CHAPTER III

CLIMATIC TEMPERATURE CHANGE THEORIES AND CALCULATION

After considering the Earth as the control volume, we then apply the chemical engineering concepts for developing the modified atmospheric-oceanic model to simulate the global climate change. This chapter is mentioned about the cause of this change. Moreover, the elevation of average global temperatures at sea-surface from preindustrial temperature caused by the increase of 4 kinds of greenhouse gases is investigated. Method and procedure of calculation is also discussed herein.

3.1 Why does Climate Change ?

The Earth's atmosphere is a thin and seemingly fragile skin of air protecting the planet from the harshness of space. From the ground, this perspective is harder to appreciate. The atmosphere protect us from the sun's most harmful radiation and, provide a moderate and stable climate, and cleans itself to provide fresh air to breathe.

Many atmospheric gases in atmosphere have properties that they can store a lot of heat energy due to its high heat capacity. They are transparent to incoming solar radiation and absorb and reemit the longer-wavelength radiation emitted from the Earth's surface. By trapping heat in this manner, so-called greenhouse gases cause the lower part of the Earth's atmosphere (the troposphere) to become warmer. A rapid increase in these gases concentration

(see Table 3.1) over the past century with the growth in human population and industrial activity, may alter atmospheric heating rates, resulting in global warming and other climatic disturbances.

The temperature of the Earth's is determined by the balance between the rate at which sunlight reaches the Earth's surface and the rate at which the warmed Earth sends infrared radiation back into space. It is well established that the warm temperatures (which make life on Earth possible) are the direct result of the trapping of part of the Earth's radiant heat.

Table 3.1 Atmospheric Concentration of Greenhouse Gases (World Resources, 1992-93)

Year	CO_2 [ppm]	CH_4 [ppb]	N_2O [ppb]	CFC-11 [ppt]	CFC-12 [ppt]
Preindustrial age	280.0a	700a	285.0a	0	0
1959	315.8	x	x	x	x
1960	316.8	x	x	x	x
1961	317.5	x	x	x	x
1962	318.3	1,354	x	x	x
1963	318.8	x	x	x	x
1964	x	x	x	x	x
1965	319.9	1,386	x	x	x
1966	321.2	1,338	x	x	x
1967	322.0	1,480	x	x	x
1968	322.8	1,373	x	x	x
1969	323.9	1,385	x	x	x
1970	325.3	1,431	x	x	x
1971	326.2	1,436	x	x	x

Table 3.1 Atmospheric Concentrations of Greenhouse Gases (Continue)

Year	CO_2 [ppm]	CH_4 [ppb]	N_2O [ppb]	CFC-11 [ppt]	CFC-12 [ppt]
1972	327.3	1500	x	x	x
1973	329.5	1,624	x	x	x
1974	330.1	1,596	x	x	x
1975	331.0	1,541	291.4	120	200
1976	332.0	1,490	293.3	133	217
1977	333.7	1,471	294.6	148	239
1978	335.3	1,531	296.4	159	266
1979	336.7	1,545	296.3	167	283
1980	338.5	1,554	297.6	179	307
1981	339.8	1,569	298.5	185	315
1982	341.0	1,591	301.0	193	330
1983	342.6	1,615	300.9	205	350
1984	344.3	1,629	300.4	213	366
1985	345.7	1,643	301.5	223	384
1986	347.0	1,656	302.5	232	404
1987	348.8	1,667b	304.5	244	410
1988	351.4	1,681b	306.3	256b	416b
1989	352.8	1,694b	x	265b	433b
1990	354.0	1,704b	x	275b	466b

Note: a = approximately

b = preliminary data, previous years are not calibrated to the same standard. All estimated are by volume.

x = not available

Gases which exist in the Earth's atmosphere has different contribution to a global climate change. Carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O) and chlorofluorocarbon (CFCs) are now considered as the greenhouse gases causing a large change in the global temperature.

3.2 Focus on Greenhouse Gas Emission

It has been proved conclusively that, in recent millennia, the "greenhouse effect" of the atmosphere has kept the planet at an average temperature of about $15^\circ C$, or 33 K warmer than it otherwise would be.

At this time, CO_2 is considered as the majority of the greenhouse effect, but the combined effect of other trace gases (especially methane, nitrous oxide and chlorofluorocarbon) could soon equal or exceed the potential warming from CO_2 alone. The addition of an incremental molecule of methane, for example, will result in the trapping of 20-30 times as much heat as the addition of an incremental CO_2 molecule. An incremental CFC molecule will result in the trapping of 20,000 times as much heat as would a CO_2 molecule. The heat-absorbing potential of these gases varies widely and depends on such factors as their relative concentration in the atmosphere and their infrared absorption profile. As mentioned above, not only carbon dioxide cause global warming, but also methane, nitrous oxide, and chlorofluorocarbon can result to increasing climatic temperature of the Earth. In order to get more precise results for prediction the sea-surface temperature, this research intends to use a model which considers all four greenhouse gases that discussed above.

