LITERATURE REVIEW

Soil CO_2 efflux

Soil CO_2 efflux or soil respiration is defined as the CO_2 efflux from the soil surface or the outcome of the production of CO_2 by both plant root (autotrophic respiration) and micro-organism (heterotrophic respiration) in soil surface. Generally, sources of CO_2 in the soil which derived from (1) growth and maintenance respiration by roots (true root respiration) (2) rhizomicrobial respiration (i.e. heterotrophic decomposition of carbohydrates derived from live roots), (3) decomposition of fresh organic matter (surface and root litter), (4) decomposition of old soil organic matter, (5) priming of soil organic matter decomposition by substrate input from live roots or plant litter, and (6) weathering of soil carbonates. In recent yeas, many studies have focused on partitioning to total soil respiration in to its components and on understanding the relationship between specific sources of CO_2 and the environmental factors controlling them. Furthermore, soil respiration provided a useful parameter of metabolic activity of heterotrophic microbes and plant roots and nutrient mineralization in the soil. For example, increase in root respiration could be an indicator of increased photosynthate translocation to the soil. Thus, measurement of soil respiration has been carried out in various ecosystems under a range of environmental conditions.

Several researchers have reported contribution of autotrophic and heterotrophic respiration to total soil respiration in various ecosystems. Since, Hanson *et al.,* (2000) indicated that root/rhizosphere respiration could account for as little as

10% to great than 90% of total *in situ* soil respiration depending on vegetation type and season of the year. The contribution of root respiration to total soil respiration also varies with 39% during the wet season and 41% during the dry season (Tang and Baldocchi, 2005). Jiang *et al.* (2005) found that rhizosphere respiration accounted for 25% of total soil respiration in old forest and 65% in the young forest. Rodeghiro and Cescatti (2006) also found annual autotrophic respiration accounts from 16-56% of total soil respiration in the seven different evergreen ecosystems and data observation shows a decrease of annual autotrophic respiration at increasing availability of soil nitrogen.

However, the relative contribution of root/rhizophere and microbial to total soil respiration is difficult to determine, as report by a wide range of estimating for soil. The factors controlling autotrophic and heterotrophic respiration are influenced by the complex interaction of environmental and biotic factors. Autotrophic respiration is influenced by the amount and activity of plant and reflects changes in environmental condition that control plant growth and development, photosynthesis and carbon allocation patterns (Shi *et al.,* 2006; Han *et al.,* 2007). While, heterotrophic respiration is dependent on the supply of respiratory substrates (primarily from plant litter, plant root exudates, plant root) as well as environmental conditions that control microbial growth and development, and supply and quality of respiratory substrate provided by plant, particularly plant root (Raich and Schlesinger, 1992; Nago *et al.,* 2007). Thus, autotrophic and heterotrophic respiration will respond differently to change in environmental conditions, it is crucial to get insight into both components of soil respiration.

Separation of soil CO_2 efflux

The separation of root respiration and microbial respiration under field conditions is very difficult but important because of root respiration is considered a primary contributor to the soil CO_2 pool and would improve understanding of plant's response to environmental change independently from that of soil conditions. In previous studies, several methods have been used to separate root and microbial respiration from total respiration including the excised-root method (Burton *et al.,* 1998), subtraction method (Gansert, 1994), the root cuvette method (Bouma and Bryla, 2000). Hanson *et al.* (2000) have reviewed the method for separating soil respiration that can be divided in to three broad categories: component integration, root exclusion and isotopic techniques. Root exclusion is used to estimate root respiration indirectly by comparing measured CO_2 efflux rates at soil surface with or with out living root. Existing root exclusion techniques may be categorized in to three broadly defined area (1) root removal (2) gap analysis and (3) trenching.

Trenching method of root exclusion has been widely used for separating total soil respiration into autotrophic and heterotrophic respiration. This method can be implemented by inserting root barriers into soil to cut off root growth and carbon supply without digging soil to estimate relative contributions of autotrophic and heterotrophic respiration to the total soil respiration. Measurement of CO_2 efflux in the trenched plots without the presence of live roots is the heterotrophic respiration from microbial decomposition of litter and soil organic matter. On the other hand, CO_2 effluxes at the soil surface in the untrenched plot where roots can normally grow are taken to quantify total soil respiration. The difference in CO_2 efflux between the trenched and untrenched plot is an estimate of autotrophic respiration. This method

can avoid the contribution of dead roots to CO_2 production due to trenching severs root, dead root usually decompose faster than soil organic matter. Root death occurs rapidly after severing and decomposition begins within the first month (Kelting *et al.,* 1998). This decomposing root-derived CO_2 efflux leads to increase in soil respiration rate. Several reviews have pointed out that plot trenching modifies biophysical condition and substrate supply for microbial respiration and may change soil temperature and soil water content (Jassal and Black 2006; Wang *et al.,* 2006). However, evaluations between soil respiration and heterotrophic respiration in a continuous are rare but are useful for improving understanding of different behavior between root respiration and heterotrophic respiration.

Factors influencing soil CO_2 efflux

Soil temperature is the most important factor in regulating soil respiration. Several researchers have proposed models or equation to predict soil respiration from more readily available biotic and abiotic measurement. Temperature is usually taken as important factors controlling soil respiration by difference relationships, including linear, quadratic, power, exponential and Arrhenius models. Soil respiration increases exponentially with increasing temperature, and this relationship is usually described with exponential and Arrhenius equations (Lloyd and Taylor, 1994).

Exponential Equation

$$R = ae^{bT} \quad (1)$$

Arrhenius Equation

$$R = R_{10} e^{\left[E_a \left(\frac{T - 283.15}{283.15 x T x R_g} \right) \right]} \quad (2)$$

$$E_a = a \left(\frac{T}{T - 227.13} \right) \quad (3)$$

where R is soil respiration, R_{10} is soil CO_2 efflux at 10 °C, T is the absolute soil temperature(K), R_g is gas constant (8.314 J $mol^{-1}K^{-1}$), E_a is Activation energy ($Jmol^{-1}$), *a*, b are constants

However, exponential relationships, especially Q_{10} relationships, are more frequently used to describe respiration rates from temperature. The increase in reaction rate per 10 ºC increases in temperature is known as the Q_{10}. The exponential function Q_{10} is commonly used to express the relationship between soil biological activity and temperature. Q_{10} can estimated from annual data sets incorporates not only temperature responses, but also seasonal changes in soil water content, root biomass, litter input, microbial populations.

