CHAPTER 2

THEORIES AND LITERATURE REVIEW

2.1 Carbon cycle and Earth's climate

Earth's climate has changed during the past century and will continue to change significantly over the next few centuries. The Intergovernmental Panel on Climate Change Report (IPCC, 2007) predicts temperature rises of 1.8 – 4.0 °C in the future. The increasing rate of mean temperature in Thailand during 1965 - 2006 was about +0.81 °C (Limsakul and Goes, 2008). The increasing rate of means temperature in Thailand was higher than the mean temperature in the world about +0.016. °C decade^{-1} (Chidthaisong, A., 2009). Moreover, the predicted changes for the next 50 to 100 years are larger and faster than previously thought. Reilly *et al.* (1990) predicted and global mean temperature over the next 100 years and indicates that it will be at the high end of, or even exceed the IPCC 2001 predictions of +1.4 to 5.8 °C above the temperatures of the 1990 and +1 to 4.5 °C compared to the present for a doubling of atmospheric CO_2 (Kattenberg, *et al.*, 1996). Hence, global warming is an important environmental problem nowadays, because it involves remains of life on the earth.

Form evidence of science, global warming has resulted from the enhanced greenhouse effect in atmosphere, caused by the elevated concentrations of greenhouse gases such as mainly carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), sulfur hexafluoride (SF_6) and Chlorofluorocarbons (CFC_s). Among these greenhouse gases, CO_2 accounts for more than 60% of total global warming potential (IPCC, 2007). Currently the global average atmospheric CO_2 concentration has increased from about 285 ppmv (parts per million on a volume basis) in 1850 to 367 ppmv in 1999, and is increasing at a rate of 0.5% yr^{-1} (IPCC, 2007).

Rising atmospheric CO_2 and temperatures are probably altering ecosystem carbon cycling, and causing both positive and negative feedback to the climate. Impacts of climate change with CO_2 on the forest carbon cycle affects to photosynthesis and respiration. They are controlling the biosphere carbon cycle. Normally, responding of forest ecosystem with climate change has two contradictory scenarios. With the "Positive feedback" mechanism, more CO_2 is produced by increased decomposition, raising CO_2 atmospheric concentration and the global temperature further, which again increase decomposition rate and CO_2 release. In the "negative feedback" scenario, an elevated CO_2 level increases primary productivity and promotes production of more recalcitrant organic matter, leading to decreased mineralization rates. With increased production and slower decomposition, soil could serve as an important sink for carbon and counterbalance the effects of global warming.

The capacity of ecosystems to store C depends on net ecosystem production (NEP), which depends on the balance between net primary production (NPP) and heterotrophic respiration. Knowledge of the underlying mechanisms driving changes in NEP is essential to predict terrestrial C cycle responses to rising temperature and CO_2. Current understanding suggests that the primary direct ecosystem response to increased CO_2 concentration is an increase of NPP, which is potentially a negative feedback on atmospheric CO_2 concentration. However, soil respiration usually increases when ecosystems are exposed to elevated CO_2 (Luo and Zhou, 2006). King, *et al.* (2004) studied soil respiration rates at ambient and elevated CO_2 from the Duke Forest FACE (Free-Air CO_2 Enrichment) experiment at 15-year-old stand of loblolly pine in North Carolina during 1996 to 2001. They found that the soil respiration increases average 22% during 5 years. Similarly, grassland community in California is exposed to elevated CO_2 for three years, the flux of CO_2 from the soil surface increases from 323 to 440 $gCm^{-2}yr^{-1}$ (Luo, *et al.*, 1996). Soil respiration increases by 12 to 40.6% in a sweet gum forest in Tennessee and developing popular forests in Wisconsin and Tuscany, Italy, at elevated CO_2 in ambient CO_2 (King, *et al.*, 2004).

One of the explanations for increases in the soil respiration at elevated CO_2 was that such increases were mainly the result of changes in the substrate supply to the rhizosphere. Rising atmospheric CO_2 stimulates plant photosynthesis and growth. Recent reviews indicate that increases in CO_2 concentration by 200 to 350 ppm usually stimulate photosynthesis by 40 to 60% (Ceulemans and Mousseau, 1994, Medlyn, *et al.*, 1999), which are biomass growths between aboveground about 22.4% and belowground about 31.6%. Increased photosynthetic carbon fixation and plant biomass growth result in delivery of more carbon substrate to belowground at elevated CO_2 than at ambient CO_2. Increased carbon substrate stimulates root biomass, specific root respiration, root turnover rates, litter production, litter decomposition, root exudation, soil priming, and microbial activity (Luo and Zhou, 2006). Especially, elevated CO_2 usually stimulates litter production and has little effect on specific rates of litter production. Litter biomass increases by 20.6%, average over 14 paired observations (Ceulemans and Mousseau, 1994, Medlyn, *et* *al.*, 1999). In general, elevated CO_2 has no effect on the specific rate of leaf litter decomposition within plant species. However, increased litter productions affect the total amount of substrate available for heterotrophic respiration, thereby contributing to increased soil CO_2 efflux.

Farquhar, *et al.* (1980) used the Farquhar Model for predicting photosynthetic and responses of leaves to temperature. The influence of increasing temperature on photosynthesis depends on the absolute temperature (Slatyer, 1977), such that at suboptimal temperature as 3°C increase in temperature could lead to as much as 40% increase in photosynthetic rate. As consequence, most modeling studies assume that the increase in soil respiration per 10 °C rise in temperature (Q_{10}) is about 2.0 (Schimel, *et al.*, 1994; McGuire, *et al.*, 1995; Cox, *et al.*, 2000). In contrast, at high temperatures (greater than approximately 25-30 °C) an increase in temperature can cause a reduced rate of photosynthesis, because the photorespiratory rate has a higher temperature coefficient than photosynthesis (Jordan and ogren, 1984; Brooks and Farquhar, 1985).

This negative effect of temperature is reduced by an increase in CO_2 partial pressure, which enhances the rate of photosynthesis relative to that of photorespiration (Harley *et al.*, 1992, McMurtrie and Wang, 1993). In addition, Cox *et al.* (2000) studied the carbon feedback model in the terrestrial ecosystem by using the general circulation model (GCM). The model was defined increasing atmospheric CO_2 concentration by scenarios in the future, as well as vegetation. Carbon in South America begins to decline and land biosphere switches from being a weak sink from CO_2 to being a strong source. The results show that an increase in the concentration of atmospheric CO_2 alone tend to increase the rate of photosynthesis and thus terrestrial carbon storage. However, plant maintenance and soil respiration rate both increase with temperature. As a consequence, climate warming (the indirect effect of a CO_2 increase) tends to reduce terrestrial carbon storage, especially in the warmer regions where an increase in temperature is not beneficial for photosynthesis.

