CO2 FLUX IN INDONESIAN WATER DETERMINED BY SATELLITE DATA

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Abstract. The oceans was considered to be a major sink for CO2. The improving of quantitative and qualitative description about the ability of sea in uptaking or emitting CO2 is a great scientific concern in meteorological and climatological science. Measurement of the ability of sea in uptake or emitting CO2 could determined by measuring the CO2 exchange coefficient on sea interface and the measuring the different partial pressure of CO2 between the air and sea. In this study, CO2 flux distribution of Indonesian waters in 2007 to 2009 was computed using monthly CO2 exchange and the different partial pressure of CO2 estimated from wind speed, salinity, SST, and sea characteristic, which were obtained from satellite data. The carbon dioxide flux thus was estimated and discussed by two different designs of transfer velocity (k), of Wanninkhof (1992), kW relationship and by Nightingale et al. (2000), kN relationship. The result indicated that generally, Indonesian water was emitting the CO2 to the air. Average CO2 emitting from sea to the air for recent year in 2007 to 2009 are 3.80 (mol m−2 year−1) and 2.85 (mol m−2 year−1) with kW relationship and kN relationship calculation, respectively. The total average CO2 emission from sea to the air in 2007 to 2009 for the Indonesian waters areas are 0.15 (PgC year−1) and 0.12 (PgC year−1) based on kW relationship and kN relationship calculations, respectively.

Keywords: CO2 flux, salinity, SST, sink and sources of CO2.

1. Introduction

Carbon dioxide (CO2) is a principal greenhouse gas (Frankignoulle et al., 1998). Measurements of the atmospheric CO2 concentration indicated that it has been increasing at a rate about 50% as of which is expected from all industrial CO2 emissions. The oceans have been considered a major sink for CO2. Hence, the improved knowledge of the net transport flux across the air–sea interface is important for understanding the fate of this important greenhouse gas emitted into the earth’s atmosphere (Takahashi et al., 1997).

Meanwhile, the world ocean plays an important role in the earth climate. It does not only absorb heat from the sun, but also plays major role in carbon cycle processes (Akiyama, 2002). For long-term climate forecasts, knowledge of the heat, momentum, and substance exchange between the atmosphere and the ocean is essential (Baliño et al., 2001).

The cycling of carbon between its various organic and inorganic forms and carbon transport from the surface to the deep sea is governed by physical and biological processes. The processes are commonly referred to as the physical (or solubility) pump and the biological pump (Frankignoulle et al., 1998). Both pumps act to increase CO2 concentrations in the ocean interior. The physical pump is driven by the slow overturning circulation of the ocean and by CO2 being more soluble in cold waters. Cold and dense water masses in high latitude oceans, particularly of the North Atlantic and Southern Ocean, absorb atmospheric CO2 before they sink to the ocean interior. The sinking water is balanced by upwelling (vertical transport) in other regions. Upwelled water warms when it reaches the surface where the CO2 becomes less soluble and some is released back to the atmosphere (by a process known as outgassing). The net effect is to pump CO2 into the ocean interior (Takahashi et al., 2002, 2008).

According to Baliño et al. (2001), the oceans currently absorb approximately one-third to one-half of the anthropogenic CO2 emitted from fossil fuels and industries. However, preliminary results from ocean-atmosphere models suggest that we will not be able to rely on the oceans to mop up excess CO2 in the future if current global warming trends continue. If the surface ocean becomes warmer, it may alter the nature and intensity of the ocean circulation or the availability of nutrients. Any such changes will have an impact on the oceans’ ability to absorb and retain the excess CO2 in the future. The fate of anthropogenic carbon emitted into the atmosphere is essential as government’s debate plans for emissions control and the utility of carbon sinks. With 50 times more carbon dioxide than the atmosphere, the ocean contains the largest reservoir of carbon actively circulating in the biosphere. In the long term, the ocean usage...

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plays the dominant role in the natural regulation of CO₂ in the atmosphere and thus exerts a powerful influence on the climate.

