

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), k_{W92} relationship and by Nightingale *et al.* (2000), k_N 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⁻²year⁻¹) and 2.85 (mol m⁻²year⁻¹) with k_{W92} relationship and k_N 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 k_{W92} relationship and k_N 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 focused 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 effects 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 Wirrick, 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; Ekayanti *et al.*, 2009; Ekayanti *et al.*, 2011).

Indonesia as an archipelagic country has a great impact from climate change (Susandi *et al.*, 2006). In recent years, a large forest fire and coral reef damage caused by changing of El Niño 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 decades, there is much deforestation in Indonesia, which decreases the ability of forest in absorbing the CO₂ (Panjiwibowo *et al.*, 2003). Besides forest, as an archipelagic country, Indonesia also has large seas with a total area 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 physico-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 (ΔpCO_2) 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), k_{w92} , relationship and Nightingale *et al.* (2000), k_N , 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

Carbondioxyde flux distribution of Indonesian waters in 2007 to 2009 were computed using monthly CO₂ exchange and ΔpCO_2 was obtained using wind speed and SST monthly derived from satellite data.

2.1 CO₂ transfer velocity

Carbondioxyde transfer velocity derived from Quikscat satellite wind speeds used two k - U relationships, k - U Wanninkhof (1992) and k - U Nightingale *et al.* (2000).

The Wanninkhof (1992), k_{w92} , relationship was deduced by assuming that k was proportional to U^2 , 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:

$$k_{w92} = (0.31 U_{10})^2 \times (660/Sc)^{0.5} \quad (1)$$

Where:

K_{w92} = 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), k_N , relationship was deduced from *in situ* tracer measurements (SF₆, 3He) performed at sea and assuming a second order polynomial k - U relationship.

$$k_N = (0.222 U_{10}^2 + 0.333 U_{10}) \times (600/Sc)^{0.5} \quad (2)$$

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).

$$Sc = -8.3 \times 10^{-2} T^3 + 6.7954 T^2 - 224.30 T + 3412.8 \quad (3)$$

2.2 Carbon dioxide (CO₂) flux calculation

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

$$F = k \times L \times \Delta pCO_2 \quad (4)$$

Where,

F = CO₂ flux (mol m⁻² year⁻¹)

ΔpCO_2 = CO₂ partial pressure between ocean and atmosphere (μ atm)

L_{CO_2} = CO₂ gas solubility (mol liter⁻¹ atm⁻¹).

Wanninkhof (1992) provided an empirical formula to estimate CO₂ gas solubility, L based on data fitting between the solubility, temperature, and salinity as the following formula:

$$\ln L = A1 + A2(100/T_{abs}) + A3 \ln(T_{abs}/100) + S / ((B1 + B2(T_{abs}/100) + B3(T_{abs}/100)^2) \quad (5)$$

Where, T_{abs} = absolute temperature (°K) = (273.15 + T °C) K,
 S = salinity (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 CO₂ 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, ΔpCO_2 , were derived from sea surface temperature. The ΔpCO_2 was a function of temperature, total inorganic CO₂ concentration ($T CO_2$), 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 CO₂ partial pressure between air and sea (ΔpCO_2) from sea surface temperature. The relationship between ΔpCO_2 and SST was investigated using *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 ΔpCO_2 and SST can be expressed as:

$$\Delta pCO_2 = 0.0147 T^3 - 0.1241 T^2 - 12.3453 T + 115.5879 \quad (6)$$

Where,

T = Sea Surface Temperature (°C)

ΔpCO_2 = CO₂ partial pressure different between atmosphere and ocean.

If seawater pCO_2 is less than the atmospheric pCO_2 , then seawater takes up CO₂ from the overlying air (indicated by negative ΔpCO_2). If it is greater than the atmospheric pCO_2 , it emits CO₂ to the air (positive ΔpCO_2) (Takahashi *et al.*, 2002).