3.2.1 Greenhouse Gases

Concentrations of infrared-absorbing gases are increasing at unprecedentedly rapid rates. Table 3.2 illustrates what is known about the growing atmospheric concentrations of the major greenhouse gases. While recent growth rates in the concentrations of individual gases vary, they all have increased significantly since preindustrial times.

Table 3.2 Atmospheric Growth of Greenhouse Gases (World Resources, 1990-91)

Greenhouse gas	Concentration		Present annual increase	
	Preindustrial	1989	Rate	Percent
CO_2	275.00 ppm	352.8 ppm	1.70 ppm	0.4
CH_4	0.70 ppm	1.7 ppm	12.30 ppm	0.7
N_2O	280.00 ppb	306 ppb	0.600 ppb	0.2
CFC-11	0.00	0.27 ppb	0.010 ppb	3.4
CFC-12	0.00	0.43 ppb	0.025 ppb	5.3

Carbon dioxide (CO_2), which now accounts for about half the greenhouse effect, is emitted to atmosphere by natural and anthropogenic processes: volcanic eruptions, fossil and biomass fuel combustion, gas flaring, cement production, forest clearing, soil disturbance, etc. The CO_2 concentration of 351 parts per million (ppm) in 1988 is 20-25 percent higher than at any time in the past 160,000 years (World Resources, 1990-91). By the beginning of the Industrial Revolution, CO_2 concentrations had climbed back up to 280 ppm. Within the past 100 years, however, atmospheric CO_2 concentrations have risen by another 70 ppm. More than half of this 25 percent increase has occurred in the past 30 year alone.

Methane (CH$_4$) is emitted in a variety of natural anoxic processes, including enteric fermentation in ruminants and anaerobic decomposition in wetlands, bogs, tundra, and lake. Emission sources associated with human activities include livestock management, anaerobic decomposition in rice paddies, and combustion of fossil fuels and biomass (fuelwood and clear forests). Ice core records of methane levels show that concentrations of this gases remained constant for thousands of years until about a century ago when they began to rise; concentrations of methane doubled higher than their preindustrial level (World Resources, 1988-89). Its greenhouse impact is one third that of carbon dioxide.

Nitrous oxide (N$_2$O) is emitted both by aerobic decomposition of organic matter in oceans and soils by bacteria and by combustion of fuels and biomass (fuelwood and cleared forests). Ice core records of nitrous oxide levels show that concentrations of this gas remained constant for thousands of years until about a century ago when they began to rise (World Resources, 1988-89). The concentrations of nitrous oxide are 8 percent higher than their preindustrial level. It may contribute one-twelfth the greenhouse impact of carbon dioxide.

Chlofluorocarbons (CFCs), which are entirlely manmade, may also contribute one fourth the greenhouse impact of carbon dioxide. Emissions of chlorofluorocarbons are estimated based on use-specific release factors derived from CFC production and sales data. CFCs are predominantly used as refrigerants, foams, and propellants.

3.2.2 Global Warming Potential (GWP) Index

The pattern of emissions of different greenhouse gases varies from kind to kind. To describe quantitatively the effects of such emissions on the atmosphere, it is useful to calculate a "Global Warming Potential" (GWP). The work of the scientific assessment panel of the Intergovernmental Program on Climate Change (IPCC) has set a de facto standard for discussion of greenhouse gas emission. The panel adopted a conceptual unit, the "global warming potential" (GWP), for comparing the impact of gases that have difference lifetimes in the atmosphere and difference potencies in absorbing heat.

One complication inherent in such a comparison is that the potential of a greenhouse gas depends on its concentration in the atmosphere, which in turns depends on assumptions about future emissions. A second complication is that the lifetimes of the greenhouse gases (or conversely, the rate at which they are removed from the atmosphere by natural processes) is not known with any precision, at least for carbon dioxide (CO_2). Estimation of the removal rates are based on models of atmosphere, oceanic, and biospheric processes.

The GWP is determined by integrating an expression for the removal rate times an expression for the infrared absorption potency of the gas. The GWP value depends on the period of years over which the integration is carried out, and can be chosen arbitrarily. The IPPC panel reported values for 20, 100 and 200 years. The GWP reported in this research makes used of an integration period of 100 years. The values calculated by the IPCC panel are normalized so that the GWP of carbon dioxide is 1. The GWP index score is calculated as follows:

$$GWP_i = \frac{\int_0^N r_i \, c_i \, dt}{\int_0^N r_c \, c_c \, dt} \qquad (3\text{-}1)$$

where

i	=	Gas i
N	=	Integrating period (or time horizon)
r_i	=	Radiative forcing of gas i
r_c	=	Radiative forcing of carbon dioxide
c_i	=	Atmospheric concentration of gas i
c_c	=	Atmospheric concentration of carbon dioxide

The values of GWP for the common greenhouse gases are summarised in Table 3.3. Although CO_2 is the least effective greenhouse gas as considered per a unit emission, the overall radiative effect of CO_2 become highest because it depends on both the GWP and the level of emissions. For example, although the GWP of CO_2 is only around 0.00018 that of CFC, current emissions of CO_2 are 52,000 times as large, so that the effect of CO_2 is nearly ten times of CFC.