The greenhouse phenomenon, for example, is very serious for the global environment and mainly caused by carbon dioxide in the atmosphere. The exchange rate of CO₂ between the atmosphere and ocean is therefore of great importance. The exchange of carbon dioxide between air and sea can be determined from air-sea CO₂ concentration differences and CO₂ gas exchange between air and sea (Takahashi et al., 2002).

The study of air-sea gas exchange is multifaceted (Wanninkhof et al., 2009). Understanding exchange processes at the molecular level has improved our understanding of how environmental factors control the rate of exchange at the air-water interface. Some of the first studies on transfer across liquid-gas surfaces were focus on industrial applications (Danckwerts, 1970). For environmental applications, a significant effort has been devoted to techniques for determining gas transfer rates, processes that control them, theories, and techniques to quantify gas fluxes. For determination of gas exchange over the ocean, many efforts have been of an opportunistic nature and take advantage of O₂ disequilibria that arise from biological productivity (Redfield, 1948) or utilize ¹⁴C excesses in the atmosphere from nuclear bomb tests, called bomb ¹⁴C, that invaded the ocean (Broecker et al., 1985, 1995, Sweeney et al., 2007). Of particular note, there is some improvements in meteorological flux techniques that measure air-sea gas fluxes in the atmospheric boundary layer on minute timescales (McGillis et al., 2001) and waterside, deliberate tracer techniques that can provide gas exchange estimates in the field with timescales on the order of 24 h (Ho et al., 2006).

However, much of the interest in gas fluxes over the ocean relates to regional or global phenomena that occur over seasonal to decadal timescales such that upscaling of field studies is a major consideration. Upscaling involves relating gas transfer to the environmental factors that influence the exchange. Most commonly, upscaling is accomplished by relating gas transfer to wind speed (Wanninkhof, 1992). The upscaling is challenging because of uncertainties in the relationship between gas exchange and wind speed, biases in different wind speed products, and issues with ignoring the cross correlations between gas transfer and wind. There is also increasing evidence that gas transfer cannot be adequately quantified with wind speed alone. Several studies have shown that other effect such as surface films (Broecker et al., 1978, Asher and Pankow, 1986, Frew et al., 1990, 2004, Asher, 1997; Bock et al., 1999, Saylor and Handler, 1999, Zappa et al., 2001, 2004, 2007, Tsai and Liu, 2003). In addition, bubble entrainment (Monahan and Spillane, 1984; Wallace and Wirick, 1992; Farmer et al., 1993; Asher et al., 1996; Zhang et al., 2006; McNeil and d’Asaro, 2007), rain (Ho et al., 1997, 2004; Takagaki and Komori 2007), and boundary layer stability (Erickson, 1993) can affect the gas transfer process.

Much effort has gone into determining a relationship between gas transfer and wind speed, such that gas fluxes between ocean and air can be determined from air-water concentration differences and wind speed over the ocean (Liss and Merlivat, 1986; Wanninkhof, 1992; Takahashi et al., 2002; Eayanti et al., 2009; Eayanti et al., 2011).

Indonesia as an archipelagic country has a great impact from climate change (Susandi et al., 2006). In recent year, a large forest fire and coral reef damage caused by changing of El Nino characteristic as a result of global warming impact. As a tropical country, Indonesia has a large tropical forest that is important for carbon cycle, as an absorbent of CO₂. Meanwhile in recent decade, there is much deforestation in Indonesia, which decrease the ability of forest in absorbing the CO₂ (Panjiwibowo et al., 2003). Beside forest, as archipelagic country, Indonesia also has large seas with total area of of 3.288.680 km² (Susandi et al., 2006). Indonesia as an archipelagic country in equatorial area has a high annual sea surface temperature affecting the partial pressure on sea interface become higher than the partial pressure in the atmosphere that could affect the emission of CO₂ from sea to the air (Takahashi et al., 2002). High annual sea surface temperature in equatorial, also affects the solubility, which is a physic-chemical process that transports carbon (as dissolved inorganic carbon) from the ocean's surface to its interior. The solubility of CO₂ is a strong inverse function of seawater temperature (i.e. solubility is greater in cooler water and lower in warmer water) (Raven and Falkowski, 1999). High annual SST in equatorial area could affect the solubility of carbon dioxide become lower so the dissolved
inorganic carbon from the ocean’s surface to its interior is pursued.