Q_{10} function by Drewitt *et al*., 2002.

$$R = R_{ref} Q_{10}^{\left(\frac{T - T_{ref}}{10} \right)} \quad (4)$$

$$Q_{10} = \frac{R_{To+10}}{R_{To}} \quad (5)$$

where R is soil respiration, R_{ref} is soil respiration at reference temperature, T is the absolute soil temperature (K), T_{ref}, is reference temperature, R_{To} and R_{To+10} are the respiration rate at reference temperature *To* and temperature T_{o+10}, respectively. When

the relationship between temperature and soil respiration is fitted by an exponential function, Q_{10} can be estimated from coefficient b, as

$$Q_{10} = 10^{b} \quad (6)$$

The estimated values of Q_{10} vary widely from little more than 1 (low sensitive) to more than 10 (high sensitive), depending on the geographic location and ecosystem type. Yuste *et al.* (2004) reported seasonal Q_{10} of soil respiration was much higher under deciduous forest than under evergreen canopies and concluded that the large differences in seasonal Q_{10} were not entirely due to differences temperature sensitivities but also different seasonal patterns of plant activity . In addition, the different temperature sensitivities showed by various components of soil respiration. The possible differential response of microbial and root respiration to temperature could be reflected in the relatively high Q_{10} values (Davidson, 1998). As soil temperature plays a significant role in accounting for the seasonal and daily variations in respiration rates and soil respiration rate also various with soil depths. The A-horizon of forest soil had highest initial rates of soil respiration, follow by the B-and E-horizontal soil. In addition, the initial rates of soil respiration from A-horizon increase with temperature in accordance with the Arrhenius equation (Winkler *et al.,* 1996).

Soil moisture is another important factor influencing soil respiration. Soil respiration could be altered dramatically by changing soil moisture since moisture affects rooting depth, root respiration, and soil microbial community composition. Scientists have explained the effect of moisture availability on soil metabolic activity. Soil respiration is usually low under dry conditions due to low root and microbial activities. In high soil moisture condition, respiration generally increases but soil

moisture may negatively affect respiration rates when it becomes very high. Soil respiration is reduced due to limitation of diffusion of oxygen and suppression of CO_2 emissions. Responded of soil microorganism respiration could be identified from three phases of moisture over time as 1) when soils are relatively dry, metabolic activity increases with increasing moisture availability; 2) when soils are 50-80% saturated, soil biological activity is almost at its potential; 3) when soils are to wet, oxygen deficiencies inhibit aerobic respiration (Raich and Potter, 1995).

The effect of soil moisture on soil respiration has been described by numerous equations, including linear, logarithmic, quadratic and parabolic functions of soil water expressed as metric potential, gravimetric water content, volumetric water content, fraction of water holding capacity, water-filled pore apace, rainfall indices and depth to water table. Davidson *et al.* (2000) reported that soil respiration was correlated with the cube of volumetric water content, which is mechanistically appropriate function for relating soil respiration at below-optimal water content. Soil respiration exhibited pronounced seasonal variations that clearly reflected those of soil water content, with minimum values below 1.6 $\mu molm^{-2}s^{-1}$ alter end of dry season and a maximum value of 5.6 $\mu mol\ m^{-2}s^{-1}$after re-wetting (Epron *et al.,* 2004). The spatial variation in soil respiration at both 10 °C and optimal soil water content were also largely explained by spatial variation in canopy-dependent parameter, basal area, and a soil characteristic, ΔpH (Vincent *et al.,* 2006).

The influencing of rainfall on soil CO_2 efflux

The amount and distribution of rainfall has been shown to be an important controlling factor of soil respiration (Lee *et al.,* 2002; Lee *et al.,* 2004; Harper *et al.,* 2005). Due to climate records suggest that rainfall may expect to increase, with an alteration in the frequency and duration of rain events (Fang *et al.,* 2005). Soil water content can change extremely rapidly during rainfall events. Additionally, soil water content is principle driving variable for soil respiration. An increased in soil water content is assumed to cause an increase in soil respiration. The recent CO_2 efflux field studies have shown that significant changes in soil respiration take place during or after rainfall.

A number of studies in forest, grassland and agricultural ecosystem have indicated that response of soil respiration to rainfall was closely related to (1) degassing or displacement of soil air by rainfall and inhibition of gaseous movement in water saturated soil; (2) translocation, quality and quantity of substrate, (3) decomposition and nitrogen mineralization, and (4) production of CO_2 in the soil due to enhanced microbial activity (Xu *et al.,* 2004; Jassal *et al.,* 2005; Misson *et al.,* 2006). There are some reported that response of soil respiration to rainfall are also rapidly increase by rainfall or add water manipulation. Soil respiration increased immediately, reached a peak and the gradually decreased after water addition into soil, probably resulted from degassing. In addition, the decreased in soil respiration following its peak is fast with the low water addition and slows with the high water addition (Liu *et al.,* 2002). The study of Lee *et al.* (2004) also found that the soil respiration increased rapidly and instantaneously after water was sprayed and returned to pre-irrigation value in <1 hours after the irrigation and the relative contribution of

litter layer to total soil respiration depends strongly on the litter layer soil moisture. Tang *et al.* (2005) also showed that soil respiration immediately and dramatically increased in response to the summer rain after a long drought periods.

Several hours to a few days after rainfalls into dry soil, microbe activities are active, resulting in an increase of soil respiration. This agree with study of Lee *et al.* (2002) found that the soil respiration increased from 380 to 560 mg CO_2 m^{-2} h^{-1} after the rainfall fall and post-rainfall increases in soil respiration represent approximately 16-21% of the annual soil carbon flux. Rochette *et al.* (1991) also state that high CO_2 fluxes from the soil following by rainfall after a dry period and in three hours after the rainfall, soil respiration was nine times higher than before and gradually decreased with time.

