscript{2} gas through the air-sea interface;
2. Chemical interactions with the oceanic dissolved inorganic carbon; and
3. Transport into the thermocline and deep waters by means of water mass transport and mixing processes.

Though there are large geographical differences and seasonal variations of the surface ocean partial pressure of CO$_2$, averaged globally and annually, the surface water value is close to that for equilibrium with the atmosphere. Therefore processes (2) and (3) are the main factors limiting the capacity of the oceans to serve as a sink on decadal and centennial time scales. The chemical buffering reactions between dissolved CO$_2$ and the HCO$_3^-$ and CO$_3^{2-}$ ions reduce the rate of oceanic CO$_2$ uptake.

The marine biota, if in steady state, is believed to play a minor role, if any at all, in the uptake of excess anthropogenic CO$_2$. The marine biota, however, play a crucial role in maintaining the steady-state level of atmospheric CO$_2$. About three-quarters of the vertical gradient in dissolved inorganic carbon is generated by the export of newly produced carbon from the surface ocean and its regeneration at depth (a process referred to as “biological pump”). In the open ocean, however, this process is believed to be limited by the availability of nutrients, light, or by phytoplankton population control via grazing, and not by the abundance of carbon (Falkowski and Wilson, 1992). Therefore, a direct effect of increased dissolved inorganic carbon (less than 2.5% since pre-industrial times) on carbon fixation and export is unlikely, although recent studies suggested that under particular conditions the rate of photosynthesis and hence phytoplankton growth might indeed be limited by availability of CO$_2$ as a dissolved gas.

Flux of carbon from the terrestrial biosphere to the oceans takes place via river transport. Global river discharge of carbon in organic and inorganic forms may be $\sim$1.2-1.4 Gt C yr$^{-1}$ (Schlesinger and Melack, 1981; Meybeck, 1993). A substantial fraction of this transport (up to 0.8 Gt C yr$^{-1}$), however, reflects the natural geochemical cycling of carbon and thus does not affect the global budget of the anthropogenic CO$_2$ perturbation (Sarmiento and Sundquist, 1992). Furthermore, the anthropogenically induced river carbon fluxes reflect to a large extent, increased soil erosion and not a removal of excess atmospheric CO$_2$. The role of coastal seas in the global carbon budget is poorly understood. Up to 30% of the total ocean productivity is attributed to marine productivity in the coastal seas, which comprise only $\sim$8% of the oceanic surface area. Here, discharge of excess nutrients by rivers might have significantly stimulated carbon fixation.

**CO$_2$ Fertilization** Increased levels of CO$_2$ can stimulate photosynthesis. Studies carried out on small-scale experimental stands of vegetation, under optimal conditions of water and nutrient supply, suggest potential increases in photosynthesis of 20 to 40% when CO$_2$ is doubled. However, when the availability of water and nutrients is taken into account, the fertilization effect is likely to be reduced (IPCC, 1994). Higher temperatures and precipitation can increase photosynthesis and plant growth and hence increase carbon storage in living vegetation and litter, which is a negative feedback. The storage of carbon in soil tends to decrease with increasing temperature due to increased rates of decomposition, which is a positive feedback. Soils, which are flooded, can store large amounts of carbon in the form of peat. Drying, due to changes in precipitation and/or evapotranspiration and possibly warming of such regions would release additional CO$_2$. Climate feedbacks also influence the storage of carbon in the ocean through changes in sea surface temperature, ocean circulation and the marine biological pump.

**Methane (CH$_4$)** Although there is less methane (Figure 6) than carbon dioxide in the atmosphere, methane is a more effective heat-trapping gas. It comes from the decay of matter without the presence of oxygen. Primary sources include wetlands, rice paddies, animal digestive processes, fossil fuel extraction, and decaying garbage.

![Figure 6: Molecular structure of methane (CH$_4$).](image)

Although methane (CH$_4$) lasts only ten years, molecule for molecule, it absorbs 20 to 30 times more heat than CO$_2$. Most CH$_4$ is generated by bacteria breaking down organic matter in the absence of O$_2$, as in flooded rice fields, the guts of cattle, wetlands and landfills. It contributes 15 to 20% of man-made additions.