According to Baliño et al. (2001), a physical pump of the CO₂ flux is driven by gas exchange at the air-sea interface and the physical processes that transport CO₂ to the deep ocean. Atmospheric CO₂ enters the ocean by gas exchange depending on wind speed and the differences in partial pressure across the air-sea interface. The amount of CO₂ absorbed by seawater is also a function of temperature through its effect on solubility i.e., solubility increase as temperatures decrease, hence cold surface waters pick up more CO₂ than warm waters.

Qualitative and quantitative study of the ability of Indonesian waters as an absorbent or emission carbon dioxide is very important for Indonesia and in the effort of improving Indonesian negotiation in international consultation about global climate policy (Susandi et al., 2006). In recent year, there are very rare researches in describing and informing CO₂ flux and the ability of Indonesian waters in absorb or emitting CO₂ in qualitative and quantitatively. In this paper, we present CO₂ flux distribution of Indonesian waters in 2007 to 2009 qualitative and quantitatively, computed using monthly data of CO₂ exchange and the different partial pressure of CO₂ (ΔpCO₂) obtained using wind speed, salinity, SST, and sea characteristic by satellite data. The carbon dioxide flux thus is estimated and discussed by two different design of transfer velocity (k). Wanninkhof (1992), kw2, relationship and Nightingale et al. (2000), kN, relationship. Those two different designs are quadratics function of wind. Meanwhile those two functions of wind were used to derive CO₂ transfer velocity in other to get a variation of CO₂ flux distribution.

2. Methods
Carbondioxide flux distribution of Indonesian waters in 2007 to 2009 were computed using monthly CO₂ exchange and ΔpCO₂ was obtained using wind speed and SST monthly derived from satellite data.

2.1 CO₂ transfer velocity
Carbon dioxide transfer velocity derived from Quickscat satellite wind speeds used two k-U relationships, k-U Wanninkhof (1992) and k-U Nightingale et al. (2000).

The Wanninkhof (1992), kw2, relationship was deduced by assuming that k was proportional to U², the global distribution of U was a Rayleigh distribution and the global k average was constrained by the Broecker et al. (1985) ocean ¹⁴C inventory (Wanninkhof, 1992).

The Wanninkhof (1992) parameterization used a quadratic fit to the bomb ¹⁴C inventory. Since high-resolution satellite wind speeds are used, the parameterization suitable for short-term wind speed, would be:

\[ k_{W2} = (0.31 U_{10}^2) \times (660/Sc)^{0.5} \]

Where:
- \( k_{W2} \) = The Wanninkhof CO₂ gas transfer velocity (cm hr⁻¹),
- \( Sc \) = Schmidt Number of radon gas = kinematic viscosity of water divided by diffusion coefficient of gas (dimensionless)
- \( U_{10} \) = wind speed at 10 meters above mean sea level (ms⁻¹)

The Nightingale et al. (2000), kN, relationship was deduced from in situ tracer measurements (SF₆, 3He) performed at sea and assuming a second order polynomial k-U relationship.

\[ k_{N} = (0.222 U_{10}^2 + 0.333 U_{10}) \times (600/Sc)^{0.5} \]

Where:
- \( k_{N} \) = The Nightingale CO₂ gas transfer velocity (cm hr⁻¹),

The transfer of inert gas through the air interface was controlled by the aqueous viscous boundary layer (Witting, 1971; Wanninkhof, 1992). Furthermore, viscosity and diffusivity show an opposite temperature dependence. Through this, diffusion coefficient fitted to a temperature relationship as following formula (Zhao, 1995).

\[ Sc = -8.3x10^{-2}T^3 + 6.7954T^2 - 224.30T + 3412.8 \]

2.2 Carbon dioxide (CO₂) flux calculation
The carbon dioxide flux was calculated as the following formula (Akiyama, 2002):

\[ F = k \times L \times \Delta p CO₂ \]