Moreover, the post-rainfall increased in soil respiration represent a significant part of annual carbon flux emitted during the growing season and the rainfall reduction or change in rainfall distribution due to climate change will effect soil respiration (Broken *et al.,* 1999). Additionally, both the magnitude and persistence of the soil respiration after rainfall were positively correlated with the amount of rain (Misson *et al.,* 2002). A reduction in rainfall amount usually results in lowered soil respiration. Harper *et al.* (2005) reported that seasonal mean soil respiration decreased by 8% under reduced rainfall amounts in grassland ecosystem. Therefore, changes in the duration between rainfall events may be important in affecting soil respiration. The duration between rainfall events and resulting temporal patterns of soil moisture relative to critical times for microbial activity, biomass accumulation, plant life histories, and other ecological properties may regulate longer-term responses to altered rainfall patterns (Fay *et al.,* 2000). Several studies suggested annual

differences in rainfall could have a significant effect on ecosystem processes such as seedling occurrence and establishment, decomposition and plant penology and growth. It is highly probable that soil respiration rate is affected by rainfall fluctuation in moisture-limited systems.

Carbon efflux measurement techniques

Technically, the rate of CO_2 production is often made at the soil surface to quantify a rate of CO_2 efflux from the soil to the atmosphere. Soil respiration can be measured using several techniques including soil CO_2 profile, micrometeorological (such as eddy covariance), and static and dynamic chamber method. An early method, gas extraction method can provide information on soil CO_2 production at several depths (Davidson and Trumbore, 1995) but this method can not provide in situ, continuous, and convenient data on CO_2 flux and it will disturb the soil environment.

Chamber-based measurements are directly measured CO_2 efflux from soils on a small scale. Chamber systems capture total soil efflux into autotrophic and heterotrophic respirations but sample small areas from spatial heterogeneous soil. There are different types of chamber systems to measure gas fluxes include static chamber system, dynamic chamber systems. Liang *et al.* (2004) compared four approaches for measuring soil respiration in a northern larch (*Larix kaempferi* Sarg.) forest. The four approaches for measuring soil respiration were: (1) a widely used non-steady-state LI-6400 chamber system; (2) a steady-state chamber system with 9 open-top chambers; (3) a steady-state chamber with 16 automated chambers; and (4) a soil CO_2 gradient system. They found that soil respiration measured with the soil CO_2 gradient approach was, on average, 45% higher than the results of the automated

chamber approach, but the correlation between the two technique was good (R^2 = 0.77). Keith *et al.* (2006) measured soil respiration using a non-flow-through steady state chamber with alkali absorption of CO_2 by soda lime compared with a flow-through non-steady IRGA method to assess suitability of using soda lime for field monitoring over large spatial scales and integrated over a day. They found that the soda lime method can be a highly practical method for field measurement if implemented with due care (in term of drying and weighing soda lime and in minimizing leakages). However, chamber-based approach has the advantage of being able to transmit natural pressure fluctuations, which can contribute to the transport of gas from porous surfaces such as a soil (Baldocchi and Meyers, 1991).

Flanagan and Johnson (2005) also compared respiration rate measured with chamber and that determined from nighttime eddy covariance measurement. They found that the chamber method produced slightly higher values than EC method by approximately 4.5% and 13.6% during 2001 and 2002 but respiration rates measured by both techniques showed very similar seasonal patterns of variation in 2001-2002. Myklebust *et al.* (2008) compared measurements of soil respiration using (1) soil chambers, (2) the soil CO_2 gradient technique and ecosystem respiration using the (3) the eddy covariance (EC) method from a surface. The result showed agreement between nocturnal EC and soil respiration measurement over an un-vegetated surface, but soil CO_2 gradient technique measured overall 7% greater values (R^2 = 0.71) than automated chamber method.

The recently, developed soil CO_2 vertical gradient measurement method provides an opportunity to measure soil respiration with high frequency (minutes to half hour) with minimum disturbance to the natural structure of soil. This method is

valuable tools in increasing of various processes governing the CO_2 exchange within the soil. This method has not been widely used earlier probably due to instrument limitations and difficulty in calculating soil respiration from gradient measurement and CO_2 diffusivity in the soil. The method account for CO_2 production in the soil at a number of depths with carbon dioxide measurement probes buried in the ground. The flux of CO_2 diffused from the soil can be calculated by applying Fick's first law of diffusion:

$$F_z = -D_s \frac{dC}{dz}, \tag{7}$$

where F_z is the soil CO_2 efflux, D_s is the gaseous CO_2 diffusion coefficient in the soil that varies with soil, C is the CO_2 mole concentration at a certain depth of the soil, and z is the depth. For flux determination, the gradient is approximated by discrete differences ΔC and Δz.

Industrial solid-state sensors can be used to measure soil CO_2 concentration have become available. Hirano *et al*. (2003) first used a type of these small CO_2 sensors (GMD20, Vaisala Inc., Finland) buried in the soil under a cool-temperate deciduous broadleaf forest to deduce soil respiration, and therefore have demonstrated the feasibility of the instrument. Tang *et al.* (2003) used the new small solid-state CO_2 sensors (GMT222, Vaisala Inc., Finland) to monitor continuously soil CO_2 profiles and soil CO_2 efflux in a dry season in savanna ecosystem by buried these CO_2 sensors at different depths of the soil. They estimated soil respiration based on the measurement of the soil CO_2 gradient and gaseous diffusivity estimated from the Millington-Quirk model. They found that the estimated respiration was very close to chamber measurements and could use for long-term continuous measurements of soil

CO_2 efflux. Similarly, Jassal *et al.* (2005) used small solid-state infrared CO_2 sensors (GMM221, Vaisala Inc., Finland) for long-term continuous real-time measurement of CO_2 concentrations at difference depths, measured half-hourly soil respiration with an automated non-steady-state chamber. They found that soil respiration that calculated from soil CO_2 concentration gradients near the surface closely agreed with the measured efflux from chamber. However, there are no more standardized approaches that suitable for all situations and considerably less information available on CO_2 dynamics below the soil surface, apparently due to the difficulty of sampling and measuring soil CO_2 concentration.