Globally, CH$_4$ increased by 7% over the decade from 1983. However, the 1980s were characterized by declining growth rates, dropping from 16 ppbv yr$^{-1}$ to about 10 ppbv yr$^{-1}$. Growth rates slowed dramatically in 1991 and 1992 although very recent data suggest that growth rates started to increase in late 1993. The magnitudes of sinks and especially individual sources of CH$_4$ are less well known than its atmospheric increase. In total, anthropogenic activities are responsible for 60-80% of current CH$_4$ emissions. The direct radiative
forcing due to the increase in CH$_4$ concentration since pre-industrial times is about 0.5 W m$^{-2}$. CH$_4$ has clearly identified chemical feedbacks. The removal process for CH$_4$ is reaction with the hydroxyl radical (OH). Addition of CH$_4$ to the atmosphere reduces the concentration of tropospheric OH concentration, which can subsequently feedback and reduce the rate of CH$_4$ removal.

A recent analysis has shown that these chemical feedbacks result in an adjustment time for the addition of a pulse of CH$_4$ to the atmosphere of 14.5 – 2.5 years. In addition, chemical feedbacks resulting from addition of CH$_4$ increase the concentrations of other greenhouse gases, in particular O$_3$ and stratospheric water vapour; these are indirect effects of CH$_4$ (IPCC, 1994). CH$_4$ is emitted by a large number of sources and total 410-660 Tg CH$_4$ yr$^{-1}$. CH$_4$ is emitted from tropical wetlands (20$^\circ$ N to 30$^\circ$ S) with a global estimate of 60 Tg CH$_4$ yr$^{-1}$. A reevaluation of the ocean source suggests only 3.5 Tg yr$^{-1}$ are emitted by the open ocean, but emissions from CH$_4$-rich areas could be higher, pushing the upper limit of the oceans to 50 Tg yr$^{-1}$. Numerous studies of emissions from rice paddies indicate a global source of 60 Tg yr$^{-1}$, with a range of 20-100 Tg yr$^{-1}$.

**Biological Formation and Consumption of CH$_4$**

Methane producing bacteria (methanogens) are a specialized group of microbes that catabolize a small number of molecules and produce CH$_4$ as the major catabolic product. They are the only known life forms that produce a hydrocarbon as a catabolic product. The known substrates that methanogens can catabolize is very limited: H$_2$ + CO$_2$, formate, acetate, methanol, methylamines and methylsulfides.

Methanogenic decomposition of complex organic matter is possible because many non-methanogenic bacteria convert organic compounds into the few molecules that methanogens can use. The decomposition process in anoxic environments may be divided into three major steps: fermentation (or acidogenesis), syntrophic acetogenesis and methanogenesis (Figure 7).

The presence of electron acceptors (O$_2$, nitrate, Fe$^{3+}$, sulfate etc.) is a major factor affecting CH$_4$ production. CH$_4$ is rarely found in oxic environments and the main reason is probably that O$_2$-using bacteria out-compete methanogens for catabolic substrates. In addition, methanogens are very strict anaerobes, being poisoned by even very small concentrations of O$_2$. This extreme sensitivity to O$_2$ is readily apparent in pure cultures of methanogens, but it may not be important in natural environments where other bacteria remove either O$_2$ itself or toxic reaction products of O$_2$. Like O$_2$, nitrate, ferric sulfate and other electron acceptors tend to block the formation of CH$_4$ by supporting activities of bacteria, which out-compete methanogens for their substrates. Like those of most bacteria, the activities of methanogens increase approximately linearly with temperature except within a few degrees centigrade of the maximum temperature of that organism. The maximum growth temperature of most methanogens is about 37$^\circ$ to 45$^\circ$ C, although many grow up to 65$^\circ$ C.