Where,
- \( F \) = CO₂ flux (mol m⁻² year⁻¹)
- \( \Delta p CO₂ \) = CO₂ partial pressure between ocean and atmosphere (µatm)
- \( L_{CO₂} \) = CO₂ gas solubility (mol liter⁻¹ atm⁻¹).
Wanninkhof (1992) provided an empirical formula to estimate CO\textsubscript{2} gas solubility, \( L \) based on data fitting between the solubility, temperature, and salinity as the following formula:

\[
\ln L = A1 + A2(100/T_{abs}) + A3 \ln(T_{abs}/100) + S ((B1+B2(T_{abs}/100)+B3(T_{abs}/100)^2)
\]

Where, \( T_{abs} = \) absolute temperature \((\text{°K}) = (273.15 + T \text{ °C}) \text{ K} \),
\( S = \) salinity \((\text{psu})\)

The value of active coefficient \( A1-A3 \) and \( B1-B3 \) are shown in Table 1 (Wanninkhof, 1992).

### Table 1. Coefficient for calculation of the solubility of CO\textsubscript{2} in molar and gravimetric units

| No  | \( L \) with unit (moles liter\(^{-1}\) atm\(^{-1}\)) |
|-----|--------------------------------------------------|
| \( A1 \) | -58.0931                                      |
| \( A2 \) | 90.5069                                       |
| \( A3 \) | 22.2940                                       |
| \( B1 \) | 0.027766                                      |
| \( B2 \) | -0.025888                                     |
| \( B3 \) | 0.0050578                                     |

Carbon dioxide partial pressure distribution between the atmosphere and ocean, \( \Delta pCO_2 \), were derived from sea surface temperature. The \( \Delta pCO_2 \) was a function of temperature, total inorganic CO\textsubscript{2} concentration \((T \text{ CO}_2)\), alkalinity, and salinity. Metzl et al. (1995) derived \( pCO_2_{water} \) from the relation with sea surface temperature in Indian Ocean. As \( pCO_2_{air} \) in the atmosphere varied slowly in both spatial and temporal scale, Zhao (1995) derived the difference of CO\textsubscript{2} partial pressure between air and sea \((\Delta pCO_2)\) from sea surface temperature. The relationship between \( \Delta pCO_2 \) and SST was investigated using \textit{in situ} data and MODIS data monthly in year 2007 to 2009. Due to data limitation, related to summer and winter period were derived using least square fit method and the relation between \( \Delta pCO_2 \) and SST can be expressed as:

\[
\Delta pCO_2 = 0.0147 T^4 - 0.1241 T^2 - 12.3453 T + 115.5879
\]

Where,

\( T = \text{Sea Surface Temperature (°C)} \)
\( \Delta pCO_2 = \text{CO}_2 \text{ partial pressure different between atmosphere and ocean.} \)

If seawater \( pCO_2 \) is less than the atmospheric \( pCO_2 \), then seawater takes up CO\textsubscript{2} from the overlying air (indicated by negative \( \Delta pCO_2 \)). If it is greater than the atmospheric \( pCO_2 \), it emits CO\textsubscript{2} to the air (positive \( \Delta pCO_2 \) (Takahashi et al., 2002).

### 3. Results and Discussion

Relationship between sea surface temperature, wind speed, and CO\textsubscript{2} flux in Indonesian waters from 2007 to 2009 are shown in Figure 1. In general, CO\textsubscript{2} flux in Indonesian waters follows the monsoon circulation of Indonesia. December to January is a winter season in the northern hemisphere and a summer season in southern hemisphere. July is the coldest temperature in winter in southern hemisphere and the hottest temperature in summer in northern hemisphere (Clift and Plum, 2008). According to Takahashi et al. (2002), seawater takes up CO\textsubscript{2} from the overlying air when \( pCO_2 \) seawater is less than the atmospheric \( pCO_2 \), and it emits CO\textsubscript{2} to the air when \( pCO_2 \) seawater is greater than the atmospheric \( pCO_2 \).