Net Ecosystem Exchange

The exchange, or flux, of carbon between the atmosphere, oceans, and land surface is called the carbon cycle. The global carbon (C) cycle consists of four principal C pools: atmosphere, oceans, reserves of fossil fuels, and terrestrial ecosystems, including vegetation and soils (Fig. 1.1). The atmosphere itself contains nearly 800 billion metric tons of carbon (or Gigatonnes carbon, GtC), which is more carbon than all of the Earth's living vegetation contains (IPCC, 2007). The amount of carbon contained in the living vegetation and soil of terrestrial ecosystems are somewhat less than that present in the atmosphere. Each year the atmosphere carbon exchanges in the system are photosynthetic uptake of ~ 120 GtC/year by terrestrial ecosystems (gross primary productivity or GPP), plant respiration which releases ~ 60 GtC / year back to the atmosphere and heterotrophic (soil) respiration which releases ~ 60 GtC / year. This cycling of carbon is also fundamental to regulating Earth's climate.

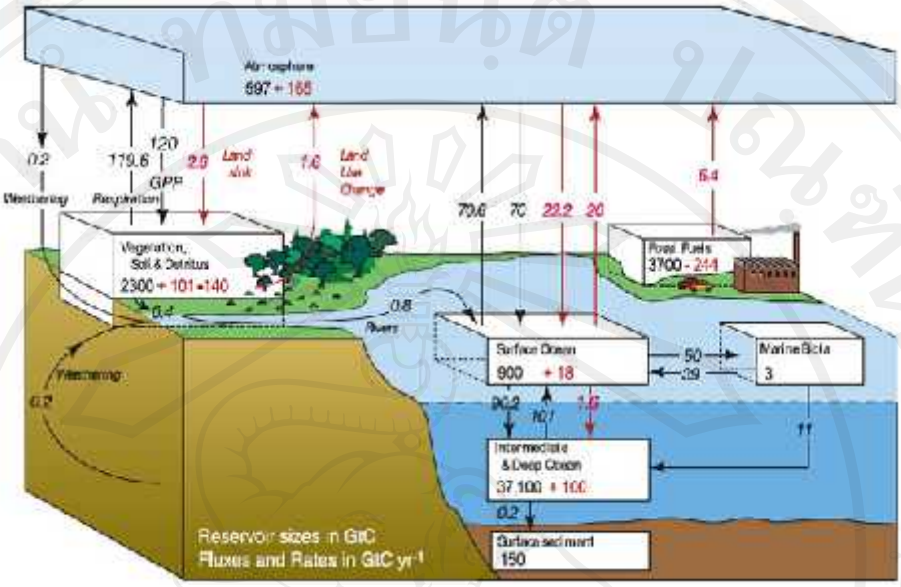

Figure 1.1 Major carbon pools and fluxes of the global carbon balance (Source: © Climate Change 2007: The Physical Scientific Basis, Intergovernmental Panel on Climate Change.)

Carbon dioxide (CO_2) is important as a major contributor to the planetary greenhouse effect and potential climate change. Global climate models predict that future increases in atmospheric CO_2 will cause significant increase in the average global surface temperature of 0.6 °C. Consequently, changes in the amount, distribution and intensity of rainfall/precipitation are also expected to occur (IPCC, 2007). Corresponding changes in air and soil temperature, soil water content and CO_2 concentration, [CO_2] are likely to alter the function of natural and managed ecosystems in terrestrial environments. Recent studies suggests that increased variability in rainfall and soil water content significantly affected C cycling processes such as net photosynthesis, aboveground productivity and soil respiration (Knapp

et al., 2002). Considerable research has been directed at understanding the effects of climate change on structural and physiological dynamics of terrestrial ecosystems (Aubinet *et al.,* 2000; Baldocchi, 2003; Syed *et al.,* 2006). Although research has been conducted in several ecosystems but there is still plenty of gaps in the understanding in respond of terrestrial ecosystem to climate change.

Understanding carbon flux through terrestrial ecosystems is important for many reasons. Recent findings suggest that feedbacks between temperature and moisture availability with climate change are highly possible in terrestrial systems because photosynthesis and respiration respond differently to climatic variables and the balance between them could change with climate changes. Therefore, the knowledge of the amount of CO_2 flux into and out of the atmosphere is important for understanding how ecosystem responds to a change climate. The majority of previous measurements were conducted using micrometeorological techniques such as eddy-covariance methods that represent net values between photosynthesis and respiration.

The carbon cycle in an ecosystem usually initiates when plant fix CO_2 from air through photosynthesis and release back into the atmosphere through plant respiration. At the same time, CO_2 is releases back into atmosphere through microbial respiration from soil (Warembourg and Paul, 1977). The transfer of CO_2 from above a plant surface to the atmosphere is the most important processes of fluxes in terrestrial ecosystem. The flux of CO_2 in the air above a plant surface and the atmosphere is usually measure of the net exchange of CO_2. The net CO_2 flux or Net Ecosystem Exchange (NEE) is the difference between two large terms: (1) the photosynthetic uptake of CO_2 by foliage and (2) the emission of CO_2 by plant and soil respiration. With regards to ecosystem CO_2 exchange, it is not only of interest how much carbon

is assimilated, but also how much CO_2 leaves the ecosystem through respiration. The ecosystem CO_2 exchange during midday is representative for total net carbon uptake by photosynthesis. While, ecosystem CO_2 exchange during night-time is representative for ecosystem respiration, which is characteristic for the CO_2 release of the ecosystem (Fig. 1.2). The net ecosystem exchange has diurnal, seasonal and annual variability.