At low CO_2 concentrations the direct effect of CO_2 dominates, and both vegetation and soil carbon increase with atmospheric CO_2. But as CO_2 increases further, terrestrial carbon begins to decrease, because the direct effect of CO_2 on photosynthesis saturates but the specific soil respiration rate continues to increase with temperature. In addition, rising atmospheric CO_2 and temperatures are probably altering ecosystem carbon cycling, causing both positive and negative feedbacks to climate. Impacts of climate change with CO_2 on the forest carbon cycle affects to photosynthesis and respiration.

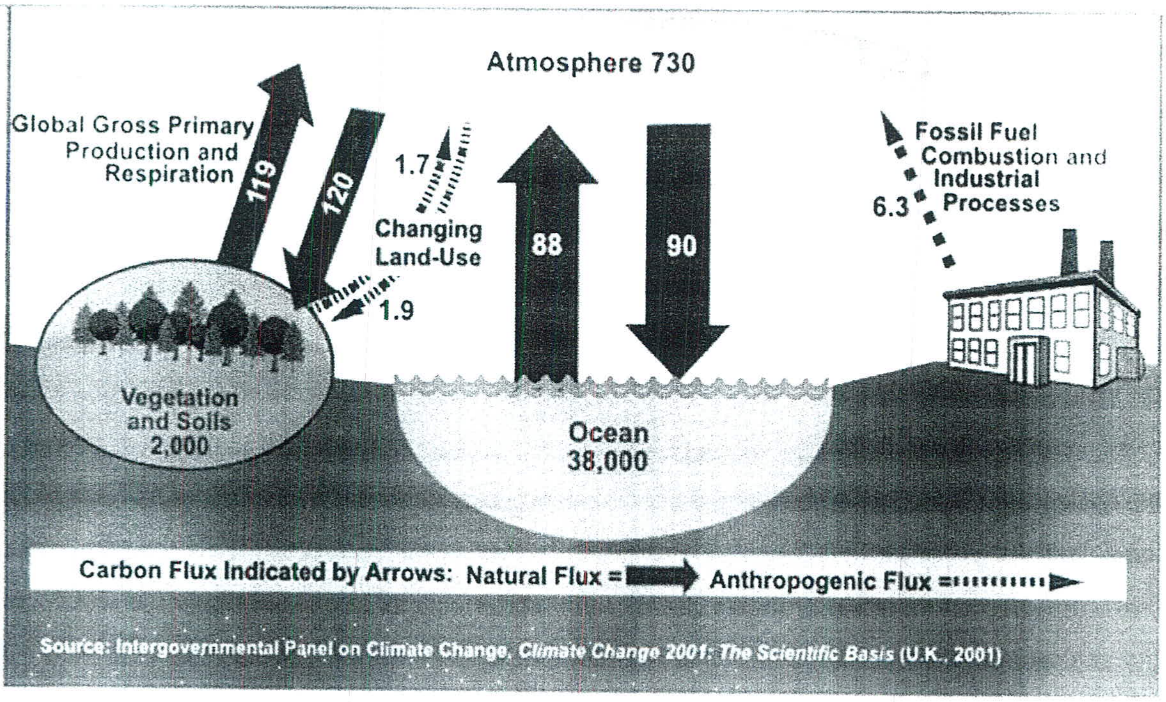
In sum, the carbon cycle is one of the key drivers of climate change with varying complexity in its responses and feedbacks. Global warming and climate change is expected to exert both positive and negative responses/feedbacks. Researches in the past decades have revealed various aspects of climate-carbon cycle interactions and there are still huge gaps in our knowledge of such interactions at different temporal and spatial scales.

2.2 Role of forest soil respiration on global carbon cycle

The global carbon cycle connects three major components of the global system, the atmosphere, the oceans, and the terrestrial ecosystem (Figure 2.1). Annually, photosynthesis of land plants takes up approximately 120 Pg (10^{15} g) C yr^{-1} from the atmosphere. About 119 PgC yr^{-1} of carbon is released back to the atmosphere through ecosystem respiration. Thus, the net flux on terrestrial ecosystem is a sink about 1.4 PgC yr^{-1} (Luo and Zhou, 2006). Oceans absorb nearly 92 PgC yr^{-1} from the atmosphere and release 90 PgC yr^{-1} back into the atmosphere through physiochemical exchanges of CO_2 at the air-sea surface, and through photosynthesis and respiration of marine organisms.

The burning of fossil fuels by humans presently adds about 6.3 PgC yr^{-1} to the atmosphere. Land clearing, deforestation, and fire release an additional 1.6 PgC yr^{-1} to the atmosphere. The amount of CO_2 added to the atmosphere by human activities may seem very small in comparison with the rates of flux through natural processes such as photosynthesis and respiration, but it takes only a small change to upset the balance of the global carbon cycle.

The main global carbon pools are soils, fossil fuels and oceans, which contain 3150, 6000, and 39000 PgC, respectively (Luo and Zhou, 2006). The soil carbon pool includes 450 PgC in wetlands, 400 PgC in permanently frozen soil, and 2300 PgC in other ecosystems (Sabine *et al.*, 2003). The latter 2300 PgC can be further divided into 1500 PgC in top-soils to the depth of 1 meter, and 800 PgC in the deeper soil layers to the depth of 3 meters (Jobbagy and Jackson, 2000). Plants contain about 650 PgC. The sum of soil and plant carbon content is 3800 PgC. This is five times the size of the atmospheric pool (750 PgC). Therefore, terrestrial ecosystems store large quantities of carbon and play an important role in controlling the concentration of CO_2 in the atmosphere.



Source: Intergovernmental Panel on Climate Change, *Climate Change 2001: The Scientific Basis* (U.K., 2001)

Figure 2.1 The global carbon cycle. Pools in Pg (= 10^{15}g) C and fluxes in PgC yr^{-1} are indicated by arrows (http://www.netl.doe.gov/)

Forests account for 52% of the world's annual carbon uptake by terrestrial vegetation (U.S. Environmental Protection Agency, 2000). The forest ecosystem is thus important with regards to carbon exchange between the earth and the atmosphere (Figure 2.2). The carbon cycle in the forest is initiated when carbon is fixed via photosynthesis. Some of the fixed organic carbon compounds are used to grow tissues. Some are broken down to supply the plants with energy. During this process, carbon is released to the atmosphere from ecosystem respiration, disturbance and herbivory. In forest ecosystems, CO_2 is produced from aboveground and belowground activity by respiration process. CO_2 from aboveground activities is produced from leaf, branch, and stem respiration but belowground, CO_2 occurs from respiration of roots, microbes, and soil animals. The largest component of the net flux of CO_2 is the efflux from the soil surface, i.e., soil respiration (40-80% of total respiration), followed by uptake for photosynthesis (Raich and Schlesinger, 1992; Goulden, *et al.*, 1996; Chambers, *et al.*, 2004; Ohkubo, *et al.*, 2007).