**Factors Affecting CH$_4$ Transport to the Atmosphere**

Probably, the most important factor affecting the transport of CH$_4$ to the atmosphere is the intervention of aerobic CH$_4$-oxidizing bacteria (Kienne, 1991). The importance of CH$_4$-oxidizing bacteria in diminishing CH$_4$ inputs to the atmosphere may be seen in soils above sanitary landfills, where CH$_4$ diffusing from underlying anaerobic zones enter aerobic zones. The high rates of CH$_4$ oxidation in soils such as these may intercept most or all of the CH$_4$ before it can reach the surface (Whalen et al., 1990). Some mechanisms of transport avoid the activities of methanotrophs almost completely. When CH$_4$ is formed slowly in the deep sediments of an oxic ecosystem, the CH$_4$ that diffuses upwards into the oxic upper sediments may be partly or completely oxidized before it reaches the surface.

On the other hand, if CH$_4$ is formed in the sediments more rapidly than it can leave via diffusion, the CH$_4$ accumulates until its partial pressure is greater than the ambient pressure, and large bubbles can form and rise through the sediments. This CH$_4$ passes through the oxic, methanotrophic sediments rapidly, and because CH$_4$ is poorly soluble, most of the CH$_4$ in the bubbles are oxidized before it reaches the atmosphere (Craig, 1957).

A more important mechanism by which CH$_4$ can elude oxidation in aquatic ecosystems is through plants.
Methane formed in ocean sediments must often pass through an anoxic, sulfate containing zone and then an oxic zone before it reaches the atmosphere. The abilities of O₂-using methanotrophs to oxidize CH₄ is well documented, but has not been confirmed. However, geochemical evidence suggests that CH₄-oxidation occurs in the anaerobic waters of the ocean (Reedurgh and Heggie, 1977). Thus, CH₄ production occurs when organic matter is degraded in environments where light and inorganic electron acceptors such as O₂, Fe³⁺, Mn⁴⁺, NO₃, SO₄ and sulfur are limiting. Methane and CO₂ are the terminal products of metabolism, and these compounds are stable. The contribution of this CH₄ to the atmosphere may be moderated by the activities of the methanotrophic bacteria, which can use O₂ to oxidize CH₄ to CO₂ and H₂O.

**Removal of CH₄**

CH₄ is removed from the atmosphere through reaction with tropospheric OH, about 445 Tg yr⁻¹ and stratospheric removal: 40 Tg yr⁻¹. Another significant sink is microbial uptake in soils, about 15-45 Tg CH₄ yr⁻¹. A growing number of studies show that CH₄ from the atmosphere and also that diffusing upward from deep soils is consumed by microbial communities in upper soils. Methane fluxes at the terrestrial and marine surface are highly variable because they result from the difference of two large processes in situ CH₄ production and consumption. It is noted that this CH₄ oxidation is important in modulating CH₄ emissions from rice paddies, wetlands and landfills. The budget imbalance depends on the recent trends, but sources must have exceeded sinks by about 7-8% over the last decade.

**CH₄ Emission from Wetlands and Oceans**

Wetlands are most likely the largest natural source of CH₄ to the atmosphere (Cicerone and Oremland, 1988), accounting for 20% of the current global emission. Approximately 35% of the global wetland area is broadly distributed in the latitude zone extending from 20°N to 30°S. This region is co-dominated by forested and non-forested swamps and marshes, with a smaller contribution from alluvial or floodplain formations. Some of these tropical wetlands, particularly floodplain habitats, undergo large-scale seasonal expansion and contraction in response to precipitation cycles. Since many of them lie along river courses with exceptionally level topography over large distances, rapid and significant changes in inundation during the year are common.

About half the global wetland area occurs in latitudes from 50° to 70°N regions, expected to undergo temperature increases in the order of several degrees C during the next 100 years (Hansen et al., 1988). These changes could lead to (i) a lengthened thaw season and associated expansion of biological activity, (ii) increased net primary productivity due either to direct fertilization from increases in CO₂ concentration or to indirect temperature effects, and/or (iii) larger areas subjected to thaw and anaerobic conditions. Based on temperature increases alone, CH₄ emissions would be expected to increase in high latitudes providing a positive feedback on the climate system. For example, nutrient limitation may limit primary productivity and microbial adaptation to the current thermal regime may be inelastic. Seasonal water supply is the major controller of tropical and subtropical CH₄ emissions, affecting both area and length of inundation periods in low latitudes. Available water supply may decline in response to increased evaporation under warmer conditions. These changes in water status may produce lowered water tables and dry soil conditions in previously waterlogged or inundated environments, thereby increasing CH₄ oxidation, reducing CH₄ emissions and perhaps causing former wetlands to act as CH₄ sinks.