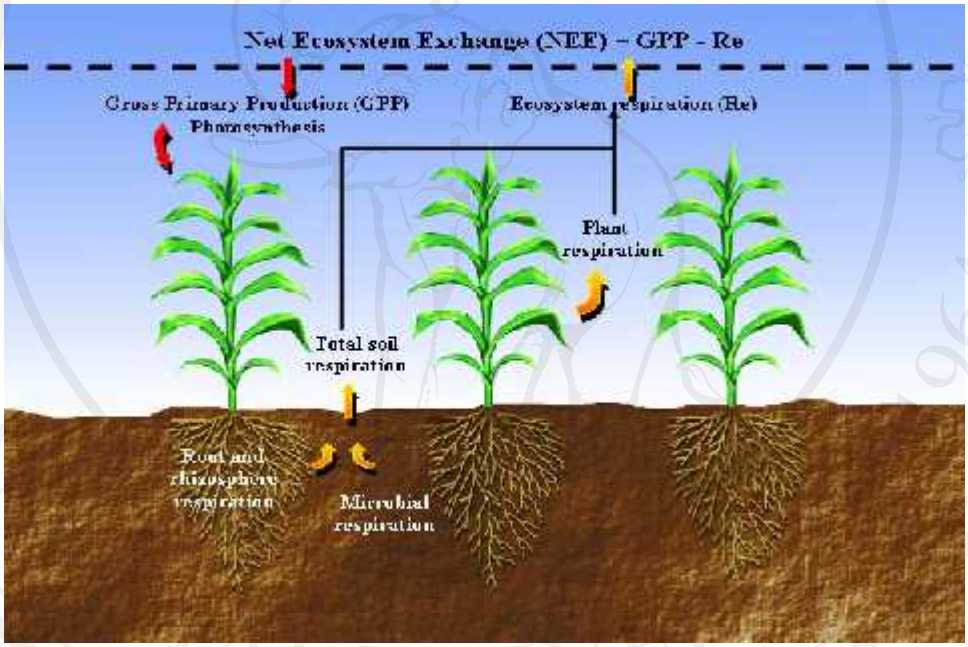

Figure 1.2 Fluxes contributing to the net ecosystem exchange (NEE).

Factors controlling net ecosystem exchange

Changes in climatic conditions influence both net carbon uptakes by photosynthesis and net carbon release by respiration. Photosynthetic uptake and respiration release are separate processes, with different responses to environmental changes. Variation in daytime CO_2 exchanges are primarily controlled by PAR (photosynthetically active radiation), LAI (leaf area index), and soil water content (Carrara *et al.,* 2004; Jaksic *et al.,* 2006). The nighttime CO_2 exchange or ecosystem respiration (Re) is primarily controlled by temperature and soil water content (Xu *et al.,* 2004; Xu and Baldocchi, 2004; Flanagan and Johnson, 2005). Ecosystem respiration is usually the production of autotrophic plant respiration (plant + roots) and heterotrophic respiration (microbial decomposition of soil organic matter). The autotrophic and heterotrophic of ecosystem respiration are responding differentially to variation in precipitation, and availability of substrate but the control exerted by temperature and moisture is dominant factors influence plant, root, and microbial activity. Respiration by plant represents about 50% of the carbon fixed by vegetation (Gifford, 1994). While, soil respiration also contributed about three-quarters of total ecosystem respiration (Law *et al*., 2001). Soil store 2-3 time as much carbon globally as exists in the atmosphere. The soil is also second largest pool in the global carbon cycle, comprising more than twice the estimated pool of carbon in living biomass. Hence, change in soil respiration can result in the net changes of ecosystem exchange and the balance between photosynthesis and ecosystem respiration.

The ecosystem respiration and soil respiration are typically dominated by disparate factors. Ecosystem respiration is controlled by complex interaction of environmental factors and biotic factors such as temperature, moisture, nutrients,

frequency, and type of distribution and varies spatially (i.e. latitudinal and intrasite) and temporally (i.e. daily and seasonal). Temperature and moisture are an important parameter well known to be a dominant environmental control on respiration rates. Many observations show that ecosystem respiration increases exponentially with increasing temperature (Moureaux *et al.,* 2006; Xiaojuan, 2007; Jassal *et al.,* 2007). Additionally, temperature sensitivity coefficient (Q_{10}) of ecosystem respiration decreases with increasing temperature and declining with soil moisture. Davidson *et al.* (2006) reported that seasonal variation in the ratio of soil respiration (R_s) and ecosystem respiration (Re) in forest ecosystem influenced by temperature. They found that the R_s/Re ratio reached a minimum of about 0.45 in early spring, gradually increased through the late spring and early summer, leveled off at about 0.65 for the summer, and then increase again to about 0.8 in autumn. Similarly, Jassal *et al.* (2007) showed that seasonal R_s/Re ratio reached a minimum of 0.52 in spring followed by 0.63 in summer, 0.81 in autumn in an intermediate-aged Douglas-fir stand.

The different responses of those soil respiration and ecosystem respiration to environmental variables arise as a result of seasonal variation in photosynthesis, mobilization and use of stored carbohydrates, and differences in the phrenology of aboveground and belowground plant tissue. Ecosystem respiration rates are also correlated with photosynthesis rates or site productivity. Xiaojuan *et al.* (2007) state that seasonal mean ecosystem respiration in wheat (2.60 $gCm^{-2}day^{-1}$) was much lower than in maize (6.00 $gCm^{-2}day^{-1}$) and ecosystem respiration in both crop increased exponentially with soil temperature. Moreover, the season distributions of daily gross primary production (GPP) and ecosystem respiration were closely liked to the respective variations in green leaf area index (Suyker *et al.,* 2005). In addition,

enhanced soil water content condition caused an increase in maize ecosystem respiration (Verma *et al.,* 2005). The same study of Prueger *et al.* (2003) also concluded that the diurnal exchanges of CO_2 and H_2O in corn and soybean are affected by the soil water content, stage of crop development, and available energy, whereas seasonal changes are caused by the interaction between the soil type (soil water-holding capacity) and management.