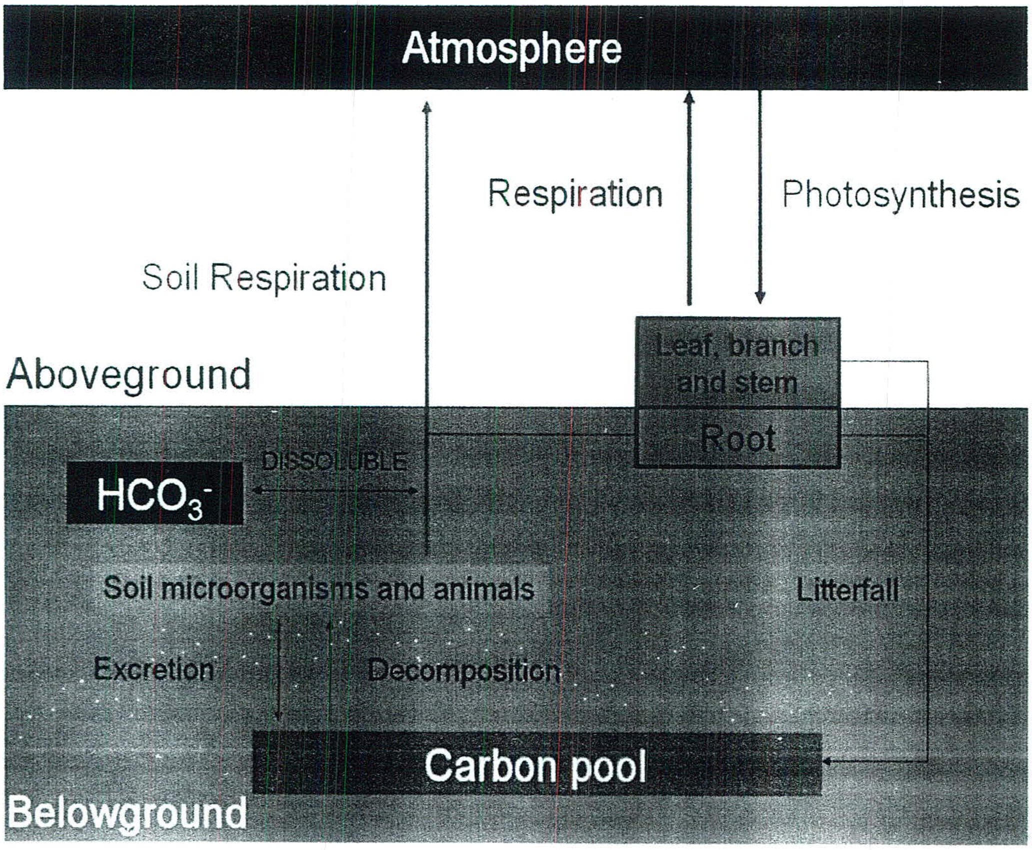


Figure 2.2 Schematic diagrams of ecosystem carbon processes in forests

2.3 Relationship between nutrients cycling and soil respiration in tropical ecosystem

The nutrients cycling processes, including photosynthesis and respiration, play an important role in the carbon cycle, especially in terrestrial ecosystems. The major pathways of nutrient cycling between soil and plants in ecosystem starts from nutrients, which are taken up from the soil by plants and are used for producing organic matter such as stems, branches, and leaves. After this living organic matter reaches a mature state, it will die and fall to forest floor in the form of litters. These litters are slowly decomposed by the activities of microorganisms and fungi in soil and release nutrients to soil for the utilization of plants growth. Dead plant material is then decomposed by microorganisms which provide energy for microbial biomass growth and other activities. Consequently, CO_2 is released back into the atmosphere through microbial respiration. These processes indicate that nutrients continue to flow in the ecosystem as long as the supply of substrates from litter fall continues. Generally, the natural nutrients cycling in forest ecosystems consist of three important processes (Brown, 1978):

1) Input process: the nutrients flow in to forest ecosystems by releasing nutrients from soil, rocks, and the constituent of nutrients with rainfall, litter decomposition, and fixed from the atmosphere.

2) Retention process: the process by which nutrients accumulate in living biomass located in soil and litter.

3) Loss process: the process by which nutrients are lost from the cycling system by soil erosion, and leaching, by man, animals, and other natural effects.

Three main processes stimulate the role of nutrient cycling in the ecosystem: primary productivity, decomposition, and soil respiration. In the forest ecosystem, each nutrient flows through a specific pathway. The differences of forest types and species are indicated by the approach of nutrient uptake, nutrient retention, and nutrient utilization for tree growth (Cole, 1986). Nutrient content in ecosystems are slowly accumulated by the input of nutrients from the atmosphere, weathering of minerals, release from litterfall, and fixing of some nutrients by microorganisms. Then, some nutrients are kept in aboveground and belowground biomass and soil. Sahunalu, *et al.*, (1984) studied in a dry dipterocarp forest, at Sakaerat Environment Research Station, Nakhon Ratchasima, and found that nutrients released from litter decomposition were composed of Ca, N, K, Mg, and P. The highest content was Ca and the next highest contents in sequence were N, K, Mg, and P. The amount of Ca, N, K, Mg, and P released into soil were 45.04, 42.92, 20.67, 9.79, and 2.61 kg ha^{-1} yr^{-1}, respectively.

Klinge and Rodrigues, (1968) studied the tropical and subtropical forests of Brazil and found that litter production in these forests were 7.3 kg ha^{-1} yr^{-1} and were composed of N, P, K, Ca, and Mg which released from litterfall 100-225, 2-11, 12-104, 18-206 and 12.53 kg ha^{-1} yr^{-1}, respectively. Brasell, *et al.*, (1980) studied a tropical rain forest in Australia and found that litter production in these forest ranged from 8.70 to 10.60 ton ha^{-1} yr^{-1}, they were composed of N, P, K, Ca, and Mg, and the amount released from the decomposition of litterfall to soil were 126-130, 10.80-11.20, 2.31-5.03, 56-63.70, 171-210, and 28.40-36 kg.ha^{-1}.yr^{-1}, respectively. These differing results indicate that the rate of nutrient release from litter decomposition in each forest are different, especially leaf litter.

Leaf litter plays an important role in releasing nutrient to soil, increase of soil fertility (Adulprasertsuk, *et al.*, 1996) and controlling soil CO_2 efflux rate also (Xu and Qi, 2001). For example, total N augments soil CO_2 efflux rate by providing a source of protein for microbial growth in mixed oak-conifer forest (Tewary, *et al.*, 1982) and P availability may limit microbial biomass in mineral soils at warm-temperate forest (Gallardo and Schlesinger,. 1994). In addition, Mg in the soil was highly correlated with soil CO_2 efflux, especially in the growing season ($R^2 = 0.58$, $N = 18$, p-value = 0.05) at a young pine have plantation in northern California (Xu and Qi, 2001). It was speculated that Mg may have affected soil microbial activity.