Table 3.3 Greenhouse Effects of various Greenhouse Gases (Dornbusch, R., and Poterba, J.M., 1991)

Greenhouse gas	Relative change in forcing	Global warming potential (time horizon, years)		
		20	100	200
CO_2	1	1	1	1
CH_4	58	63	21	9
N_2O	206	270	290	190
CFC	4,860	5,800	5,400	3,000

3.3   Modified Atmospheric-Oceanic Model

After using the conservation law of energy and defining the essential assumptions for both of the atmospheric model and the oceanic model, we can develop this modified model. The characteristics of these two components of the control volume and the method for calculating the sea-surface temperature is described as follows:

3.3.1   Ocean Dynamics

The ocean system, which covers about three fourths of the planet, is the greatest single climate-feedback system. The ocean can act initially as a buffer for global climate change because it can be considered as both a heat sink and, possibly, as a sink for some CO_2 emissions. As the climate becomes warms, some of the excess heat produced by greenhouse processes is siphoned off to warm the enormous bulk of ocean waters. Thus, the full effect of greenhouse warming is not manifest at first in the atmosphere. Thermal inertia of the oceans could be explained why the global warming observed to date is only about half of the $1.0\,^{\circ}C$ warming while climate models predict the warming should have occurred already on the basis of present levels of greenhouse gases.

The mixing of the ocean waters and their global circulation by ocean currents are key processes governing how the ocean becomes warm and transports heat. Heat transport by the oceans plays a crucial role in shaping details of the world's climate. The Earth is heated unevenly by the sun, with the tropics absorbing more solar energy than the polar regions. Ocean currents provide an important means to redistribute this heat. Indeed, estimates are that the oceans account for 40 percent of the total heat transported from low to high

latitudes in the Northern Hemisphere (the rest being transported through the atmosphere by weather systems) (World Resources, 1990-91).

3.3.2 Atmospheric-Oceanic Interaction

It is useful at times to take a somewhat distant view of the Earth and to describe it in the way one would describe another planet. From this point of view, the surface of the Earth is dominated by existence of two fluid envelops: a gaseous one, the atmosphere, that covers the entire Earth and a liquid one, the ocean, lying below the atmosphere and covering nearly three quarters of the surface. As the scientists have known so far, none of the other planets has oceans or large reserves of liquid water in any form.

Fundamental Characteristics of Oceans and the Atmosphere

A most important characteristics of the atmosphere is its transparency. A large fraction-about half-of the incoming radiation from the sun passes through the atmosphere and reaches the Earth's surface, most of which is ocean. All the heat received by the Earth from the sun is again radiated away to space. However, the atmosphere is not transparent to the infrared radiation produced by the Earth. (The radiative transparency to incoming radiation and the relative opacity to outgoing radiation create the natural greenhouse effect without which the planet would be icebound).

Weather and climate would both exist in the absence of the ocean. However, the ocean dramatically modifies the situation. The most important way it does so is through evaporation at the ocean surface. When water evaporates, it absorbs the latent heat of vaporisation (about one megajoule, or one third of a

kilowatt hour per kilogram). This heat is released again when the vapour condenses into clouds. The clouds may evaporate and recondense several times, but eventually they precipitate. When they do, the water returns to the surface and the heat remains aloft. This evaporation-precipitation process thus transports heat both vertically and geographically.

The second central characteristics of the ocean is its ability to store heat. The top three meters of the ocean have as much thermal capacity (i.e., can store as much heat energy) as the entire atmosphere. Thus the ocean is able to absorb and to give up great quantities of heat with little change in temperature. This capability has dramatic effects upon climate. For example, in areas where the air is dry, the day-night temperature difference can be $40^\circ C$. Over the ocean, the difference is generally only a few degrees or less at the surface. At middle latitudes in mid-continent, the temperature difference between the summer maximum and winter minimum can reach $80^\circ C$. Over the ocean, the spread rarely excess $10^\circ C$. As a result, to create a model for estimating the climatic temperature change have to consider the ocean heat capacity (3×10^8 $Ws/m^2 K$). This research will try to calculate to find out the increasing mean climatic temperature of the Earth at sea level.