The contribution of atmospheric CH₄ made by the oceans is not in proportion to the ratio of sea to land. Nevertheless, it is by no means negligible and the oceans occupy the majority of the earth’s surface (73%). Oceanic sources of CH₄ are both natural (physical and biological) and anthropogenic; CH₄ is also generated beneath the soil by both bacterial and geological processes and may be introduced into the oceans by rivers (Scranton and McShane, 1991). Several of the sources that are active in (or beneath) the ocean environment include hydrothermal sources, volcanic and related sediments, hydrocarbon-bearing sediments, marine algae, seepages, and gas hydrates. A case study on the emission of CH₄ from unpolluted and polluted mangrove wetlands of South India is provided later in this chapter.

**Nitrous Oxide (N₂O)**

Soils and oceans are the primary natural source of nitrous oxide (Figure 8). Humans contribute through soil cultivation and use of nitrogen fertilizers, nylon production, and the burning of organic material and fossil fuels. Nitrous oxide (N₂O) is an important, long-lived greenhouse gas that is emitted predominately by biological sources in soils and water, which are not well quantified in global terms. It is removed mainly in the stratosphere by photolysis and reaction with excited oxygen atoms [O (¹D)]. A small soil sink has been suggested. The loss of N₂O in the stratosphere yields
NO\textsubscript{x}, providing the major input of NO\textsubscript{x} in the stratosphere, thus in part regulating stratospheric ozone and influencing the NO\textsubscript{x} balance in the troposphere.

**Figure 8: Molecular structure of nitrous oxide (N\textsubscript{2}O).**

Nitrous oxide (N\textsubscript{2}O) or laughing gas plays a sombre role in the atmosphere, where it can last 180 years. Produced by microbes in soils, its increase comes from chemical fertilizers and slash-and-burn farming as well as by fossil fuel emissions. N\textsubscript{2}O is 200 times as heat absorbant as CO\textsubscript{2}, adding about 5% to the man-made burden.

**Sources of Nitrous Oxide**

Tropical forest soils are probably the single-most important source of nitrous oxide to the atmosphere. New data on tropical land-use changes and intensification of tropical agriculture indicate significant, growing sources of N\textsubscript{2}O (Matson and Vitousek, 1990). The flux of N\textsubscript{2}O depends on the age of the pasture, with young pasture (<10 years) emitting 3-10 times more N\textsubscript{2}O than tropical forests. The total N\textsubscript{2}O sources from tropical soils (forest, savannah) is estimated at 4 Tg(N)/yr (range 2.7-5.7). The magnitude of N\textsubscript{2}O emissions from intensively fertilized tropical agricultural soils has not been quantified. No attempt has been made to speciate the tropical soil source into natural and anthropogenic components. More reliable estimates require better models of N\textsubscript{2}O emissions from soils and better databases of underlying soil and ecosystem properties.

The Earth’s oceans are significant N\textsubscript{2}O sources. N\textsubscript{2}O fluxes from upwelling regions of the Indian (Law and Owen, 1990) and Pacific (Codispoti et al., 1992) Oceans suggest that oceans may be a larger source than previously estimated (1.4-2.6 Tg(N) yr\textsuperscript{-1}; IPCC, 1992). The total pre-industrial N\textsubscript{2}O sources was approximately 9 Tg(N) yr\textsuperscript{-1} of which approximately 3 Tg(N) yr\textsuperscript{-1} was oceanic (Weiss,1994). An isotopic study of atmospheric N\textsubscript{2}O (Kim and Craig, 1993) suggests a large gross flux of N\textsubscript{2}O between the atmosphere and the ocean, but the possible implications for net fluxes are not clear.