3. Results and Discussion

Relationship between sea surface temperature, wind speed, and CO₂ flux in Indonesian waters from 2007 to 2009 are shown in Figure 1. In general, CO₂ 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 Plumb, 2008). According to Takahashi *et al.* (2002), seawater takes up CO₂ from the overlying air when pCO_2 seawater is less than the atmospheric pCO_2 , and it emits CO₂ to the air when pCO_2 seawater is greater than the atmospheric pCO_2 .

The research of annual CO₂ flux data in Figure 1 indicated by positive CO₂ flux, in general, Indonesian waters was out gassing the CO₂ 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 CO₂ absorbed become lower in higher sea surface temperature. In higher SST, water column tended to emit CO₂ (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⁻²year⁻¹) and 2.85 (mol m⁻²year⁻¹) by k_{w92} relationship and k_N relationship calculation, respectively. This mean, for the total Indonesian water area which is approximately 3.3 x 10⁶ km², 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_{w92} relationship and k_N relationship calculations, respectively. The k_N relationship in deriving CO₂ transfer velocity was used the k_{w92} relationship, although the value of k_{w92} was little bit higher than k_N . This result had a similar value with Susandi *et al.* (2006) which the CO₂ emission from Indonesian sea to the air varied from 2.64 (mol m⁻²year⁻¹) to 3.78 (mol m⁻²year⁻¹).