The third important characteristics of the ocean is its ability to transport heat. The best-known and one of the most important examples in the North Atlantic circulation. Water heated by the sun in tropical and subtropical Atlantic is transported northward in the Gulf Stream and then distributed over the entire North Atlantic, some proportion spreading as far north as the Norwegian Sea. The return-flowing water, much of it at great depths, is much colder. The resulting heat transfer, quantitatively about one petawatt (1 quadrillion watts), is a significant fraction of the north-south heat transfer effected by the atmosphere

itself. The prevailing westerly winds at the latitude of the North Atlantic transport this heat across all of Europe, producing a significantly warmer climate there than would be anticipated from its latitude.

Atmospheric-Oceanic Coupling

Although the ocean strongly influences the atmosphere, it is the atmosphere that largely drives ocean movements. The ocean is also a heat engine, but a relatively inefficient one; the temperature range over which it works is much smaller than that of the atmosphere. The deep circulation of the ocean, below about 100 meters, is largely caused by heating and cooling and by changes in salinity produced by evaporation and precipitation. However, the more vigorous surface motions, which are responsible for much of the heat transport, are wind driven.

The way in which the ocean responses to surface wind is greatly complicated by the fact that the Earth is a rotating spherical body. Although the details are beyond the scope of this discussion, what is most important to note is that the ocean-atmosphere system is closely coupled. The distribution of sea-surface temperature has important effects on activity in the atmosphere, and wind driving is a dominant factor in setting up ocean currents, which in turn determine the distribution of sea-surface temperature. Ideally, this system would be described and studied through a couple ocean-atmosphere model, but today the remains a long-term scientific goal.

Couples analysis of the two systems is further complicated by the fact that they operate on difference time scales. Atmospheric fluctuations are typically on the order of a few days, which is much too fast a time scale to be

followed by the ocean. Oceanic fluctuations generally take place over several months.

3.3.3 Modified Atmospheric-Oceanic Model

Comprehensive global monitoring does not mean just more and better physical observations, but improved methods for using these observations to understand global cycle and to predict how these cycles will behave in the future. An accurate, quantitative model of a natural system is a powerful tool that allows observations of the system's present state to be used to forecast the future state of the system.

For global climate model, which is used to study climate changes induced by the release of greenhouse gases into the atmosphere, makes use of the mathematical equations to simulate the dynamics of the atmosphere. Climate models, however, are generally more complex and involve a wider range of processes than weather forecasting models. The fundamental difference in how climate models and weather models behave stems from the timescales over which they are required to act. Weather models simulate atmospheric dynamics for short periods of time-usually one to 10 days. But a typical run of a climate model exploring the greenhouse effect is 10 to 100 years and simulations of ancient Earth climates may treat time periods of thousands of years.

Models to better understand future climate resulting from changing atmospheric composition have developed over the last two decades. The climate system involves transfer of energy between a three-dimensional turbulent and radiatively active atmosphere and spatially heterogeneous land, ocean, and cryosphere surfaces. This system is very complex and it has not yet been

possible to produce models that use "state-of-the-art" descriptions of either the atmospheric or surface processes. Furthermore, our state-of-the-art understanding of some processes is still not satisfactory. Thus, various approximations and simplifications have been made to develop climate models, some based on empiricism. Because of lack of consensus as to which approximations do least damage to modeling results, many different modeling approaches have been developed. The simple models still remain of interest because their relative computational economy allows consideration of a much wider range of parameter values. Furthermore, the simpler models can usefully provide insight and description of the dominant processes in the more elaborate models. That is, they are diagnostic and education. As discussed, the ocean has a strong relation to climate system but can be explained by simple model. The model used in this research, consists of atmospheric model and oceanic model, has been modified from the models of *Oeschger et al.* [1975] and *Cess et al.* [1981] as list below:

Oeschger et al. [1975] proposed a box-diffusion model for studying the carbon cycle, which describes the world ocean in terms of a well-mixed surface layer (about 75 m deep) and a deep sea reservoir, within which vertical transfer is accomplished by eddy diffusion. The rate of air-sea exchange is given in accordance with measurements and the eddy diffusivity (thermal diffusivity) is assigned a value of $1.3 \text{ cm}^2\text{s}^{-1}$, which yields a vertical distribution.

Cess et al. [1981] proposed an ocean-land global climate model, to estimated time-dependent global warming due to increasing levels of atmospheric carbon dioxide. Ocean heat capacity is incorporated by means of a global ocean model having a 70 m deep mixed layer, with heat being transports from the mixed layer to deeper waters by eddy diffusion (conduction). The time-

dependent increase in atmospheric CO_2, from 1860 to 2025, is taking from carbon-cycle models. The model results suggest that ocean heat capacity will produce a lag in CO_2-induced global warming of about 2 decades.