Nitrous oxide is also emitted by a large number of smaller sources, most of which are difficult to evaluate. These include soils of other natural ecosystem, biomass burning, degassing of ground water used for irrigation and specialized industrial processes. There are very few studies of these sources, particularly in the tropics, and uncertainties in their emission estimates are large.

**Atmospheric Distribution**

Owing to its long lifetime, N\textsubscript{2}O exhibits only small spatial and temporal variations in the free troposphere. By empirically scaling with CFC interhemispheric gradient, the interhemispheric difference of about 1 ppbv (0.3%) corresponds to a source imbalance with an excess Northern hemispheric source of 5 Tg(N) yr\textsuperscript{-1}.

**Nitrogen Fertilization and Influence of Land Use Changes**

Nitrous oxide (N\textsubscript{2}O) is formed through the microbial transformation of inorganic nitrogen compounds in soils and waters. N\textsubscript{2}O is produced also by combustion of biomass and fossil fuels. In soils, N\textsubscript{2}O (Figure 9) is produced by ‘denitrification’ generally supposed to occur under anoxic conditions. N\textsubscript{2}O (also called ‘laughing gas’) production from anthropogenic sources adds 5% of the enhanced greenhouse effect. Present estimates of sources and sinks of N\textsubscript{2}O are highly uncertain (Papen and Butterbach-Bahl, 1999). The increase of N\textsubscript{2}O in the atmosphere annually by 3-4.5 million tons is well documented and indicates that annual rate of N\textsubscript{2}O transformation into molecular nitrogen is exceeded by this amount. It is beyond doubt that human activities cause a massive input of N into the environment. The influence of land use accounts for 0.5-3.5 t N\textsubscript{2}O-N yr\textsuperscript{-1} or 50% of the total N\textsubscript{2}O-N release from terrestrial ecosystems. The major parts of these emissions are related to fertilization as well as slash and burn practice in the tropics.

**Figure 9: Nitrogen cycle in aquatic ecosystems.**

**Pathways**

1. Uptake of NH\textsubscript{4} or NO\textsubscript{3} by organisms.
2. Release of NH\textsubscript{4} by decomposition.
3 & 4. Microbial oxidation of NH\textsubscript{4} (yields energy in aerobic conditions).
5. Denitrification (NO\textsubscript{3} respiration) by microbes in anaerobic conditions (NO\textsubscript{3} is used instead of O\textsubscript{2} as the terminal electron acceptor during decomposition of organic matter).
6. Nitrogen fixation and 7. NO\textsubscript{3} leaching from soil/sediment.
Climate Change and Coastal Ecosystems: An Overview

Enhanced eutrophication of waters has become a further N\textsubscript{2}O emission source. Ecosystems receive substantial inputs of anthropogenic nitrogen (N), which in many areas can act as a fertilizer and could have increased terrestrial carbon storage by 0.2 to 1.0 GtC yr\textsuperscript{-1} in the 1980s. However, high levels of N-addition are often associated with acidification and high surface O\textsubscript{3} concentrations, which in the long term may damage ecosystems and possibly reduce carbon storage. Our own studies on the effect of N-fertilization in certain mangrove ecosystems of South India (Purvaja and Ramesh, 2000) has highlighted the fact that anthropogenic introduction of N has resulted in ecosystem degradation.

Spatial and temporal variations and the factors influencing primary production have been studied in three different mangrove waters (Pichavaram, Ennore Creek and Adyar Estuary) of South India characterised by different anthropogenic impacts. The gross primary productivity in the unpolluted Pichavaram mangrove was 113 g C m\textsuperscript{-2} yr\textsuperscript{-1} exhibiting natural variability with the environmental forcing factors. Human activities have elevated primary productivity in the Ennore Creek mangrove (157 g C m\textsuperscript{-2} yr\textsuperscript{-1}) primarily through the direct discharge of fertilizer effluents. By contrast, a combination of domestic and industrial effluent discharges into the Adyar estuary mangrove has considerably reduced phytoplankton primary productivity (83 g C m\textsuperscript{-2} yr\textsuperscript{-1}).