In the retention process, the nutrients in soil and air are taken up by plants and some microorganisms, and then transferred to stock the nutrient in aboveground and belowground biomass, and soil. The carbon stocks in different forests and locations vary significantly. For example, the carbon stocks in tropical zones are higher than in temperate zones, which may relate to higher variety of chemical, physical, and biological processes (Table 2.1).

Table 2.1 Carbon stocks in tropical and temperate zone.

Carbon stocks (tC ha^{-1})	**Tropical zone (Diloksumpun *et al.*, 2008)**		**Temperate zone (Duvigneaud and Denaeyer-De Smet, 1970)**
	Dry evergreen forest	**Mixed deciduous forest**	**Typical temperate forest**
Above-ground biomass	152.15	86.17	62.90
Below-ground biomass	71.51	40.50	17.50
Soil carbon	210.89	223.91	NA
Soil organic matter	NA	NA	125.00

In the loss process, the rate of nutrients released from the forest will determine the availability of nutrients, especially the nutrients that are obtained from decomposition and litterfall (Binkley, 1986). As has been stressed, the mainly CO_2 production from soil respiration in the tropical forest occurs from litter decomposition by microbial activity. This is the only process enabling massive recycling of chemical elements on the scale of ecosystems and the whole biosphere. The complete decomposition of organic matter means that the release of all energy fixed in organic compounds, which is tied to oxidation of carbon to carbon dioxide by microbial respiration.

Carbon dioxide is one of the main greenhouse gases in the atmosphere, responsible for maintaining the global temperature at a certain level. The public has become familiar with the danger of global warming due to the increase in CO_2 level in the atmosphere caused by massive fuel combustion and deforestation (Berg and Laskowski, 2006). However, the carbon emission from forest to atmosphere is composed of plant and microbe respiration. The magnitude of soil respiration is about 10 times that of fossil fuel burning and cement manufacturing combined (7.8 Pg C y^{-1} in 2005) (Forster, *et al.*, 2007). This suggests that ~ 10% of atmospheric CO_2 cycles through the soil annually (Reichstein and Beer, 2008). Thus, even small changes in soil respiration may greatly influence the CO_2 airborne fraction (Veenendaal, *et al.*, 2004; Kane, *et al.*, 2005).

Carbon dioxide efflux from soil respiration beneath a forest canopy to the atmosphere is a primary pathway by which carbon dioxide is exchanged between the ecosystem and the atmosphere; soil beneath a forest canopy can represent as much as 40-90% of the forest ecosystem respiration (Schlesinger and Andrews, 2000). In addition, Saigusa, *et al.* (2008) studied the CO_2 flux in various forest types (Figure 2.3). These indicate that ecosystem respiration (R_e) in tropical zones were higher than temperate and sub-arctic zones (Figure 2.4) because of the difference in the length of summer. This observation is likely due to the fact that the most important controlling factor of R_e in sub-arctic and temperate forests is the temperature, and other factors, such as drought stress, are not dominant. The maximum monthly value of R_e was highest in the temperate larch forest in mid-summer (Figure 2.4 d). This high R_e might be related to the high photosynthetic activity of needles during the growing period of the temperate larch forest. The R_e of tropical seasonal forests (Figure 2.4 i and j) decreased from December to April, due to the dry season. In the tropical rain forest (Figure 2.4 k), the seasonality of R_e was less clear.

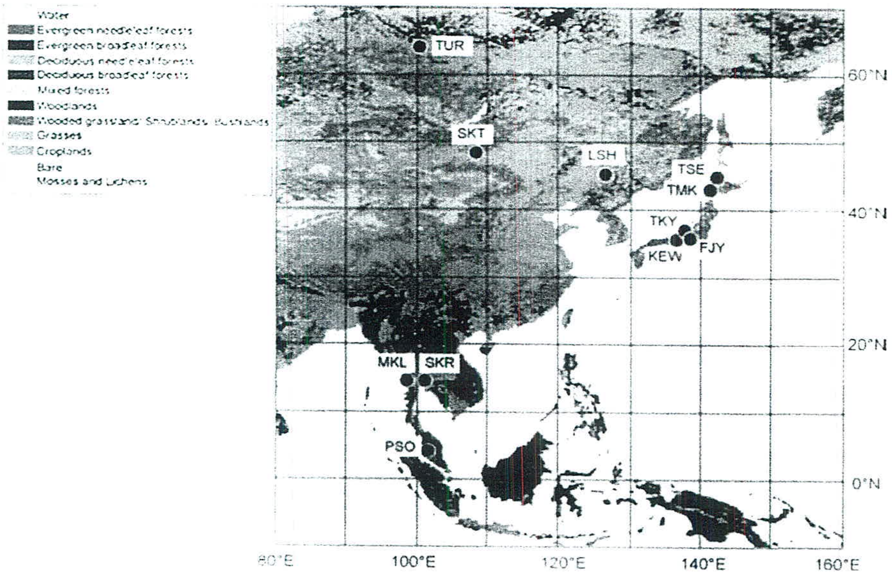


Figure 2.3 Location of study sites in East Asia plotted on a land classification map (De Fries, *et al.*, 1998).