According to above discussions, it is clear that ocean has great effect to climate system, especially to delay the temperature changes. Researchers mentioned above have been calculated the increasing of Earth's climatic temperature which is resulted from CO_2 only. However, other greenhouse gases also affect the global warming too. From this reason, it is expect to cause the lower estimation of the climatic temperature change. Therefore, this research includes other greenhouse gases for consideration by using the modified atmospheric-oceanic model. Moreover, besides the prediction at an interval of times, average temperature of the Earth can be calculated when the value of greenhouse gases in the atmosphere was clarified. The detail of model was as followed:

Oceanic Model

Figure 3.1 illustrates the presently employed global ocean model, which is equivalent to the box-diffusion model utilised by *Oeschger et al.* [1975] in studying the carbon cycle. This consists of a mixed layer having the heat capacity $R_m = 3 \times 10^8$ Ws/m^{2}K, which corresponds to a layer 70 m deep. By definite $R_m = \rho C_p \Delta Z$, where ρ is the density of sea water, C_p is its specific heat of the ocean, and $\Delta Z = 70$ m.
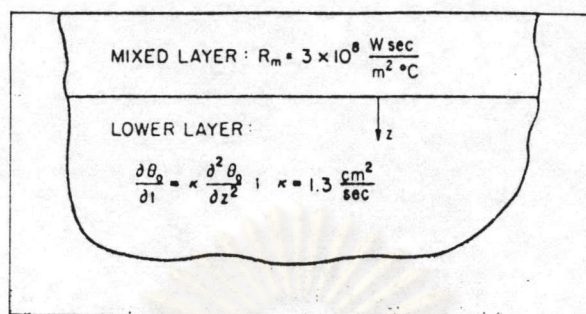
Figure 3.1   Schematic Illustration of the Global Oceanic Model

Below the mixed layer heat is transported by turbulent diffusion, with κ representing a global eddy diffusion coefficient (thermal diffusivity). $\theta_o(Z,t)$ is denoted as the ocean temperature within the lower layer relative to the 1860's, and it is a function of depth and time as derived in chapter II.

$$\frac{\partial \theta_o(z,t)}{\partial t} = \kappa \frac{\partial^2 \theta_o(z,t)}{\partial z^2} - w \frac{\partial \theta_o(z,t)}{\partial z} \tag{3-2}$$

Change in the temperature depends on both variables, therefore this model equation is classified as partial differential equation type. In general, a partial differential equation can be replaced by two ordinary differential equations using the method of separation of variables. By defining the initial condition and boundary conditions, solutions of the partial differential equation can be obtained. However, in this case, change of temperature is still dependent on time, therefore the separation method is not suitable to the modified atmospheric-oceanic model equation derived above.

Cess and Goldenberg [1981] purpose that at any depth of the lower layer ocean, the oceanic temperature increases with time by exponential ratio as follows:

$$\theta_o(z,t) = (c)\exp\left(\frac{t}{\tau_c}\right) \tag{3-3}$$

where

τ_c = time scale for CO_2 heating process (33 years)

From Equation (3-3), differentiate the average oceanic temperature with time (t)

$$\frac{\partial\theta_o(z,t)}{\partial t} = c\frac{\partial\exp(t/\tau_c)}{\partial t} = \frac{c}{\tau_c}\exp(t/\tau_c) = \frac{\theta_o(z,t)}{\tau_c} \tag{3-4}$$

Substitute Equation (3-4) into Equation (3-2), we obtain

$$\frac{\theta_o(z,t)}{\tau_c} = \kappa\frac{\partial^2\theta_o(z,t)}{\partial z^2} - w\frac{\partial\theta_o(z,t)}{\partial z} \tag{3-5}$$

or

$$\kappa\frac{\partial^2\theta_o(z,t)}{\partial z^2} - w\frac{\partial\theta_o(z,t)}{\partial z} - \frac{\theta_o(z,t)}{\tau_c} = 0 \tag{3-6}$$

Solution of above equation, simple second order differential equation, is

$$\theta_o(z,t) = c_1\exp(\lambda_1 z) + c_2\exp(\lambda_2 z) \tag{3-7}$$

where

$$\lambda_1 = \frac{1}{2\kappa}\left[w + \sqrt{(-w)^2 + \frac{4\kappa}{\tau_c}}\right] = \frac{w}{2\kappa} + \frac{1}{2}\sqrt{\frac{(-w)^2}{\kappa^2} + \frac{4}{\kappa\tau_c}} \tag{3-8}$$

and

$$\lambda_2 = \frac{1}{2\kappa}\left[w - \sqrt{(-w)^2 + \frac{4\kappa}{\tau_c}} \right] = \frac{w}{2\kappa} - \frac{1}{2}\sqrt{\frac{(-w)^2}{\kappa^2} + \frac{4}{\kappa\tau_c}} \tag{3-9}$$

Because the oceanic depth is relatively deep, and the temperature of the ocean must be finite, therefore the constant c_1 in Equation (3-7) must be equal to zero.