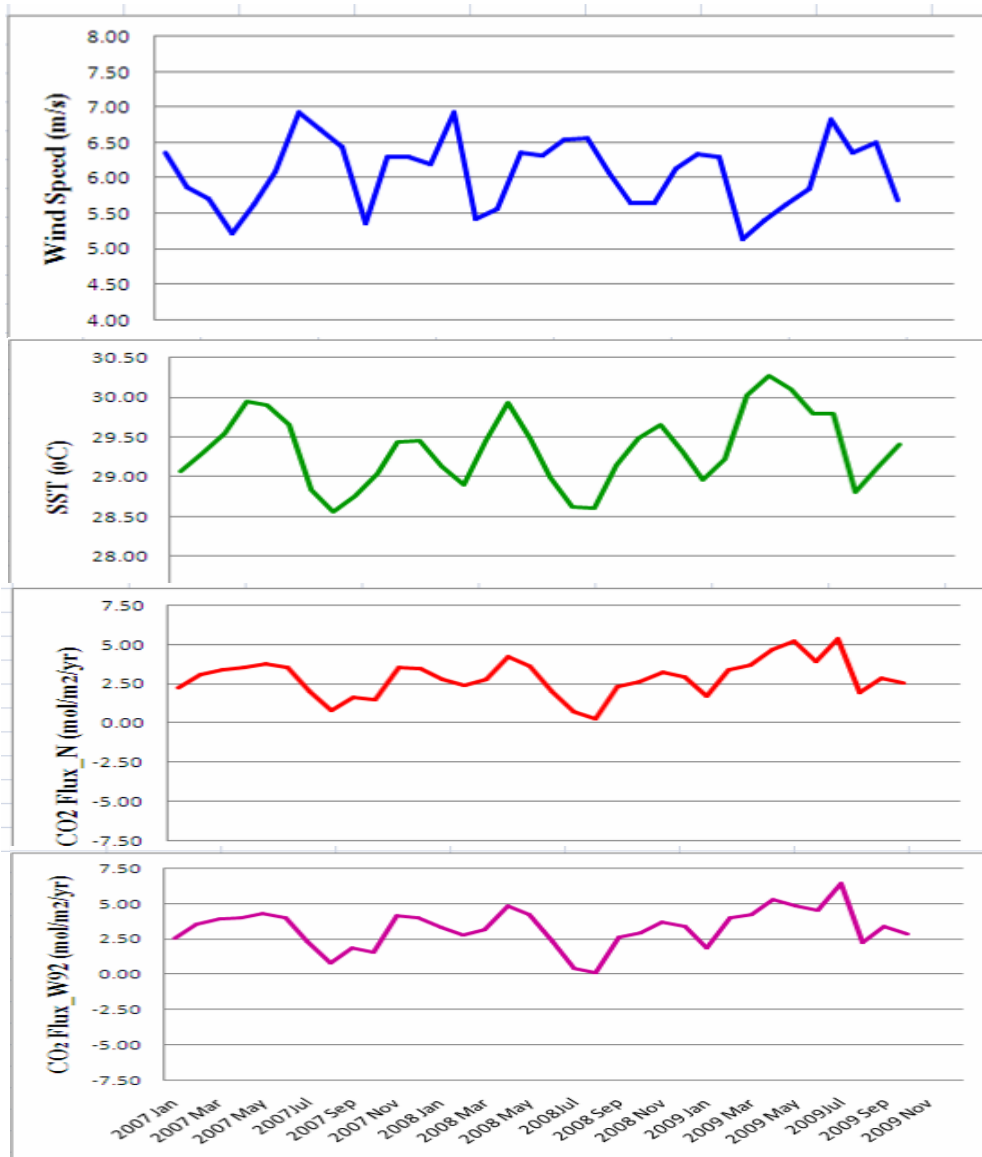


Figure 1. Variation in sea surface temperatures, wind speed, and CO₂ flux in Indonesian waters from 2007 to 2009

According to Baliño *et al.* (2001) the equatorial ocean, such as equatorial Pacific, was the largest continuous natural source of CO₂ in the ocean. This was due to the combination of a strong upwelling of CO₂-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, CO₂-enriched water to the surface. As this water warms up during its journey to the surface, it holds less CO₂ and the gas trapped in the water escapes to the air.