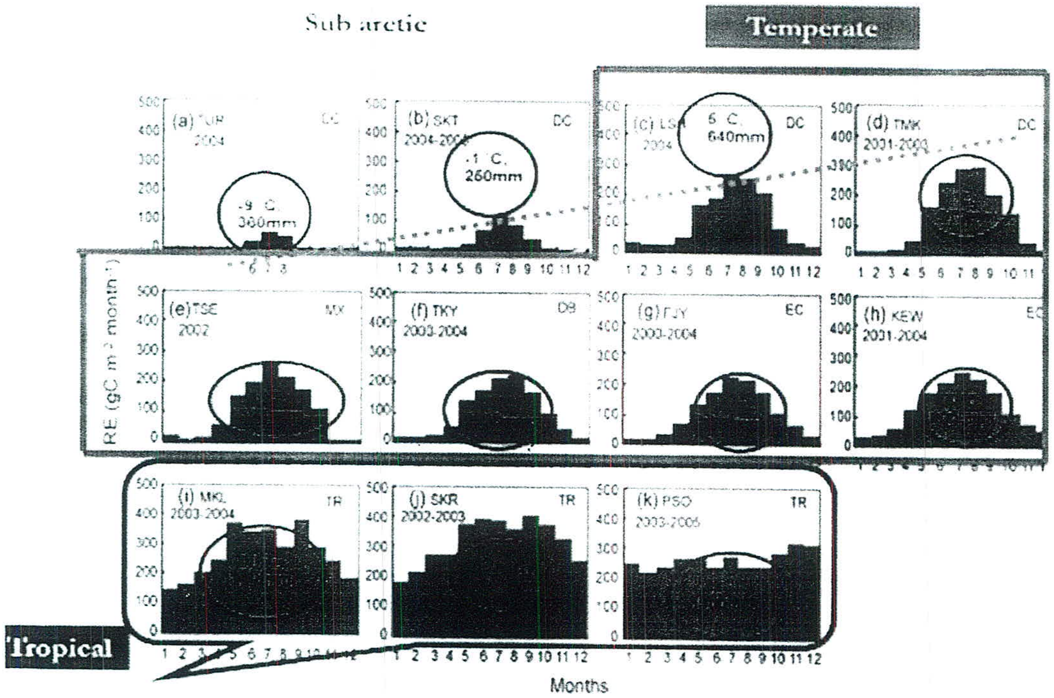


Figure 2.4 Monthly Re observed at the 11 sites listed in Table 2.2. The site code, year(s) of ecosystem respiration (Re) evaluation, and ecosystem type are indicated in each figure (Saigusa, *et al.*, 2008).

2.4 Components of ecosystem respiration and their relationship with environmental factors

Carbon dioxide in forests is produced by both plant respiration (R_p) and microbial respiration (R_M, Figure 2.2). The plant respiration (R_p) is often called autotrophic respiration and can be separated into aboveground plant respiration (R_a) and belowground plant respiration (R_b). The belowground plant respiration is mainly root respiration. The microbial respiration occurs during the decomposition of litter and soil organic matter (SOM) and is called heterotrophic respiration. The CO_2 efflux rate measured at the soil surface (R_s) is the sum of root respiration and microbial respiration (Eq. 2.1).

$$R_s = R_b + R_M \qquad \text{(Eq. 2.1)}$$

The ecosystem respiration (R_e) is the sum of aboveground respiration (R_a) and soil respiration (R_s). Therefore, the total CO_2 emission from an ecosystem can be estimated by Eq. 2.2.

$$R_e = R_a + R_s \qquad \text{(Eq. 2.2)}$$

The total ecosystem respiration in each forest plays an important role in the global carbon cycle. It is regarded as the second largest CO_2 emission pathway to the atmosphere (see Figure 2.1). The ecosystem respiration rate in different vegetations and locations of several forests varies due to differences in various environmental factors at each site such as temperature and moisture of air and soil, soil C and N contents, biomass of fine roots and microbes, climate zone, etc.

Curtis, *et al.* (2005) studied the relationship between R_s (soil respiration) and R_e (ecosystem respiration) in an aspen-dominated mixed hardwood forest in Michigan from 1999 to 2003. They found that the average contribution of R_s is 71% of R_e. However, the relative contribution of R_s to R_e varies considerably in a year. R_s contributes nearly 100% of R_e for most of the winter; the contribution drops to about 60% during the period of fast last leaf expansion and then gradually increases during the growing season as soil warms, reaching about 75% at the time of leaf abscission in the autumn (Curtis, *et al.*, 2005). Typically, R_s contributes 30-80% of R_e in forests. The wide range of R_s/R_e ratios of forests available in the literatures suggests differences in controlling factors and responses for both R_s and R_e. Typically, the annual R_s/R_e ratio was 0.68 in temperate hardwood forest in

Massachusetts, USA (Goulden, *et al.*, 1996); 0.76 in a mixed-age ponderosa pine forest in Oregon, USA (Law, *et al.*, 1999); 0.48-0.71 among coniferous boreal forests of central Canada (Lavinge, *et al.*, 1997); 0.73 in a boreal aspen forest in Saskatchewan, Canada (Griffis, *et al.*, 2004); 0.38-0.99 among Euroflux forested study sites (Janssens, *et al.*, 2001); 0.38 in a mature evergreen forest in the central Amazon Basin (Chambers, *et al.*, 2004); and 0.31 in a mature evergreen forest of the eastern Amazon Basin (Saleska, *et al.*, 2003).

Moreover, Davidson *et al.* (2006) studied the seasonal pattern of the ratio of soil respiration in a spruce-dominated forest, in the USA. They found that R_s/R_e ratio reached a minimum of about 0.45 in the early spring, gradually increased through the late spring and early summer, leveled off at about 0.65 for the summer, and then increased again to about 0.8 in the autumn. For R_s/R_e in tropical forest, Hanpattanakit, *et al.* (2012) studied this ratio in a dry dipterocarp forest in Western Thailand. They found that the annual R_s/R_e ratio ranges from 0.24-1.20, with a mean of 0.57. R_s/R_e ratio was relatively high in the dry season [0.91, during November to April] when compared with the wet season [0.55, during May to October]. The seasonal change of R_s/R_e was controlled by soil environmental factors (soil moisture and temperature) and plant phenology (aboveground and belowground plan tissues).

The important environmental factors that control the ecosystem respiration and soil respiration are usually temperature and moisture (Figure 2.4). Figure 2.4 indicates that in tropical forests the ecosystem respiration is controlled by temperature and precipitation. Table 2.2 compares the annual amount of soil respiration in different locations. It is clear that the soil respiration rates were different because of the temporal variability. Thus, in a given forest ecosystem, it is essential to evaluate the relationship between soil respiration and environmental variables. In many cases, differences in soil temperature and water content may explain most of the temporal variation in soil respiration (Davidson, *et al.*, 1998; Qi and Xu, 2001; Reichstein, *et al.*, 2002; Rey, *et al.*, 2002). In tropical Amazonian forests, for example, Davidson, *et al.* (2000) reported that the soil respiration rate was high during the rainy period and low during the dry period.

Hashimoto, *et al.* (2004) and Hanpattanakit, (2008) reported that soil respiration rates increased following the increase of soil moisture in a tropical monsoon forest in Thailand. Tang, *et al.* (2006) studied effects of soil temperature and moisture on soil respiration in a broadleaf forest, a mixed forest, and a pine forest in Southern China. The results shown that soil respiration were positively correlated with both soil temperature and soil moisture (Figure 2.5). When soil temperature increases, root development is faster and microbial activity is elevated, resulting from the temperature sensitivity of enzymatically catalyzed reactions involved in respiration and the sensitivity of the increased ATP requirements as metabolic rates increase.

The temperature stimulation of respiration also reflects the increased demand for energy necessary to support the increased rates of biosynthesis, transport, and protein turnover that occur at higher temperature (Luo and Zhou, 2006). At higher temperatures, the availability of nutrients is greater because of elevated biotransformation rates of various compounds. In addition, the rate of water movement out of the soil usually becomes higher. The relationship between respiration and temperature is well-known, and understanding how temperature affects soil respiration is important to predict the soil response to changes in the climate. In addition, soil moisture influences soil respiration via physiological processes of roots and microorganisms. At high soil moisture, respiration is regulated primarily by oxygen concentration. Water in soil pores at high water content slows exchanges of gaseous O_2 and CO_2 at sites of microbiological and root activity. The effect of moisture on microbe growth varies with rates of biosynthesis, energy generation, and substrate uptake, as well as the nature and mode of water perturbation (Luo and Zhou, 2006).