$$\theta_o(z,t) = c_2 \exp(\lambda_2 z) \tag{3-10}$$

at $z = 0$

$$c_2 = \theta_o(0,t) = \theta_o(t) \tag{3-11}$$

Substitute constant c_2 from Equation (3-11) into Equation (3-10)

$$\theta_o(z,t) = \theta_o(t)\exp(\lambda_2 z) \tag{3-12}$$

From the above equation, whether we consider the heat convection or not, the average lower oceanic temperature change is not affected too much. In this work, the energy transferring to the lower oceanic layer is transferring by heat conduction only. The correlation of the average lower layer oceanic temperature change with respect to time is described by equation (3-13).

$$\frac{\partial \theta_o(z,t)}{\partial t} = \kappa \frac{\partial^2 \theta_o(z,t)}{\partial z^2} \tag{3-13}$$

where

$$\theta_o(z,t) = \theta_o(t)e^{-\frac{z}{\kappa\tau_c}} \tag{3-14}$$

From the above equation, differentiating the lower oceanic temperature change with oceanic depth (z), will provide

$$\frac{\partial \theta_o(z,t)}{\partial z} = -\frac{\theta_o(t)}{\sqrt{\kappa \tau_c}} e^{-\frac{z}{\sqrt{\kappa \tau_c}}}$$

(3-15)

As discussed, the temperature gradient along z-axis, at interface of the mixed layer and lower layer, depends on the heat transferring to the lower layer, $F_o(t)$, as shown in chapter II.

$$\left(\frac{\partial \theta_o(z,t)}{\partial z} \right)_{z=0} = -\frac{F_o(t)}{\kappa \rho C_p}$$

(3-16)

In this work, we consider only the temperature at sea-surface, therefore, let z = 0 in Equation (3-15), and compare Equation (3-15) to Equation (3-16), we will obtain

$$\frac{\partial \theta_o(t)}{\partial z} = -\frac{\theta_o(t)}{\sqrt{\kappa \tau_c}} = -\frac{F_o(t)}{k \rho C_p}$$

(3-17)

Rearranging Equation (3-17), we can estimate the average global temperature change as function of heat flux transferring to lower layer.

$$\theta_o(t) = F_o(t) \frac{\sqrt{\kappa \tau_c}}{k \rho C_p}$$

(3-18)

Heat flux transferring to the lower layer derived from the law of energy conservation at the interface of the mixed layer and the lower layer as discussed in chapter II is

$$F_o(t) = \Delta F(t) - B\theta_o(t) + \frac{\upsilon}{f_o}[\theta_{LS} - \theta_o(t)] - R_m\left\{\frac{d\theta_o(t)}{dt}\right\} \qquad (3\text{-}19)$$

where

$\Delta F(t)$ = Heat flux of the surface-troposphere system due to increasing

atmospheric greenhouse gas (W/m^2)

B = Feedback parameter (1.26 W/m^{2}K)

υ = The overall heat transfer coefficient $(0,\infty)$

υ/f_o = Heat transfer coefficient based on ocean

f_o = The global ocean fraction (0.71)

R_m = The heat capacity of mixed layer (3x10^8 Ws/m^{2}K)

θ_{LS} = Land-surface temperature change relates to the 1860's (K)

$\theta_o(t)$ = Sea-surface temperature change relates to the 1860's (K)

The term $\Delta F(t)$ represents heating of the surface-troposphere system due to increasing atmospheric greenhouse gases. The second term represents modification of the surface-temperature radiation budget resulting from increased sea-surface temperature, where

$$B = \frac{dF_{LW}}{d\theta_o(t)} + Q\frac{d\alpha}{d\theta_o(t)} \qquad (3\text{-}20)$$

with $dF_{LW}/d\theta_o(t)$ representing the change in outgoing infrared flux with surface temperature, Q is the global insolation, $d\alpha/dT_s$ is the change in albedo (reflectivity) due to ice-albedo feedback. The third term on the rightside of (3-19) represents heat transfer from land to ocean surfaces, with υ denoting a land-ocean coupling coefficient or overall heat transfer coefficient. The final term represents heat storage by the mixed layer. In this work, we will consider only

the extreme cases of complete land-ocean heat transfer (υ=infinite) and no land-ocean heat transfer (υ=zero).

case 1 complete land-ocean heat transfer

From Equation (3-19), applies to land surface, such that

$$0 = \Delta F(t) - B\theta_o(t) + \frac{\upsilon}{f_o}\left[\theta_{LS} - \theta_o(t)\right] \qquad (3\text{-}21)$$

for which heat storage by the land is assumed to be negligible compared with that by the ocean. When heat transfer between land and ocean is completed, the temperature change over land surface is equal to the sea-surface temperature change.

$$\theta_{LS} = \theta_o(t) \qquad (3\text{-}22)$$

Substituting Equation (3-22) into Equation (3-21), we will obtain

$$0 = \Delta F(t) - B\theta_o(t) \qquad (3\text{-}23)$$

In the case of complete land-ocean heat transfer, the considered heat transfer area is based on that of ocean, therefore

$$0 = \Delta F(t) - B\theta_o(t) = \frac{\Delta F(t)}{f_o} - \frac{B\theta_o(t)}{f_o} \qquad (3\text{-}24)$$