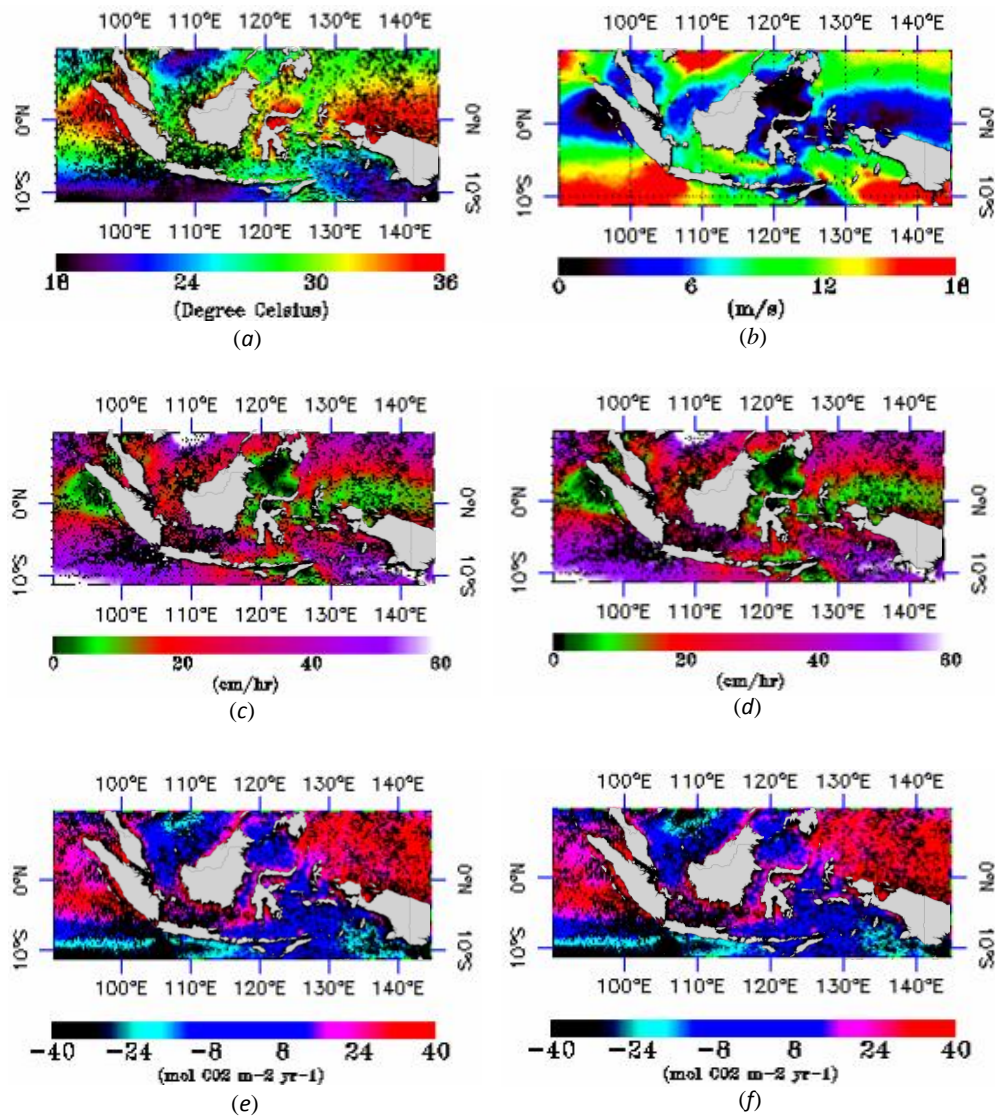


Figure 2. Mean SST (a), wind speed (b), $k_{W92} \text{ CO}_2$ (c), $k_N \text{ CO}_2$ (d), $\text{CO}_2 \text{ Flux}_{W92}$ (e), and $\text{CO}_2 \text{ Flux}_N$ (f) in Indonesian waters in 2007 to 2009.

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.