The influence of rainfall events on ecosystem exchanges

In water-limited regions, changes in rainfall may have an even greater impact on ecosystem dynamics than the singular effects of rising CO_2 concentration or temperature because the availability of water will have direct impacts on plant recruitment, growth and reproduction, nutrient cycling, and net ecosystem productivity (Knapp *et al.,* 2000; Huxman *et al.,* 2004; Patrick *et al.,* 2007). The highly random and variable precipitation patterns, along with the variation in timing and magnitude can produce unexpected plant growth responses and provide insights into climate change impacts on the ecosystems. For example, plants may increase photosynthetic rates in response to rainfall through an increase in leaf-level CO_2 exchange or through the incremental addition of more leaf area, or both. Fay *et al,* (2003) found that root to shoot ratio and canopy photon flux density at 30 cm above the soil surface were increased in increased rainfall variability, due to reduce in aboveground biomass and stimulate root growth. However, rainfall variability has the potential to impact NEE, through changed in leaf area and stomatal behavior which directly influence the photosynthesis, transpiration and respiration. Niu *et al.* (2007)

state that increased rainfall not only enhanced ecosystem C fluxes, but also ameliorated the negative impacts of climatic warming on ecosystem C fluxes.

Rainfall timing and frequency influence on total carbon uptake are more important than total rainfall. The study of Laporte *et al.* (2002) showed that reduction of the number of monthly rainfall event reduced soil respiration and plant growth through soil moisture deficits. Other studies have reported that the shifts in the season timing or frequency of rainfall may have a larger impact on ecosystem respiration than shifts in the event-size distribution of rainfall (Potts *et al.,* 2006). But Chimner and Welker (2005) found that decreasing in summer rainfall amount reduced ecosystem respiration in a Mixed-grass Prairie. They concluded that ecosystem respiration rates were more limited by soil water content than temperature. Ecosystem respiration may also change after a rainfall due to respiration from plant, plant root and soil respiration. Huxman *et al.* (2004) state that the increase in plant respiration rates immediately following a rainfall event can also significantly contribute to overall ecosystem respiration. This supports Jenerette *et al.* (2008) who found that ecosystem respiration increases following rainfall event with up to 21%, counts observed within 5 days- post rainfall event and Xu *et al.* (2004) who concluded that ecosystem respiration increases with increasing in amount of rainfall, depending on ambient soil water conditions.

Eddy-covariance technique

During the last decade the interest in carbon dioxide, water vapor, and momentum fluxes has increased rapidly in appear with increasing in climate change problem. The various micrometeorological methodologies are carrying out such the vertical turbulent fluxes of carbon dioxide, water, and other scalar entities on both short and long time scales by using flux measurement methods. Several alternative techniques have been developed to estimate fluxes. The most commonly used micrometeorological methods can be separated into four categories: mass balance, flux-gradient, eddy accumulation and eddy covariance (Baldocchi *et al.,* 1988). The most currently explored micrometeorological technique is the eddy covariance method. Environmental scientists are increasingly using eddy covariance as a routine tool for the measurement of surface fluxes of momentum, sensible heat, water vapor, and trace atmospheric constituents such as CO_2.

The eddy-covariance technique is the most direct micrometeorological flux measurement method and provides a direct long –term measure of net carbon dioxide exchange between vegetated canopies and the atmosphere from hourly to inter-annual time scales. (Foken and Wischura, 1996; Baldocchi, 2003). In recent year, the eddy-covariance technique has emerged as an alternative way to assess ecosystem carbon exchange (Running *et al.,* 1999; Canadell *et al.,* 2000; Geider *et al.,* 2001). This technique involves the exchange rate of CO_2 across the interface between the atmosphere and plant canopy by measuring the covariance of the fluctuation in vertical velocity (w) and in the CO_2 mixing ratio or density (c). This technique is also able to measure over short and long times scales (hour, days, seasons, and years) (Wofsy *et al.,* 1993; Baldocchi *et al.,* 2001) and the area sampled, called the flux

footprint, possesses longitudinal dimensions ranging between a hundred meters and several kilometers (Schmid, 1994).

Eddy covariance measurements impose strong technical requirements on sensor technology, including (1) rapid response time that sufficient to capture the high-frequency transport flux; (2) sensor stability that sufficient to integrate continuous measurements without drift over a sufficiently long time to capture the low-frequency transport flux); and (3) high precision, the small fluctuations that constitute the eddy flux signal can be resolved. At present, the eddy covariance technique usually consists of a three-dimensional sonic anemometer that measures fluctuations of wind speed in three directions and an infrared gas analyzer (IRGA) that measures fluctuations of densities of CO_2 and water vapor.

Theory of eddy-covariance technique

The foundations basis of eddy-covariance technique begin with the atmosphere that contains turbulent motions of upward and downward moving air that transport trace gases such as CO_2. This technique measures these turbulent motions to determine the net difference of material moving between the canopy and the atmosphere. In general, turbulent fluxes are calculated as the covariance between the two high frequency time series of vertical wind velocity and a scalar, which can be temperature, humidity or any other trace gas, measured at the same point in space and time. The equation defining the conservation of mass provides theoretical guidance for implementing the eddy-covariance technique (Baldocchi *et al.,* 1988). The principles of measurement begin with the conservation of mass in a system define by the rate of change of the mixing ratio of CO_2 versus the flux of CO_2 in three

dimension, plus a biological source or sink term. The general form of this conservation equation is

$$\frac{d\bar{c}}{dt} = \frac{\partial \bar{c}}{\partial t} + \left[\bar{u}\frac{\partial \bar{c}}{\partial x} + \bar{v}\frac{\partial \bar{c}}{\partial y} + \bar{w}\frac{\partial \bar{c}}{\partial z} \right] = -\left[\frac{\partial F_x}{\partial x} + \frac{\partial F_y}{\partial y} + \frac{\partial F_z}{\partial z} \right] + S_B(x, y, z), \quad (8)$$

I II III IV

In equation (8), the conservation of mass states of CO_2 that the sum of the local time rate of change of the CO_2 mixing ratio (term I),c, and advection (term II) is balance by the sum of the flux divergence of CO_2 in the vertical(z), lateral(y) and longitudinal(x) directions(term III) and the biological source-sink strength (SB)(term IV). $\bar{u}$, $\bar{v}$ and $\bar{w}$ are the vertical velocities in the x, y and z direction, respectively.