The GWP of a greenhouse gas is the ratio of global warming, or radiative forcing—both direct and indirect—from one unit mass of a greenhouse gas to that of one unit mass of carbon dioxide over a period of time. While any time-period can be selected, the 100-year GWPs recommended by the IPCC have been used here in this paper (IPCC, 1994). GWP values are listed below in Table 1.

### Table 1: Global warming potentials (100-year time horizon)

<table>
<thead>
<tr>
<th>Gas</th>
<th>Global Warming Potential (GWP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Carbon dioxide (CO\textsubscript{2})</td>
<td>1</td>
</tr>
<tr>
<td>Methane (CH\textsubscript{4})*</td>
<td>21</td>
</tr>
<tr>
<td>Nitrous Oxide (N\textsubscript{2}O)</td>
<td>310</td>
</tr>
<tr>
<td>HFC-23</td>
<td>11700</td>
</tr>
<tr>
<td>HFC-125</td>
<td>2800</td>
</tr>
<tr>
<td>HFC-134a</td>
<td>1300</td>
</tr>
<tr>
<td>HFC-143a</td>
<td>3800</td>
</tr>
<tr>
<td>HFC-152a</td>
<td>140</td>
</tr>
<tr>
<td>HFC-227ea</td>
<td>2900</td>
</tr>
<tr>
<td>HFC-236fa</td>
<td>6300</td>
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<tr>
<td>C\textsubscript{2}F\textsubscript{6}</td>
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</tr>
<tr>
<td>C\textsubscript{4}F\textsubscript{10}</td>
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</tr>
<tr>
<td>C\textsubscript{6}F\textsubscript{14}</td>
<td>7400</td>
</tr>
<tr>
<td>SF\textsubscript{6}</td>
<td>23900</td>
</tr>
</tbody>
</table>

* The methane GWP includes the direct effects and those indirect effects due to the production of tropospheric ozone and stratospheric water vapour. The indirect effect due to the production of CO\textsubscript{2} is not included.

Source: IPCC, 1996.
Human Factors Affecting Climate
Naturally occurring greenhouse gases (e.g. water vapour, carbon dioxide, methane, nitrous oxide) keep the Earth warm enough to support life. Scientific studies show that a variety of human activities release greenhouse gases. These include the burning of fossil fuels for producing electrical energy, heating and transportation. By increasing their concentrations and by adding new greenhouse gases like CFCs, humankind is capable of raising the average global temperature.

Land-Use Change: As humans replace forests with agricultural lands or natural vegetation with asphalt or concrete, they substantially alter the way the Earth’s surface reflects sunlight and releases heat. All these changes also affect regional evaporation, runoff and rainfall patterns.

Atmospheric Aerosols: Humans are adding large quantities of fine particles (aerosols) to the atmosphere, from both agriculture and industrial activities. Although most of these aerosols are soon removed by gravity and rainfall, they still affect the radiation balance in the atmosphere. Whether this effect adds to or offsets any warming trend depends on the quantity and nature of the particles as well as the nature of the land or ocean surface below. The regional effects, however, can be significant.

Climate Trends
An analysis of temperature records shows that the Earth has been warmed up by an average of 0.6°C over the past 100 years. There appears to have been a warming until the early 1940’s, then a moderate cooling until the mid 1970’s, followed by a renewed and pronounced warming continuing through the present. The 1990’s were the warmest decade on record. (In fact, ice core and other proxy data indicates that the 1980’s and 1990’s were the warmest decades of the past millennium.) Nighttime temperatures over land have generally increased more than daytime temperatures. Regional changes are also evident. For example, recent warming has been greatest over the mid-latitude continents in winter and spring, with a few areas of cooling such as the North Atlantic Ocean. Precipitation has increased over land in high latitudes of the Northern Hemisphere, especially during the cold season.

This is consistent with predictions of climate change due to an enhanced greenhouse effect and increased aerosols. Yet, it could also be within acceptable limits for natural temperature variation.