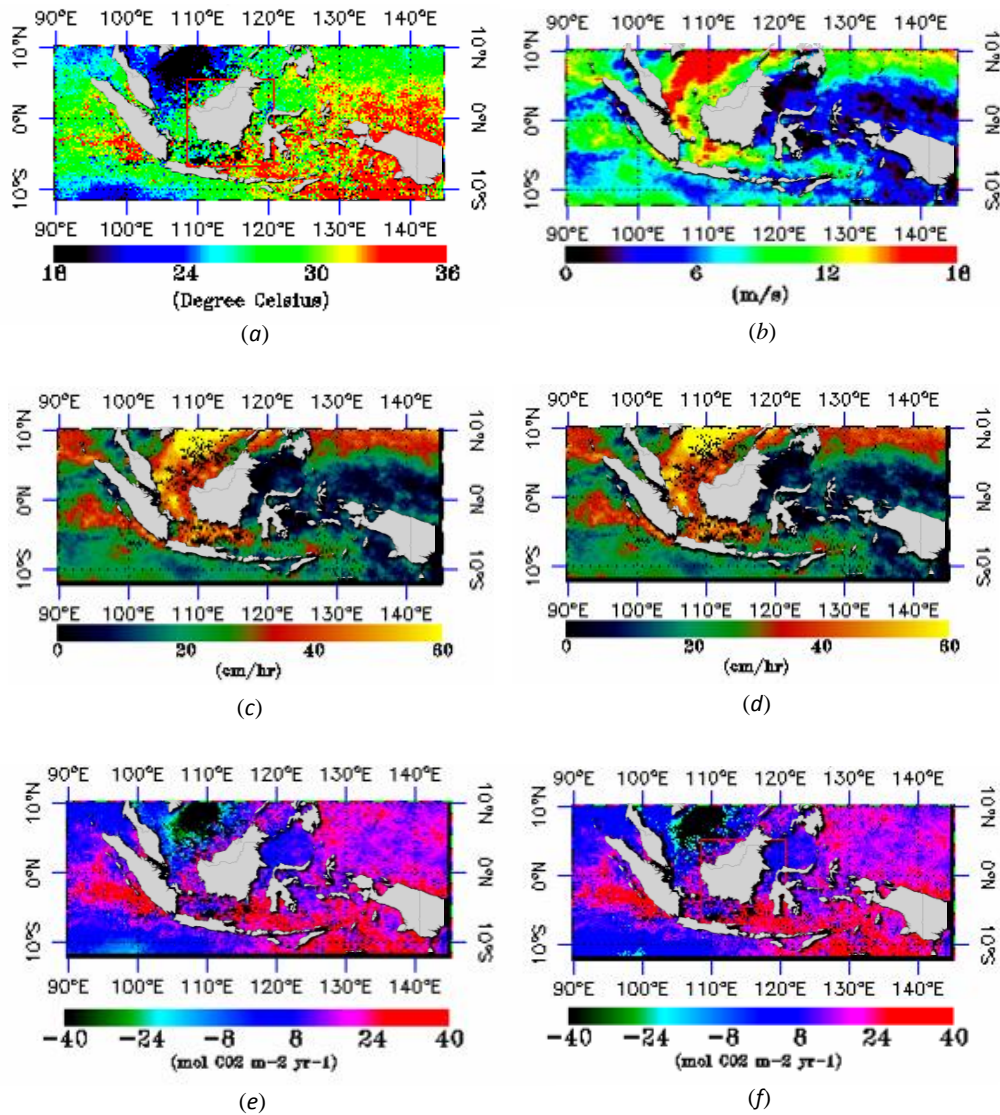


Figure 3. SST (a), wind speed (b), k_{W92} CO₂ (c), k_N CO₂ (d), Maximum CO₂ uptake (CO₂ Flux_{W92}) (e), and CO₂ uptake (CO₂ Flux_N) (f) in Indonesian waters at December 2008.

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 ΔpCO_2 between the atmosphere and ocean in positive value (mean $\Delta pCO_2 = +12.97 \mu\text{atm}$ to $+39.29 \mu\text{atm}$). 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 ($k_{W92} = 12.42 \text{ cmhr}^{-1}$ to 23.24 cmhr^{-1} and $k_N = 10.55 \text{ cmhr}^{-1}$ to 18.61 cmhr^{-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 $0.026 \text{ mol liter}^{-1} \text{ atm}^{-1}$).

Maximum CO₂ absorption in the South China Sea in December 2008 was shown in Figure 3(e) (k_{W92}) and Figure 3(f) (k_N). Using the k_{W92} relationship, maximum value of CO₂ absorption was 46.96 (mol m⁻² year⁻¹). Using the k_N 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 ΔpCO_2 (mean $\Delta pCO_2 = -40.01 \mu\text{atm}$). The negative ΔpCO_2 indicated that the sea could absorb CO₂ from air in a maximum value. In

adition, 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 ($k_{W92} = 12.0 \text{ cmhr}^{-1}$ to 42.29 cmhr^{-1} and by $k_N = 10.14 \text{ cmhr}^{-1}$ to 32.49

cmhr^{-1}), as shown in Figure 3(c) and 3(d), respectively.