The eddy-covariance technique is most accurate when the atmospheric condition (wind, temperature, humidity, CO_2) are steady state, the underlying vegetation is homogenous and it is situated on flat terrain for an extended distance upwind (Baldocchi, 2003). Under ideal condition, steady state ($\partial \bar{c}/\partial t$, so term I =0) and horizontally homogenous (there is no advection, term II) so the horizontal flux divergences $\partial Fx/\partial x$, and $\partial Fy/\partial y$ in term III = 0. Based on these assumptions, the conservation equation simplifies to a balance between the vertical flux divergence of CO_2 and its biological source-sink strength, SB. Thus equation 1 reduces to

$$\frac{\partial F_z}{\partial z} = S_B(z), \quad \text{or}$$

$$\bar{\rho}_a \frac{\partial \overline{w'c'}}{\partial z} = S_B(z), \quad (9)$$

Now, following Baldocchi (2003) and integrating equation 9 with respect to height (z), from ground level (z=0) to some measurement height (z=h) above the canopy.

$$F_z(h) = F_z(0) + \int_0^h S_B(z)dz\,, \qquad \text{or}$$

$$\rho_a \overline{w'c'(z_r)} = \rho_a \overline{w'c'(0)} + \int_0^h S_B(z)dz\,, \qquad (10)$$

where, $F_z(h)$ is the mean vertical turbulent flux density of CO_2 (material) at measurement height, $F_z(0)$ is the net flux density of CO_2 (material) in and out of the underlying soil (soil respiration at ground level), and the integral term is the net storage or source of CO_2 between ground level and height. In practice, $F_z(h)$ is the term that is evaluated as the covariance of w' and c' using the eddy covariance technique.

Net ecosystem exchange (NEE) is given by conservation of mass for total CO_2.

$$NEE = F_z(h) + \overline{\rho_a} \int_0^h \frac{\partial \bar{c}}{\partial t} dz\,, \qquad (11)$$

NEE can be directly determined by add the eddy covariance measurement at height to the time rate of change of concentration measured over specific heights within the canopy up to the height and integrated from z=o to z=h. The eddy covariance measurements of net ecosystem exchange (NEE) of CO_2 (in μ moles m^{-2} s^{-1}) are typically calculated as (Goulden *et al.,* 1996):

$$NEE = \rho \overline{w'[CO_2]'} + \rho \frac{d}{dt} \int_0^h [CO_2](z)dz\,, \qquad (12)$$

Evaluating the accuracy of this technique is complicated. This technique is most accurate when the atmospheric condition (wind, temperature, humidity, CO_2) are steady state, the underlying vegetation is homogenous and it is situated on flat terrain for an extended distance upwind (Baldocchi, 2003). Factors contributing to instrument errors include time response of the sensor, signal to noise ratio, sensor separation distance and height of the measurement. Uncertainties with eddy covariance technique is occurred when the thermal stratification of the atmosphere is stable CO_2 (low turbulent condition) at nighttime. Under this condition, CO_2 exiting leave and the soil may not reach a set of instrument at a reference height, above canopy, causing the eddy covariance technique to underestimate ecosystem respiration (Grace *et al.,*. 1996). The underestimation of CO_2 flux during nighttime periods has led to development of screening criteria for eddy covariance data. For a practical solution, many researchers filter their nighttime measurements based on atmospheric turbulence condition using friction velocity (u*) as an indicator. The measured CO_2 efflux become negligible as u* decreases to zero. Recently, Falge *et al*. (2001) corrected eddy covariance data under stable condition by use u*correction method that based on the rejection of measurements below a certain threshold of friction velocity (u*). Pattey *et al.* (2002) set a limitation for agriculture at a friction velocity threshold of 0.075 to 0.1 ms^{-1} when the turbulence was strong enough to give reasonable flux estimate. The observation of Massman and Lee (2002) found that u* thresholds used at different sites ranged from 0.0 to 0.6 ms^{-1}. However, the threshold of u*seem to be strictly site dependent.

Signal processing used with the eddy-covariance technique

This technique is base on the determination of the statistical correlation of the fluctuations in the wind and carbon dioxide concentration to deduce the vertical flux of the gas. The applicability of the eddy-covariance technique is limited by a member of restrictive assumptions (Baldocchi *et al.,* 1988; Foken and Wichura, 1996). These include horizontal homogeneity of the upwind surface, homogeneity of the turbulence and mean flow, and stationarity. Other factors contributing to instrument errors include time response of the sensor, signal to noise ratio, sensor separation distance, height of the measurement, and signal attenuation due to path averaging. There are different sources of uncertainties in the eddy covariance measurements that sometimes difficult to assess. Therefore, quality test of the raw data, several corrections of the covariance and quality tests for the resulting turbulence fluxes are necessary.

Spike detection

The test for data spikes is the first quality control test. Eddy-covariance measurements are often affected by spikes, due to different reasons both bio-physical (changes in the footprint or fast changes in turbulence conditions) and instrumental (e.g. water drops on sonic anemometer or on open path IRGA) (Papale *et al.,* 2006). Data spikes can be caused by random electronic spikes in the monitoring or recording systems as might during precipitation as well as raindrops blocking the sonic path and indicates faulted data when too many spikes occur. The spikes affecting the single instantaneous measurement are removed before the half-hourly average flux is calculated. Based on previous work of Højstrup (1993) and Vicker and Mahrt (1997),

they considered electronic spikes to have a maximum width of 3 consecutive points in the time series and amplitude of several standard deviations away from the mean. The method computes the mean and standard deviation for a series of moving window of length. The window moves one point at a time through the series. Any point in the window which is more than 3.5 standard deviations from the window mean is considered a spike. The point is replaced using linear interpolation between data points. When 4 or more consecutive points are detected, they are not considered spikes and are not replaced. The entire process is repeated until no more spikes are detected. During the second pass, when the standard deviations may be smaller if spikes were replaced on the previous pass, the threshold for spike detection increases to 3.6 standard deviations and a like amount for each subsequent pass. The threshold of 3.5 standard deviations is limited spike events to 3 or fewer consecutive points.