Temperature
Global temperatures are rising. Observations collected over the last century suggest that the average land surface temperature has risen by 0.45-0.6°C (0.8-1.0°F) in the last century (Figure 10).

![Figure 10: Global temperature changes.](image)

Precipitation
Precipitation has increased by about 1 percent over the world’s continents in the last century. High latitude areas are tending to see increases that are more significant in rainfall, while precipitation has actually declined in many tropical areas.

Sea Level Rise (SLR)
Sea level has raised worldwide approximately 15-20 cm (6-8 inches) in the last century. Approximately 2-5 cm (1-2 inches) of the rise has resulted from the melting of mountain glaciers. Another 2-7 cm has resulted from the expansion of ocean water that resulted from warmer ocean temperatures. The rise and fall of sea level has occurred on many occasions in the past over the long periods of time due to changes in climate, which influences geological, biological and hydrological processes that affect sea level. The natural phenomena that affect sea level are the following:

- Natural variability (i.e., the geomorphology of the coast)
- Vertical land movements (such as oceanic lithosphere cooling)
- Tectonics
- Meteorological factors such as
  - Changes in weather system leading to changes in pressure and wind fields
  - Specific events such as cyclones
  - Large scale events such as El Niño

In addition, recent increases in anthropogenic activities have also contributed to sea level rise, which include:

- Global warming due to increasing greenhouse gas concentrations
• Land subsidence due to extraction of oil, gas and water, mining of sand, shells etc.
• Changes in land-use such as bay reclamation, impoundments of water by dam construction.

While sea level undergoes short-term and long-term changes, land levels are also subjected to vertical movement. There are three groups of factors which determine relative changes between the levels of oceans and those of land areas (Sterr, 2001).

• Changes in the size and holding capacity of the ocean basin. Ocean volume changes of this sort result from large-scale plate tectonic movements of the Earth’s crusts as well as from the infill of sediment into ocean basins. Both processes operate rather slow, however, on time scales of millions of years and therefore have no discernable effects on Holocene to recent sea level adjustments.

• Small-scale or large-scale vertical crustal movements at continental rims. These may result from two different processes: On a smaller scale regional tectonics related to earthquakes and volcanism can cause uplift or subsidence of land areas bordering the sea. On a larger scale crustal deformations result mainly from glacio-hydro-isostatic effects.

• Changes in volume and density of water masses in the ocean basins. First, temperature and salinity near the ocean surface directly determine the water density, thus affecting the upper boundary of the ocean surface. Secondly and more important, the shifts in the global water balance change the available water masses in the oceans. These shifts are brought about by the storage or release of water in ice sheets, in permafrost and groundwater layers and in artificially dammed lakes or reservoirs.

As vertical movement both of land and sea levels may occur, there is possibility of either an increased rate of SLR, where higher sea levels are facing subsiding land areas; on the other hand, when the uplift on the land side partially compensates or even outgrows sea level rise, this reduces the regional rate of SLR and may result in ‘negative SLR’ i.e., a net drop over time. Combining all the possible influences renders the assessment of SLR a rather complex and difficult task, which leaves a range of uncertainties about past and future climate impacts on sea level. The major factors causing a distinctive relative sea level rise are the following:

• Melting of land-bound ice sheets due to a global temperature increase, caused by natural climate variability and/or an anthropogenically enhanced greenhouse effect. By this, large volumes of water may be added to the oceans, resulting in the so-called “eustatic effect”
• Enhanced melting of mountain glaciers and ice caps as a result of eustatic effect
• Increasing thermal expansion of the upper ocean layers as a result of higher sea surface temperatures (“steric effect”)
• Widespread effect of the upper Earth’s crust adjacent to the regions that were covered by thick Pleistocene ice sheets (“glacio-isostatic effect”)

Late-Quaternary subsidence movements in areas covered by large masses of unconsolidated sediments, usually due to progressive sedimentary compaction (“geologic effect”). This type of relative subsidence is prevalent in deltaic environments along shallow continental margins e.g., the deltas of Mississippi, Nile, and Ganges.

References