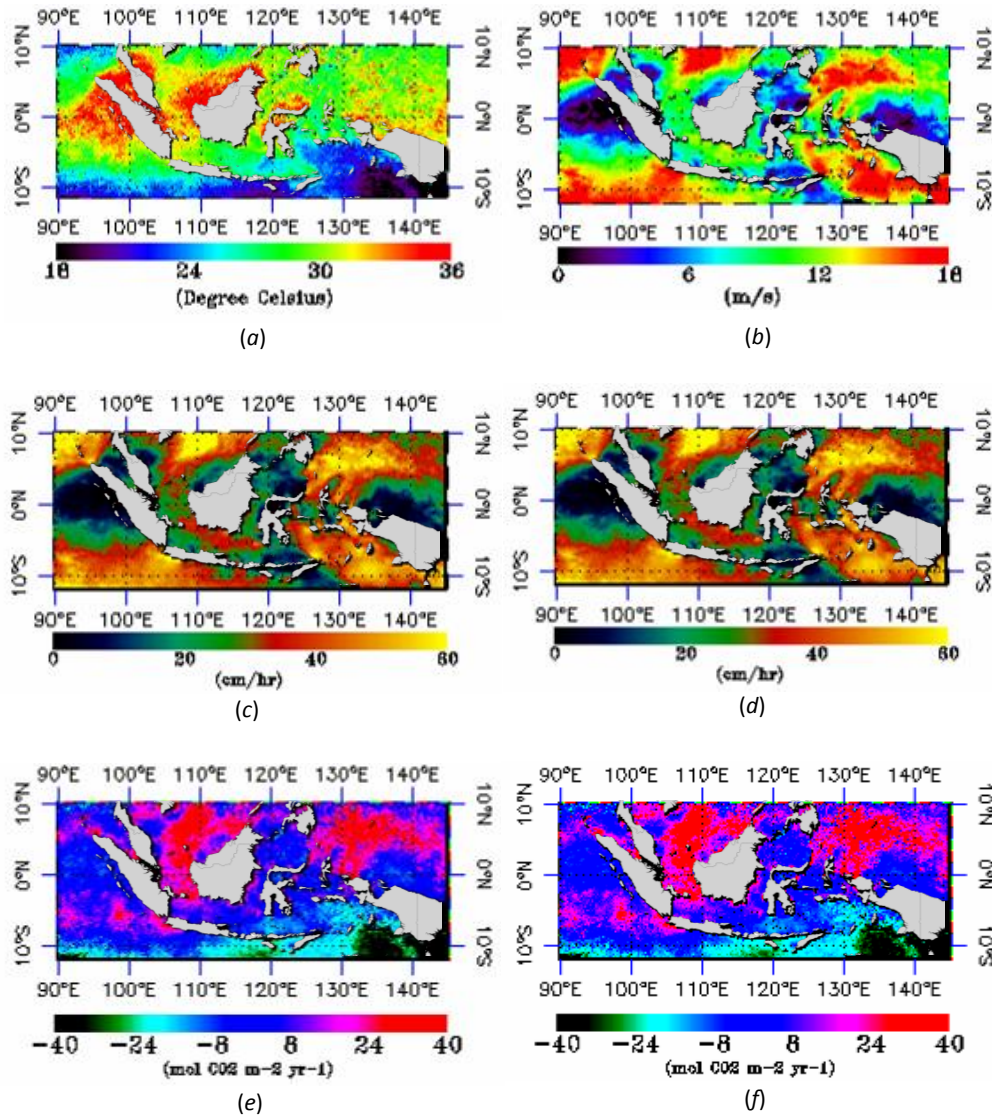


Figure 4. SST (a), wind speed (b), k_{W92} CO_2 (c), k_N CO_2 (d), Maximum CO_2 uptake ($\text{CO}_2 \text{ Flux}_{W92}$) (e), and CO_2 uptake ($\text{CO}_2 \text{ Flux}_N$) (f) in Indonesian waters at July 2008.

In southern hemisphere of Indonesian waters, although the CO_2 transfer velocity was smaller than in northern hemisphere ($k_{W92} = 1.07 \text{ cmhr}^{-1}$ to 12.0 cmhr^{-1} and by $k_N = 1.34 \text{ cmhr}^{-1}$ to 10.14 cmhr^{-1}), summer season caused big effect in southeastern part of Indonesian waters for SST from 29.32°C to 35.31°C . This sea surface temperature could increase the $\Delta p\text{CO}_2$ (mean $\Delta p\text{CO}_2 = +79.94 \mu\text{atm}$) 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 $32.00 \text{ (mol m}^{-2}\text{year}^{-1}\text{)}$ and with k_N maximum CO_2 flux was $27.14 \text{ (mol m}^{-2}\text{year}^{-1}\text{)}$ (Figure 5).