The research of annual CO\textsubscript{2} flux data in Figure 1 indicated by positive CO\textsubscript{2} flux, in general, Indonesian waters was out gassing the CO\textsubscript{2} to the air (Figure 2(e) and Figure 2(f)). Even though, in some part of Indonesian waters, carbon dioxide was absorbed from air to the sea, the CO\textsubscript{2} absorbed become lower in higher sea surface temperature. In higher SST, water column tended to emit CO\textsubscript{2} (Baliño et al., 2001; Takahashi et al., 1997, 1999, 2002, 2008, and 2009). Average carbon dioxide emitted from sea for recent year of 2007 to 2009 was 3.80 (mol m\(^{-2}\) year\(^{-1}\)) and 2.85 (mol m\(^{-2}\) year\(^{-1}\)) by \( k_{w2} \) relationship and \( k_N \) relationship calculation, respectively. This mean, for the total Indonesian water area which is approximately 3.3 x 10\(^6\) km\(^2\), its total average carbon dioxide emitted from sea to the air in 2007-2009 was almost 0.15 (Pg C year\(^{-1}\)) and 0.12 (Pg C year\(^{-1}\)) by \( k_{w2} \) relationship and \( k_N \) relationship calculations, respectively. The \( k_N \) relationship in deriving CO\textsubscript{2} transfer velocity was used the \( k_{w2} \) relationship, although the value of \( k_{w2} \) was little bit higher than \( k_N \). This result had a similar value with Susandi et al. (2006) which the CO\textsubscript{2} emission from Indonesian sea to the air varied from 2.64 (mol m\(^{-2}\) year\(^{-1}\)) to 3.78 (mol m\(^{-2}\) year\(^{-1}\)).
According to Baliño et al. (2001) the equatorial ocean, such as equatorial Pacific, was the largest continuous natural source of CO$_2$ in the ocean. This was due to the combination of a strong upwelling of CO$_2$-rich waters and low biological activity, as well as due to a strong physical pump combined with a sub-optimal biological pump. Vigorous upwelling along the equator brings cold, CO$_2$-enriched water to the surface. As this water warms up during its journey to the surface, it holds less CO$_2$ and the gas trapped in the water escapes to the air.

Figure 1. Variation in sea surface temperatures, wind speed, and CO$_2$ flux in Indonesian waters from 2007 to 2009
Figure 2. Mean SST (a), wind speed (b), $k_{w92}$ CO$_2$ (c), $k_N$ CO$_2$ (d), CO$_2$ Flux$_{w92}$ (e), and CO$_2$ Flux$_N$ (f) in Indonesian waters in 2007 to 2009.

The North Atlantic, on the other hand, is the most intense region of CO$_2$ uptake in the global ocean. As the Gulf Stream and the North Atlantic Drift transport warm water northwards, it cools and absorbs CO$_2$ from the atmosphere. This region is also one of the more biologically productive ocean regions because of an abundant supply of nutrients. Thus, in contrast to the equatorial Pacific, biological and physical factors combine to create a substantial, though seasonal, net flux of CO$_2$ from the atmosphere into the North Atlantic and North Pacific. The Southern Ocean is another important uptake region where the cold surface water masses sink and biological activity is sometimes intense. Seasonal changes in the flux also occurred and the fluxes were affected on longer time scales in response to large-scale oceanic and atmospheric perturbations like the El Niño-Southern Oscillation (ENSO) cycles (Takahashi et al., 1999). Additionally, by Takahashi et al. (2002), the new global uptake flux obtained with the Wanninkhof (wind speed) dependence was compared with those obtained previously using a smaller number of measurements, about 250,000 and 550,000, respectively, and were found to be consistent within ±0.2 (Pg C year$^{-1}$). This estimate for the global ocean uptake flux was consistent with the values of 2.0±0.6 (Pg C year$^{-1}$) estimated based on the observed changes in the atmospheric CO$_2$ and oxygen concentrations during the 1990s.
According to Takahashi et al. (2008), the annual sea surface temperature in equatorial area affected the partial CO₂ pressure on ocean surface, becoming higher than the partial pressure in the atmosphere that could affect the CO₂ emission from sea. The annual SST in Indonesian waters as an equatorial area with annual average value of 29.34°C to 31.94°C (Figure 2(a)), affected the ΔpCO₂ between the atmosphere and ocean in positive value (mean ΔpCO₂ = +12.97 µatm to +39.29 µatm). This SST condition could affect the CO₂ outgassing from sea, which also supported by high wind speed (from 6.02 ms⁻¹ to 8.44 ms⁻¹) (Figure 2(b)). The high wind speed could cause the higher annual CO₂ transfer velocity (k₇₀₂ = 12.42 cm hr⁻¹ to 23.24 cm hr⁻¹ and k₉ = 10.55 cm hr⁻¹ to 18.61 cm hr⁻¹) (Figure 2(c) and 2(d)). High annual SST in Indonesian waters also affected the CO₂ solubility with strong inverse function of seawater temperature (Raven and Falkowski, 1999). The solubility of carbon dioxide become lower with annual solubility of 0.026 mol liter⁻¹ atm⁻¹).