Replace Equation (3-22) and Equation (3-24) into Equation (3-19), it follows that

$$F_o(t) = \frac{\Delta F(t)}{f_o} - \frac{B\theta_o(t)}{f_o} - R_m \left(\frac{d\theta_o(t)}{dt} \right) \qquad (3\text{-}25)$$

Substitute heat flux to lower layer from the above equation into Equation (3-18), we will obtain the relation between the sea-surface temperature change with increasing heat flux of the surface-troposphere system, $\Delta F(t)$, and the differential equation of sea-surface temperature change with time as follows:

$$\theta_o(t) = \left(\frac{\frac{\sqrt{\kappa \tau_c}}{f_o \kappa \rho C_p}}{1 + \frac{B\sqrt{\kappa \tau_c}}{f_o \kappa \rho C_p}} \right) \Delta F(t) - \left(\frac{\frac{R_m \sqrt{\kappa \tau_c}}{\kappa \rho C_p}}{1 + \frac{B\sqrt{\kappa \tau_c}}{f_o \kappa \rho C_p}} \right) \frac{d\theta_o(t)}{dt} \qquad (3\text{-}26)$$

case2 no land-ocean heat transfer

From equation of heat flux transferring to the lower layer, substituting overall heat transfer coefficient ($\mathbf{V}$) with zero will provide

$$F_o(t) = \Delta F(t) - B\theta_o(t) - R_m \left(\frac{d\theta_o(t)}{dt} \right) \qquad (3\text{-}27)$$

Replacing above equation into Equation (3-18), we will obtain the relation between the sea-surface temperature change with $\Delta F(t)$, and differential equation of sea-surface temperature change with time as Equation (3-28).

$$\theta_o(t) = \left(\frac{\frac{\sqrt{\kappa \tau_c}}{\kappa \rho C_p}}{1 + \frac{B\sqrt{\kappa \tau_c}}{\kappa \rho C_p}} \right) \Delta F(t) - \left(\frac{\frac{R_m \sqrt{\kappa \tau_c}}{\kappa \rho C_p}}{1 + \frac{B\sqrt{\kappa \tau_c}}{\kappa \rho C_p}} \right) \frac{d\theta_o(t)}{dt} \qquad (3\text{-}28)$$

To estimate the sea-surface temperature change at any time, we have to define the heating of the surface-troposphere system, $\Delta F(t)$, due to increasing greenhouse gases that explained in the atmospheric model.

Atmospheric Model

The final quantity which remains to be specified is $\Delta F(t)$. Recall that this term represents infrared heating of the surface-troposphere system due to increasing levels of carbon dioxide, methane, nitrous oxide, chlorofluorocarbon.
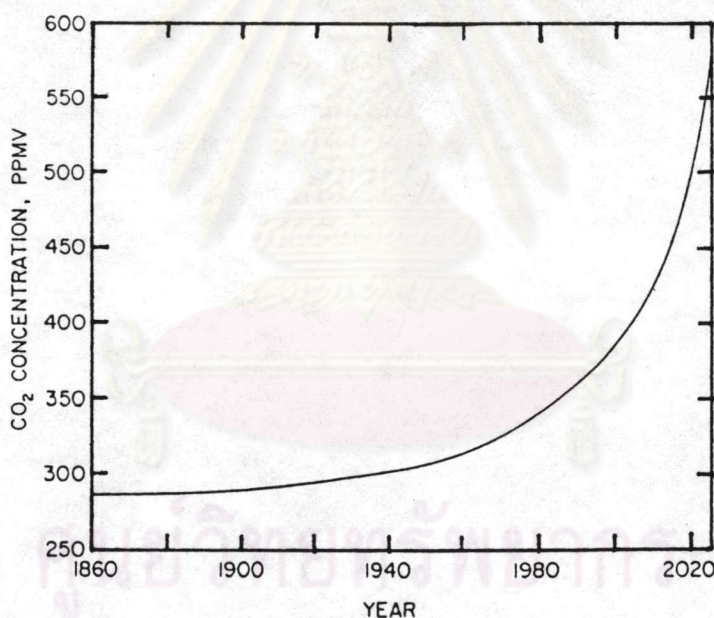

Figure 3.2   Atmospheric CO_2 Concentration from 1860 to 2025.