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

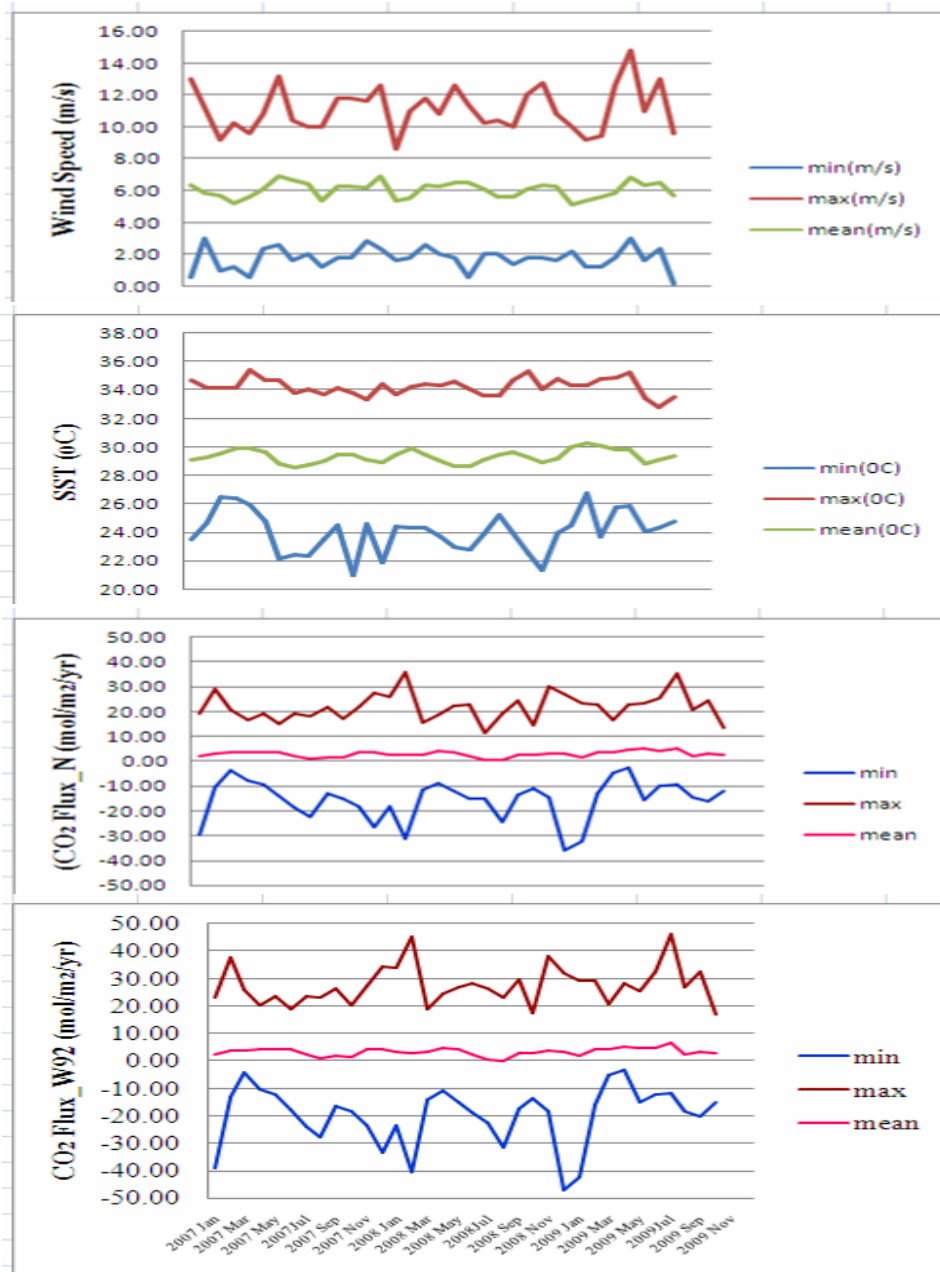


Figure 5. Minimum, mean, and maximum value of sea surface temperatures, wind speed, and CO₂ flux in Indonesia waters from 2007 to 2009.