Planar fit method

Planar fit method is applicable for the rotation of velocity covariance into the streamline coordinate system. The axis rotation applied to eddy-covariance data consists of a rotation of the measured velocity vectors around three axes. The velocity components in the coordinate direction x, y and z will be indicated with u, v and w, respectively. Here x and y are two horizontal direction and z is the vertical. Ideally, a single axis anemometer could be installed to measure only the vertical velocity component, as that is the only velocity signal required for the measurement of scalar fluxes. In practice the coordinate system of the anemometer and the surface will not be perfectly aligned.

Following Wilczak *et al.* (2001) consider a sonic anemometer that is oriented

with its vertical axis perpendicular to the local terrain so that its x and y axes measure the two component of the streamwise flow. If the anemometer is then tilted, we can write.

$$\vec{u}_p = \mathbf{P}(\vec{u}_m - \vec{c}), \tag{13}$$

where $\vec{u}_m$ is the measured wind vector, $\vec{u}_p$ is the wind vector in a mean streamline coordinate system (not yet rotated into the mean wind direction), **P** is a partial rotation matrix that places the *z*-axis perpendicular to the plane of the mean streamline, and $\vec{c}$ is the mean offset error in the measured winds due to instrument error. Note that the planar fit method can only be applied to set of data when the position of the anemometer does not change. If the anemometer is moved or remounted, or if the bias in the vertical component is adjusted during an experiment, then a separate planar fit must be done for each period between changes.

Linear detrend

The linear detrend is most frequently used in calculation of turbulent fluxes. The error of the systematic in flux arising from overlapping of the diurnal cycle with time scales of turbulent motion, changes in meteorological conditions and/or sensor drift which add to turbulent change in data being seen as trend or low-frequency change. The trends contaminating the signal need to remove by suitable detrend method. Indeed, the method for obtaining the fluctuating components for covariance calculations according to Reynolds averaging rules is subtraction of signal from their time average, the way of obtaining the fluctuating components referred to below as mean removal. But simple averaging would lead to overestimation of variance of turbulent quantities, and underestimation of fluxes if trends are present in the time

series. Over a suitable time interval the trend can be approximated as linear and the fluctuations with respect to the regression line can be obtained by linear detrending. Rannik and Vesala (1999) have previously developed schemes of linear detrendind method. In covariance calculations the fluctuations are obtained by subtracting a signal from a realization mean $\bar{x}$, or in the case of detrending/filtering from an instantaneous mean X_t, $x_t' = x_t - X_t$, where $x = w, c$.

In linear detranding, the mean is given by the linear ligression line $X_t = St + I$ over the period T (= $N_i\Delta t$), the regression slope S and intercept I be determined by:

$$S = \frac{N_i \Sigma t x_t - \Sigma t \Sigma x_t}{N_i \Sigma t^2 - (\Sigma t)^2}, \tag{14}$$

$$I = \frac{\Sigma x_t - S\Sigma t}{N_i}, \tag{15}$$

where $t = i\Delta t$ and the summation is made over $i = 1,\ldots, N_i$

WPL correction

Measuring CO_2 fluxes with open-path eddy-covariance sensors typically requires correcting the observed flux data for heat and moisture effects (WPL correction) caused by heat and water vapor transfer to obtain corrected CO_2 fluxes (final CO_2 flux). Historically, the WPL correction was justified by relating the density effects to a mean vertical velocity. Webb *et al.* (1980) published a theory that explained the effects of density fluctuations on flux measurements with devices that measure scalar densities as opposed to scalar mixing ratios. Scalar concentrations in the atmosphere can be expressed in various ways. Among those are densities, ρ, (mass

per unit volume), the mole fraction, χ, (moles scalar per mole humid air) and mixing ratio, r, (moles scalar per mole dry air). However, the open-path CO_2/H_2O gas analyzer dose not measure dimensional CO_2 and H_2O concentrations as mixing ratio, rather it measures CO_2 and H_2O density. Practically, the CO_2/H_2O fluxes evaluated by eddy covariance technique are used to calculate CO_2/H_2O by the mean vertical flow. For this reason, the CO_2/H_2O fluxes need to correct for the mean vertical flow due to air density fluctuation. The key elements of the approach taken by WPL are (a) the ideal gas law and Dalton's law of partial pressures for defines expansion/compression processes in terms of the total number density, and (b) the assumption of zero total dry air flux.

WPL correction basic results can be summarized in the following two equations:

$$E = (1+\frac{\overline{\rho_v} m_a}{\overline{\rho_a} m_v}\frac{m_a}{m_v})(\overline{w'\rho_v'} + \overline{\rho_v}\frac{\overline{w'T'}}{\overline{T}}), \tag{16}$$

$$F_c = \overline{w'\rho_c'} + \frac{m_a}{m_v}\frac{\overline{\rho_c}}{\overline{\rho_a}}\overline{w'\rho_v'} + (1+\frac{\overline{\rho_v}' m_a}{\overline{\rho_a}' m_v})\frac{\overline{\rho_c}}{\overline{T}}\overline{w'T'}, \tag{17}$$

where E and F_c are the latent heat flux and the CO_2 flux after application of the WPL correction. $\overline{w'T'}$, $\overline{w'\rho_v}$, and $\overline{w'\rho_c}$ are the sensible heat flux, latent heat flux, and CO_2 flux measured by eddy-covariance systems, respectively. $\overline{\rho_c}$, $\overline{\rho_a}$, and $\overline{\rho_v}$ are the densities of CO_2, dry air, and water vapor, respectively; m_a and m_v are the molecular mass of dry air and water vapor, respectively. T is the air temperature.