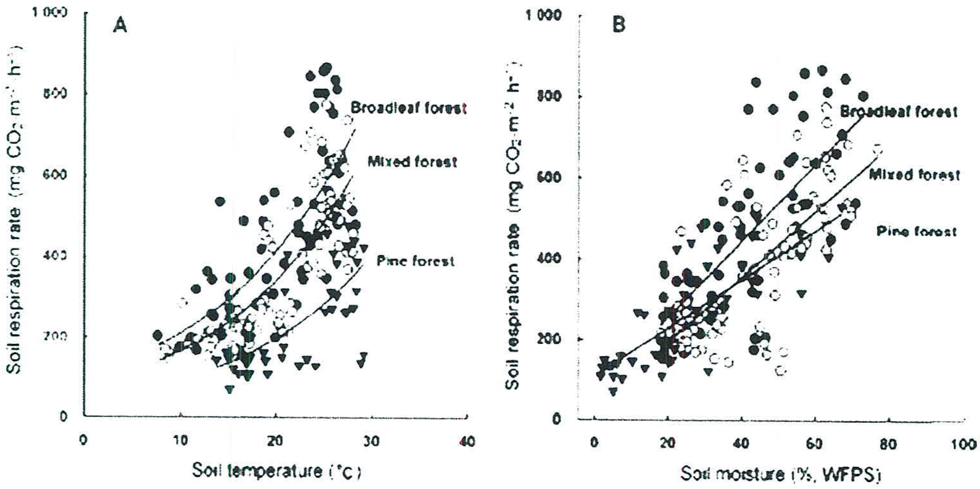


Figure 2.5 Relationships between soil respiration rate and soil temperature and moisture in Dinghushan Nature Reserve forests, Southern China (Tang, *et al., 2006*).

Occasionally, the relationship between soil respiration and soil moisture was found to be negative (Figure 2.6). This was because of the effects of daily changes in precipitation: precipitation after a dry period resulted in a large increase in ecosystem respiration, whereas additional precipitation after a rainy period had little effect. This response was similar to that of surface litter (coarse and fine), where respiration is greatly reduced when moisture is limited, but increases markedly and quickly saturates with an increase in moisture.

The correlation between soil respiration and soil temperature of a tropical forest in China were also negative (Figure 2.6). The reason for such a negative relationship is not known yet. However, this could be due to the possibility that the temperature at the site may be too high, especially in the afternoon, so that soil respiration was partially suppressed. Flanagan and Veum, (1974) found the optimal rate of soil microbial respiration at a temperature of 23 °C in the tundra ecosystem in Sweden, but the average soil temperature at the site was more than 25 °C (range 25 – 35 °C). The soil temperature and moisture values in figure 2.5 and 2.6 were not quit different but the soil respiration rate of both experiments differs greatly. Therefore, there are something that we could not explain when analyzing the relationships between respiration and temperature. Other physical, chemical and biological factors in each forest may play important roles and should be observed to improve our understanding on the mechanism of soil respiration in response to the environmental factors and changes.

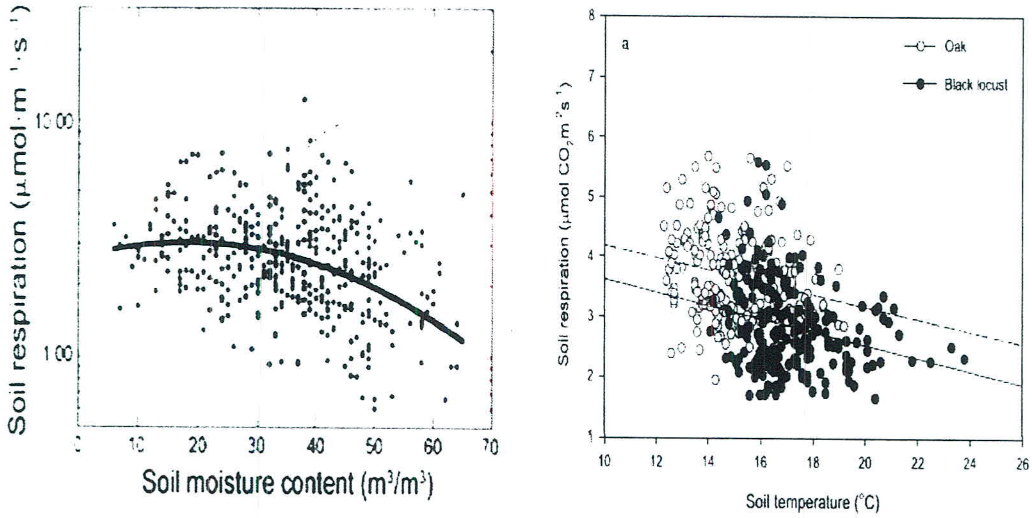


Figure 2.6 Relationships between soil respiration rate with soil moisture in Brazil (Chamber, *et al.*, 2004), and soil temperature in China (Shi, *et al.*, 2011).

In boreal regions, seasonal variations in soil respiration rates were determined to a large degree by soil temperature. Differences in soil water content had almost no effect on variation in soil respiration rate (Goulden, *et al.*, 1998; Moren, *et al.*, 2000; Rayment and Jarvis, 2000). In temperate regions, the soil respiration rate exponentially increased with soil temperature, and soil water content suppressed soil respiration rates under dry conditions (Dong, *et al.*, 1998; Londo, *et al.*, 1999; Kosugi, *et al.*, 2005; Mitani, *et al.*, 2007), or under both dry and wet conditions (Martin and Bolstad, 2005) or did not have a significant effect on soil respiration (Fang, *et al.*, 1998; Ohashi, *et al.*, 1999). Therefore, soil temperature plays a greater role than water content in governing soil respiration rates in temperate regions where the range of temperatures are wider. However, in tropical regions where seasonal variation in soil temperatures are small, soil water content should be tested as the most effective index to estimate the seasonal variation of soil respiration rate.