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⁻²year⁻¹) and with k_N relationship maximum CO₂ flux was 35.22 (mol m⁻²year⁻¹). The SST at that time was 29.80°C to 35.20°C (see Figure 4(a)) could change ΔpCO_2 in positive value (mean $\Delta pCO_2 = +78.56 \mu\text{atm}$), which means that the sea was emitting CO₂ to the air in a maximum value. In addition, high wind speed, from 6.83 ms⁻¹ to 14.80 ms⁻¹ (see Figure 4(b)) could make the CO₂

transfer velocity higher ($k_{W92} = 14.44 \text{ cmhr}^{-1}$ to 74.34 cmhr⁻¹ and $k_N = 12.79 \text{ cmhr}^{-1}$ to 55.90 cmhr⁻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 ($k_{W92} = 3.23 \text{ cmhr}^{-1}$ to 15.45 cmhr⁻¹ and by $k_N = 3.31 \text{ cmhr}^{-1}$ to 12.79 cmhr⁻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 ΔpCO_2 in negative value (mean $\Delta pCO_2 = -17.05 \mu\text{atm}$) meaning that sea could absorb CO₂ from

air. With k_{w92} relationship maximum value of CO₂ absorb from air to the sea was 11.76 (mol m⁻²year⁻¹) and with k_N relationship maximum CO₂ flux is 9.37 (mol m⁻²year⁻¹) (Figure 5).

4. Conclusion

Generally, Indonesian waters were emitting the CO₂ with average of almost 3.80 (mol m⁻²year⁻¹) or 0.15 (Pg C year⁻¹) and 2.85 (mol m⁻²year⁻¹) or 0.15 (Pg C year⁻¹) for the Indonesian waters from 2007 to 2009, with k_{w92} relationship and k_N relationship calculation, respectively. Maximum CO₂ absorbtion occurred in northern hemisphere (the area of South China Sea) in December 2008. With k_{w92} relationship, maximum value of CO₂ flux from the atmosphere to the sea was 46.96 (mol m⁻²year⁻¹) and with k_N relationship maximum CO₂ flux was 36.14 (mol m⁻²year⁻¹). Meanwhile in southern hemisphere, CO₂ was emitted to the air, with k_{w92} relationship maximum value of CO₂ flux from the sea to the air was 32.00 (mol m⁻²year⁻¹) and with k_N relationship maximum CO₂ flux was 27.14 (mol m⁻²year⁻¹). Maximum CO₂ emission in the northern hemisphere of Indonesia (area of South China Sea) occurred in July 2009. With k_{w92} relationship, maximum value of CO₂ flux from the sea to the air was 46.17 (mol m⁻²year⁻¹) and with k_N relationship maximum CO₂ flux was 35.22 (mol m⁻²year⁻¹). Meanwhile in southern hemisphere, CO₂ was absorbed with k_{w92} relationship maximum value of 11.76 (mol m⁻²year⁻¹) and with k_N relationship maximum value of 9.37 (mol m⁻²year⁻¹). Additionally, CO₂ flux in Indonesia waters seemed to follow the monsoon circulation in Indonesia. The k_N relationship in deriving CO₂ transfer velocity seemed to follow the k_{w92} relationship, although the value of k_{w92} was little higher than k_N [1 (mol m⁻²year⁻¹) or 0.03 (PgC year⁻¹)].

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