This first necessitates knowledge of the variation in atmospheric CO_2 with time, which has been taken from the carbon cycle models of *Machta and Telegadas* [1974] and *Keeling* [1976]. The time-dependence of CO_2 concentration,

which is a composite of these two models, is shown in Figure 3.2, with the preindustrial (1860) concentration of 287 ppmv then doubling to 574 ppmv by 2025.
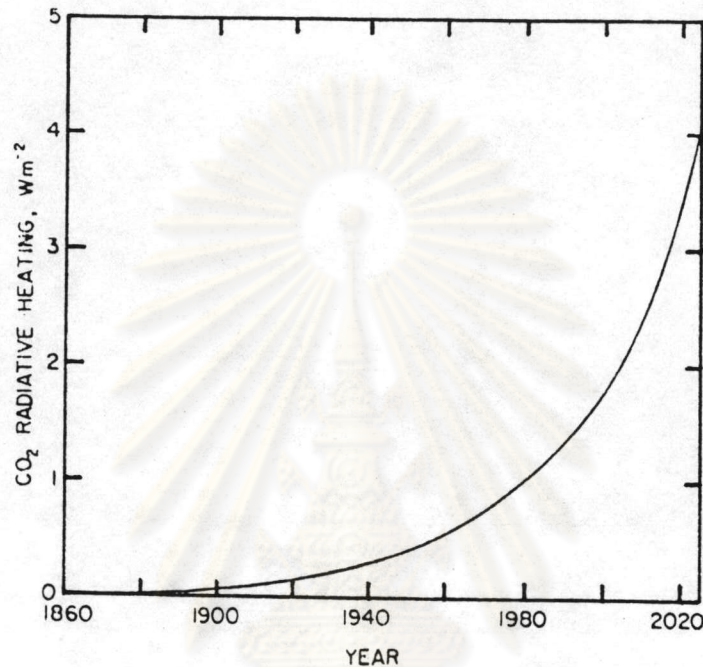

Figure 3.3   The CO_2 Radiative Heating Function from 1860 to 2025

Heat fluxes occurred from the carbon dioxide gases at various periods of time can be estimated using Equation (3-29). This correlation is the net amount of the solar energy that affects the Earth and the thermal energy radiating from the global surface of 2 levels; at the global surface and the upper level of the troposphere (tropopause). The difference of the net heat from these two levels is the amount of the heat accumulated in the lower part of the atmosphere. This estimating is considered in various scenarios (320, 426, 534, and 640 ppmv (*Ramanathan et al.*[1979])). Finally, by extrapolating and interpolating, we have

the relation between the heat fluxes and the concentrations of carbon dioxide. Then the relation with time is also acquired.

$$\Delta F(t) = 0.0277(e^{\frac{t}{\tau_c}} - 1) \qquad (3\text{-}29)$$

where $\tau_c = 33$ years

By comparing the amounts of other greenhouse gases with the CO_2 at any time and fixing the global warming potential index at the integrating time of 100, heat fluxes from greenhouse gases are determined. The example data are shown in Table 3.4.

Table 3.4   Global Warming Potential and Concentration of Greenhouse Gases

Greenhouse gas	Global warming potential	Concentration
CO_2	1	351 ppm
CH_4	21	1670 ppb
N_2O	290	305 ppb
CFC	5400	697 ppt

Then the amounts of heat fluxes due to the increasing of greenhouse gases are calculated by comparison of global warming potential of these gases with that of CO_2. The result are shown in Table 3.5.

Table 3.5   Heat Flux of the Surface-Troposphere System of Greenhouse Gases compared to CO_2

Greenhouse gas	Global warming potential	Concentration	$\Delta F(t)$
CO_2	1	351 ppm	$\Delta F(t)$
CH_4	21	1670 ppb	$0.0999\,\Delta F(t)$
N_2O	290	305 ppb	$0.2520\,\Delta F(t)$
CFC	5400	697 ppt	$0.0107\,\Delta F(t)$

Therefore, heat generated in the troposphere in the term of the increasing in four greenhouse gases can be expressed as:

$$\Delta F(t) = 0.0378(e^{\frac{t}{\tau_c}} - 1) \qquad (3\text{-}30)$$

Substitute heat flux accumulated in the troposphere as the above equation into Equation (3-26) as complete heat transfer case and Equation (3-28) as zero heat transfer case, we can estimate the sea-surface temperature interval change from the year of 1860's by analytical method.

Determining of $\Delta F(t)$ as a Function of the Amount of Greenhouse Gases

The influence of CO_2 or other greenhouse gases on the elevation of global temperature can be estimated by defining the amount of greenhouse gas and then calculating the heat flux occurred in the system. Concentration of CO_2 must be fixed first and heat fluxes generating from other greenhouse gases then can be found by comparing influence of global warming potential value and amount of these gases with that of CO_2.

The method for determining the heat flux in the system is corresponding to the method of time dependence. That is, after the value of CO_2 or other greenhouse gases had been fixed, the ratios of concentration between those greenhouse gases and that of CO_2 are then calculated. Then the influence of global warming potential values of those gases are compared with the calculated one.

By using the method mentioned above, heat flux due to the increasing of greenhouse gases can be estimated. Time in the heat flux equation can be approximated by expressing concentration of greenhouse gases as a function of time and then determined sea-surface temperature by using the same method.