Maximum CO₂ absorption in the South China Sea in December 2008 was shown in Figure 3(e) (k₇₀₂) and Figure 3(f) (k₉). Using the k₇₀₂ relationship, maximum value of CO₂ absorption was 46.96 (mol m⁻² year⁻¹). Using the k₉ relationship, maximum CO₂ flux was 36.14 (mol m⁻² year⁻¹). The SST values of 22.51°C to 29.32°C (see Figure 3(a)) could produce negative value of ΔpCO₂ (mean ΔpCO₂ = -40.01 µatm). The negative ΔpCO₂ indicated that the sea could absorb CO₂ from air in a maximum value. In
addition, high wind speed, from 6.13 ms\(^{-1}\) to 12.0 ms\(^{-1}\) (see Figure 3(b)) could make the CO\(_2\) transfer velocity higher (\(k_{W92} = 12.0\) cm hr\(^{-1}\) to 42.29 cm hr\(^{-1}\) and by \(k_N = 10.14\) cm hr\(^{-1}\) to 32.49 cm hr\(^{-1}\)), as shown in Figure 3(c) and 3(d), respectively.

In southern hemisphere of Indonesian waters, although the CO\(_2\) transfer velocity was smaller than in northern hemisphere (\(k_{W92} = 1.07\) cm hr\(^{-1}\) to 12.0 cm hr\(^{-1}\) and by \(k_N = 1.34\) cm hr\(^{-1}\) to 10.14 cm hr\(^{-1}\)), summer season caused big effect in southeastern part of Indonesian waters for SST from 29.32\(^{\circ}\)C to 35.31\(^{\circ}\)C. This sea surface temperature could increase the \(\Delta pCO_2\) (mean \(\Delta pCO_2 = +79.94\) µatm) which mean that CO\(_2\) was emitted to the air. With \(k_{W92}\) relationship maximum value of CO\(_2\) flux from the sea to the air was 32.00 (mol m\(^{-2}\) year\(^{-1}\)) and with \(k_N\) maximum CO\(_2\) flux was 27.14 (mol m\(^{-2}\) year\(^{-1}\)) (Figure 5).

During southeast monsoon, in northern hemisphere, the sea surface temperature becomes higher than in southern hemisphere.