Table 2.2 Soil respiration rates in different vegetation types in the tropical regions

Soil respiration rate (mg CO_2 m^{-2} h^{-1})	Vegetation/location	References
948	Tropical forest (primary forest)/Malaysia	Minako, *et al.* (2006)
707	Tropical forest (secondary forest)/Malaysia	Minako, *et al.* (2006)
966	Oil palm plantation/Malaysia	Minako, *et al.* (2006)
625	Tropical semi deciduous forest/Thailand	Tulaphitak, *et al.*(1983)
338-503	Three types of tropical forest/Australia	Kiese, *et al.* (2002)
469-914	Tropical forest/Panama	Kursar, (1989)
231-444	Tropical bare soil/Brazil	La Scala, *et al.* (2000)
216-510	Tropical forest/Brazil	Fernandes, *et al.* (2002)
183-1162	Pasture/Brazil	Fernandes, *et al.* (2002)
430-675	Tropical forest/Costa Rica	Schwendenmann, *et al.* (2003)
435.4	Pine and broad-leaf forest/China	Yi, *et al.* (2007)
429.5	Pine forest/China	Yi, *et al.* (2007)
477.9	Monsoon evergreen broad-leaf forest/China	Yi, *et al.* (2007)
73.67	Subtropical mountain cloud forest/ Taiwan	Chang, *et al.* (2008)
349.32	Dry Dipterocarp forest/ Thailand	Hanpattanakit *(2008)*

Over the last decade, research has focused on the measurement of fluxes at the soil surface using a variety of chamber and micrometeorological methods. However, there is considerably less information available on CO_2 dynamics below the soil surface, apparently due to the difficulty of sampling and measuring soil CO_2 concentrations. Though process-based models (Fang and Moncrieff, 1999; Jassal, *et al.*, 2004; Simunek and Saurez, 1993) are valuable tools in increasing our understanding of various processes governing the CO_2 exchange within the soil, they need to be validated using measurements.

In addition, the litter production and decomposition in a forest are important for controlling the carbon cycle in a terrestrial ecosystem because the chemical compositions in litter are converted from fresh litter to humus by decomposition processes. This process produces CO_2 emission to the atmosphere by microbial respiration. Microorganisms start degrading plant litter as soon as it has fallen to the ground and been invaded by decomposers, fungal mycelium and bacteria. In tropical forest ecosystems, decomposition processes of litter are very rapid (Sukwong, 1982).

Glumphabutr, (2004) studied the annual CO_2 release of litterfall in a Moist Evergreen Forest (MEF), Dry Evergreen Forest (DEF), and Hill Evergreen Forest (HEF), in Chanthaburi Province, Thailand. Their results showed that, the decomposition rate of DEF, MEF, and HEF were 7.44, 5.86, and 2.80 ton ha^{-1} yr^{-1}, respectively, which are different in each forest type depending on species of plant and the climatic condition. The difference in decomposition of litter as a result of interaction among controlling factors is usually reflected in the difference in the average value of k (normalized decomposition coefficient). This has been estimated in each region of the world such as; in boreal region is in the rage of 0.223-0.446, cold temperate region in the range of 0.140-0.693, warm temperate region in the range of 0.162-0.751 and tropical region in the range of 0.162-2.813 (Landberg and Gower, 1997). The k value in tropical regions was higher than temperate and boreal forest ecosystems indicate the more rapid decomposition rates.

Changes in climatic conditions along a latitudinal gradient influence decomposition processes through temperature and water conditions. In temperate forest ecosystems, decomposition processes have been well studied and the variability of decomposition rates has been related to litter qualities and site conditions (Aber, *et al.*, 1982, McClaugherty, *et al.*, 1985, Takeda, *et al.*, 1987). Important issues relating microorganism activity to biomass decomposition are physical and chemical properties of biomass such as the value of C/N ratio. Furthermore, climatic conditions affect to the amount of microorganisms. The important factors of climatic conditions are temperature and moisture (Vitousek and Sanford, 1988). These factors affect the development of vegetation and microbial activity which decompose organic matter on surface soil (Singh, 1969). However, weight loss accorded a linear decline and was better correlated with soil moisture than temperature (Isaac and Nair, 2005).

In the future, the global precipitation and temperature will be changed from the climate change effects, which will affect the litter decomposition rate and ecosystem respiration. Therefore, our understanding on biophysical parameters effects on litter decomposition rate such as temperature, moisture, soil texture, soil microbial biomass, root biomass, and growth rate would help us in improving the understanding of the impacts of future climate changes on tropical and global carbon cycle.

2.5 Dipterocarp forest; importance, distribution in the world and Thailand

Tropical rainforests cover approximately 60 per cent of the region's total forest area with tropical moist deciduous forests and tropical dry forests each accounting for around 15 per cent and mountain forests another 10 per cent (FAO 2005). Dry dipterocarp forest (DDF) is an important deciduous forest type in the world. Appanah and Turnbull (1998) studied plant distribution in DDF from life and alife (fossil) of dipterocarpaceae species. They found that the distribution patterns of dipterocarps were thought to reflect routes of colonization and past climatic conditions (Figure 2.7). Living dipterocarps *sensu lato* are spread over the tropical belt of three continents of Asia, Africa and South America. They occupy several photogeographical zones that mainly conform to climatic and ecological factors.

Southeast Asia, contains the world's third largest tropical forests, and is experiencing deforestation rates higher than almost anywhere else on Earth. The region's forests are endangered by conversion to agriculture or other land uses, such as oil palm plantations, logging (both legal and illegal) and climate change. In addition, the high value in DDF products such as camphor, resins, and timber stimulated deforestation from community people and entrepreneur. Royal Forest Department, RFD (2010) has mentioned status of DDF in Thailand that is in dangerous stage due to tree species are hard wood and used for construction. Additionally, many tree species may be appropriately used for fuel. Consequently, people want to use hardly wood from this forest type (Kutintara, 1999).

Forests in Thailand are divided into two main types: evergreen forests and deciduous forests (Santisuk, 2006). The evergreen forest consists of 14 types: tropical evergreen rain forest or tropical rain forest, seasonal rain forest or semi-evergreen forest or dry evergreen forest, lower montane rain forest, lower montane oak forest, lower montane pine-oak forest, lower montane coniferous forest, lower montane scrub, upper montane rain forest or cloud forest, upper montane scrub, montane peat bog or sphagnum bog, mangrove forest, peat swamp forest or coastal peat swamp forest, freshwater swamp forest, and strand vegetation. On the other hand, deciduous forests consist of 3 types: mixed deciduous forest, deciduous dipterocarp forest or dry dipterocarp forest, and pine-deciduous dipterocarp forest.

Tropical seasonal forests in Thailand are divided into three types: (1) dry dipterocarp forest (DDF), (2) mixed deciduous forest (MDF), and (3) dry evergreen forest (DEF). The ratios of each forest to total forest area are 21% for DDF, and 34% for MDF. Thus, 55% of forests are deciduous. On the other, tropical evergreen forests occupy 40% of forests in Thailand. The total forest in Thailand in 2005 was about 14.5 million hectares (FAO, 2005). Aboveground biomass of dry dipterocarp forest is about 126 ton ha^{-1}, relatively low compared to other types of forest (Table 2.3).

Table 2.3 Values of aboveground and stem wood biomass and carbon content of dry biomass by various forest types (Puangchit, 2000)

	EGF	**MDF**	**DDF**	**PE**	**MGF**
Carbon content of dominant species (%)	54	52	49	48	55
Aboveground biomass (ton ha^{-1})	337	266	126	160	200
Stem wood biomass (ton ha^{-1})	229	149	88	102	162

Note:
EGF = tropical evergreen forest
MDF = mixed deciduous forest
DDF = dry dipterocarp forest
PE = pine forest
MGF = mangrove forest

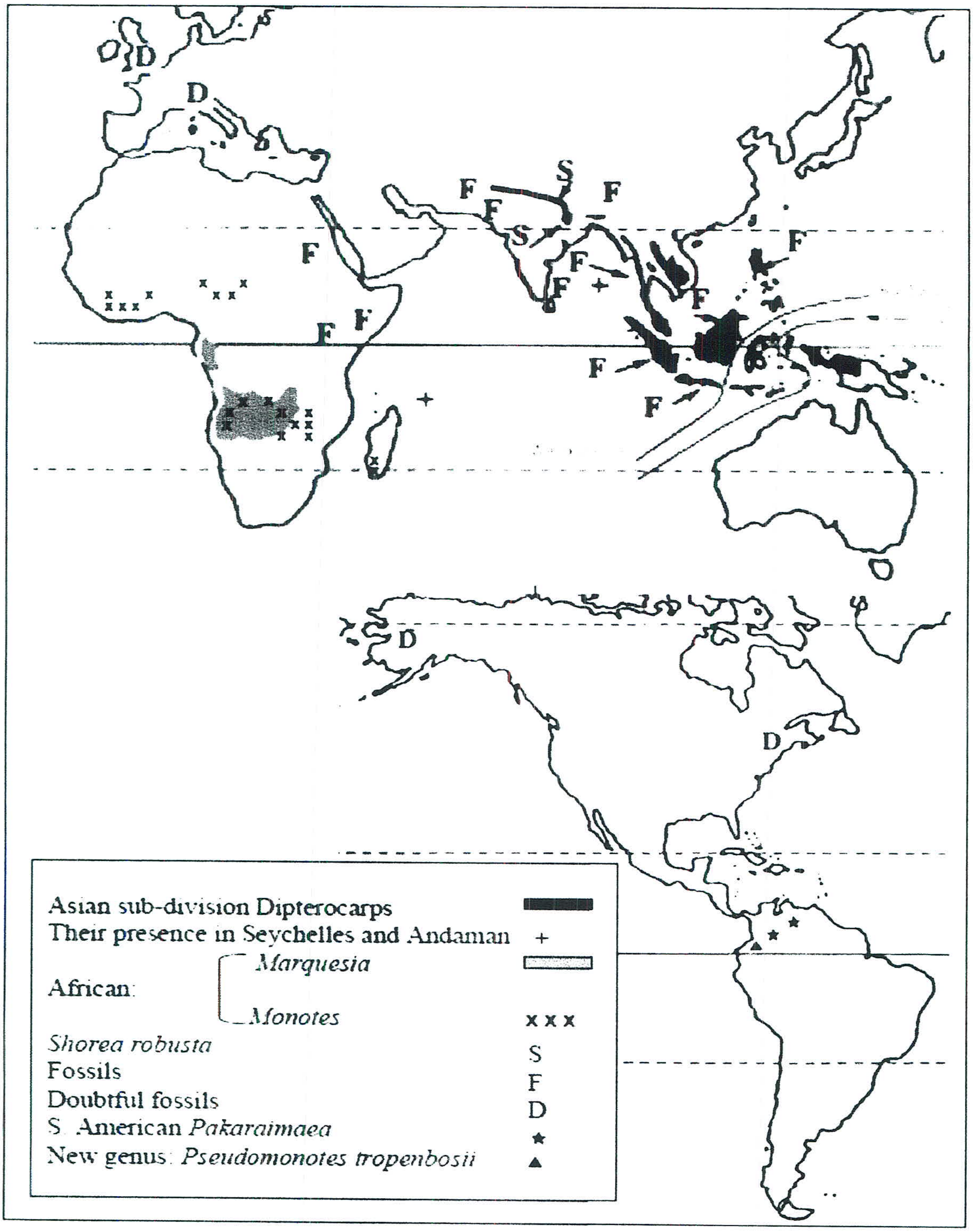


Figure 2.7 Distribution of DDF plant species in the world (Appanah and Turnbull, 1998)

2.6 Studies of carbon cycle in dipterocarp forest

The Tropical forests have higher carbon storage than other regions. Approximately 60 percent of total carbon in biomass stock on the Earth (Dixon, *et al.,* 1994). Its annual net primary production is about 32% of global terrestrial photosynthesis (Field, *et al., 1998*). Dlioksumpun, *et al.* (2008) studied the net primary production (NPP) in the Sakaeral Dry Evergreen (DEF) and the Maeklong Mixed Deciduous Forests (MDF) in Thailand. They found that NPP in DEF (15.3 tC ha^{-1} yr^{-1}) was higher than MDF (6.2 tC ha^{-1} yr^{-1}). However, the litter biomass product in MDF was higher than DEF about 7% because the plant in MDF is to shed leaves. Many investigation results suggest that tropical forests act as both the carbon sources and sinks, depending on the balance between photosynthesis and respiration and their temporal responses to environmental variables.

Research and knowledge of CO_2 flux measured in tropical forests in Thailand is sparse. For example, Gamo and Panuthai, (2005) studied CO_2 flux in the tropical seasonal forests at Sakaerat and Maeklong sites, which represent dry evergreen and mixed deciduous forests, respectively, during 2001-2004. Annual ecosystem exchange (NEE) indicated that the Sakaerat evergreen forest was a carbon sink with a magnitude of 1.5 and 1.3 tC ha^{-1} yr^{-1} in 2001 and 2002 respectively. In 2003 and 2004, the magnitude of the sink increased to 4.6 and 4.7 tC ha^{-1} yr^{-1} respectively. In the Maeklong deciduous forest, annual NEE was 5.6, 3.7 and 4.5 tC ha^{-1} yr^{-1} in 2001, 2002 and 2003, respectively. Soil respiration in DEP, MDF and DDF emissions was 13.37, 12.14 and 10.67 tC ha^{-1} yr^{-1}, respectively (Panuthai, *et al.*, 2008).

Tropical forests have a large potential to sequester carbon primarily through reforestation, agroforestry and conservation of existing forests (Brown, 1996) and the high productivity of tropical forests may make them particularly responsive to growth enhancement from increasing atmospheric CO_2 concentrations (Prentice and Lloyd, 1998). Furthermore, old-growth tropical forests can continue to serve as carbon sinks rather than reaching a steady state C equilibrium (Luyssaert, *et al.*, 2008). Thus, accurately characterizing the stocks and fluxes of carbon in tropical forests will have great implications for understanding the global carbon cycle.