Figure 4. SST (a), wind speed (b), \(k_{W92}\) CO\(_2\) (c), \(k_N\) CO\(_2\) (d), Maximum CO\(_2\) uptake (CO\(_2\) Flux\(_{W92}\)) (e), and CO\(_2\) uptake (CO\(_2\) Flux\(_N\)) (f) in Indonesian waters at July 2008.
Maximum CO₂ emission in the northern hemisphere of Indonesia (area of South China Sea) in July 2009 was shown in Figure 4(e) \( (k_{W92}) \) and Figure 4(f) \( (k_N) \). With \( k_{W92} \) relationship, maximum value of CO₂ emission was 46.17 (mol m\(^{-2}\) year\(^{-1}\)) and with \( k_N \) relationship maximum CO₂ flux was 35.22 (mol m\(^{-2}\) year\(^{-1}\)). The SST at that time was 29.80°C to 35.20°C (see Figure 4(a)) could change \( \Delta pCO_2 \) in positive value (mean \( \Delta pCO_2 = +78.56 \) µatm), which means that the sea was emitting CO₂ to the air in a maximum value. In addition, high wind speed, from 6.83 m s\(^{-1}\) to 14.80 m s\(^{-1}\) (see Figure 4(b)) could make the CO₂ transfer velocity higher \( (k_{W92} = 14.44 \) cm hr\(^{-1}\) to 74.34 cm hr\(^{-1}\) and \( k_N = 12.79 \) cm hr\(^{-1}\) to 55.90 cm hr\(^{-1}\) ), respectively (Figure 4(c) and 4(d)).

Meanwhile, in southern hemisphere of Indonesian waters, although the CO₂ transfer velocity was smaller than in northern hemisphere \( (k_{W92} = 3.23 \) cm hr\(^{-1}\) to 15.45 cm hr\(^{-1}\) and by \( k_N = 3.31 \) cm hr\(^{-1}\) to 12.79 cm hr\(^{-1}\) ), winter season caused big effect in southeastern part of Indonesia waters for SST from 25.86°C to 29.80°C. This SST condition could affect the \( \Delta pCO_2 \) in negative value (mean \( \Delta pCO_2 = -17.05 \) µatm) meaning that sea could absorb CO₂ from
air. With $k_{W2}$ relationship maximum value of CO$_2$ absorb from air to the sea was 11.76 (mol m$^{-2}$year$^{-1}$) and with $k_y$ relationship maximum CO$_2$ flux is 9.37 (mol m$^{-2}$year$^{-1}$) (Figure 5).

4. Conclusion

Generally, Indonesian waters were emitting the CO$_2$ with average of almost 3.80 (mol m$^{-2}$year$^{-1}$) or 0.15 (Pg C year$^{-1}$) and 2.85 (mol m$^{-2}$year$^{-1}$) or 0.15 (Pg C year$^{-1}$) for the Indonesian waters from 2007 to 2009, with $k_{W2}$ relationship and $k_y$ relationship calculation, respectively. Maximum CO$_2$ absorption occurred in northern hemisphere (the area of South China Sea) in December 2008. With $k_{W2}$ relationship, maximum value of CO$_2$ flux from the atmosphere to the sea was 46.96 (mol m$^{-2}$year$^{-1}$) and with $k_y$ relationship maximum CO$_2$ flux was 36.14 (mol m$^{-2}$year$^{-1}$). Meanwhile in southern hemisphere, CO$_2$ was emitted to the air, with $k_{W2}$ relationship maximum value of CO$_2$ flux from the sea to the air was 32.00 (mol m$^{-2}$year$^{-1}$) and with $k_y$ relationship maximum CO$_2$ flux was 27.14 (mol m$^{-2}$year$^{-1}$). Maximum CO$_2$ emission in the northern hemisphere of Indonesia (area of South China Sea) occurred in July 2009. With $k_{W2}$ relationship, maximum value of CO$_2$ flux from the sea to the air was 46.17 (mol m$^{-2}$year$^{-1}$) and with $k_y$ relationship maximum CO$_2$ flux was 35.22 (mol m$^{-2}$year$^{-1}$). Meanwhile in southern hemisphere, CO$_2$ was absorbed with $k_{W2}$ relationship maximum value of 11.76 (mol m$^{-2}$year$^{-1}$) and with $k_y$ relationship maximum value of 9.37 (mol m$^{-2}$year$^{-1}$). Additionally, CO$_2$ flux in Indonesia waters seemed to follow the monsoon circulation in Indonesia. The $k_y$ relationship in deriving CO$_2$ transfer velocity seemed to follow the $k_{W2}$ relationship, although the value of $k_{W2}$ was little higher than $k_y$ [1 (mol m$^{-2}$year$^{-1}$) or 0.03 (PgC year$^{-1}$)].

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