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Seasonal variation of the CO_2 exchange coefficient over the global ocean using satellite wind speed measurements

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ABSTRACT

We used one year (July 1987 to June 1988) of wind speed measurements made by the microwave radiometer SSM/I to determine the CO_2 exchange coefficient between air and sea on a global scale through the Liss and Merlivat relationship. We determined the seasonal variation in every region of the world ocean. It can be as high as a factor of 4 in some areas. An estimate of the accuracy of the retrieved wind speeds showed that this variation is likely to be underestimated. The global averaged CO_2 exchange coefficient obtained is $3.34 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}, close to previous estimates. A study of the errors on the retrieved wind speed showed that the exchange coefficient is likely to be underestimated and could be as high as 4 10^{-2} mol m^{-2} yr^{-1} μatm^{-1}. We combined these results with various estimates of the annual CO_2 partial pressure gradient at the air sea surface and get a net flux absorbed by the ocean. The flux so determined is not meaningful, however, since the covariant variability of the exchange coefficient and of the CO_2 partial pressure gradient is not taken into account.

1. Introduction

The recent estimation by Liss and Merlivat (1986) of a relationship between the CO_2 transfer velocity and the wind speed over the ocean, has made it possible to deduce the CO_2 transfer velocity, a quantity that is presently hard to measure in the open ocean, from the sea surface wind speed. With the advent of satellite-borne instruments able to measure the wind at the ocean surface, wind speed fields with a global coverage and a repetition period of a few days are now available. A first global map of the exchange coefficient, that is the transfer velocity multiplied by the solubility of CO_2 in sea water, was made by Etcheto and Merlivat (1988) using 3 months of satellite data. The present paper uses the first year of data from a satellite launched in June 1987 to derive the seasonal variation of the exchange coefficient on a global scale.

2. Data

The satellite DMSP (Defense Meteorological Satellite Program) launched on 19 June 1987, is on a circular, sun-synchronous orbit with an altitude of 833 km and an inclination of 98.8° (Hollinger et al., 1987). It carries a microwave radiometer, SSM/I (Special Sensor Microwave Imager), measuring the brightness temperatures of the sea at 19.35; 22.2; 37.0 and 85.5 GHz, from which the wind speed, the water vapor and cloud liquid water content of the atmosphere can be deduced. We used the wind speed retrieved by F. Wentz from the 22 V; 37 V and 37 H GHz channels using the algorithm described in Wentz et al. (1986) and Wentz (1989). The swath of the instrument is 1400 km wide. The accuracy which is claimed for the retrieved wind speed is 2 m s^{-1} and the resolution is 25 km (Wentz, 1989). We used the data from 9 July 1987, to 30 June 1988. There is no measurement from 3 December 1987 to 12 January 1988. We will discuss further the accuracy of the wind speed retrieval in a subsequent part of this paper. Measurements affected by the proximity of land were removed by the masking method described in Ozieblo and Etcheto (1991) at 111 km from large land masses and 56 km from small islands. Data compromised by sea ice or rain were removed too, using the brightness temperatures from the radiometer. Wind speeds measured when the simultaneously determined cloud liquid water content was more than 25 mg cm^{-2}, were discarded.

3. Method

The measurements made during 1 month are binned in boxes $2.5° \times 2.5°$ in latitude and longitude. Then the CO_2 transfer velocity is computed for each measurement using the Liss and Merlivat relationship between the transfer velocity at 20°C, k in cm h^{-1}, and the wind speed at 10 m height, U in m s^{-1},

$$
\begin{aligned}
k &= 0.17U, && 0 \leqslant U \leqslant 3.6, \\
k &= 2.85U - 9.65, && 3.6 < U \leqslant 13, \\
k &= 5.9U - 49.3, && U > 13.
\end{aligned}
\tag{1}
$$

Satellite wind speed is retrieved at 19.5 m effective height and is converted to 10 m effective height assuming the atmosphere to be neutral and using $1.5 \cdot 10^{-3}$ for the drag coefficient.

Then the transfer velocity is averaged in each box resulting in a monthly map with a 2.5° resolution. Etcheto and Merlivat (1988) showed that the dependence of the exchange coefficient on the sea surface temperature can be neglected, thus resulting in the exchange coefficient K being proportional to the transfer velocity at 20°C, k:

$$
\underset{(\text{mol m}^{-2}\,\text{yr}^{-1}\,\mu\text{atm}^{-1})}{K} = \underset{(\text{cm h}^{-1}).}{2.89\;10^{-3}k}
\tag{2}
$$

Even though the Liss and Merlivat relationship gives the transfer velocity, we prefer to derive maps of the exchange coefficient, which is not temperature dependent.

Then we can compute the average transfer velocity over a given region of area A by weighting the transfer velocity in each $2.5° \times 2.5°$ square inside this region by its area and integrating over the region

$$
\langle k \rangle = \frac{1}{A} \iint_A k \, \mathrm{d}A.
\tag{3}
$$

The wind-speed measurements in the regions covered with sea ice were removed, so these areas of no gas exchange contain no k data and are not taken into account in the integration. The existence of polynyas was not considered, since their area does not exceed 10% of the area of the arctic ocean, which itself covers no more than 10% of the global ocean.

If we know the gradient of CO_2 partial pressure ΔP in each square of the region, we can compute the net flux exchanged between air and sea in that region:

$$
\phi = \iint_A k \, \Delta P \, \mathrm{d}A.
$$

4. Results

4.1. Exchange coefficient

The maps of exchange coefficient obtained by this method are shown in Fig. 1 for August 1987, November 1987, February 1988 and May 1988. The number of measurements per $2.5° \times 2.5°$ bin varies between 2000 and 4000 depending on the region and on the period. The seasonal variation for latitudes larger than 40° is clearly seen, with strong winds in autumn and winter in the northern hemisphere and a longer season of strong winds in the southern hemisphere, where the wind weakens in summer and to a lesser extent in autumn. The seasonal variation of the trade winds, stronger in the winter hemisphere, is also visible. One should notice that this is an anomalous year since an El Niño event was observed during the winter of 1986–1987. The monsoon, off the Horn of Africa, is nearly absent in August. At the same time, the trade winds are very weak in the Atlantic Ocean, while they are displaced to the west in the Pacific Ocean, suppressing the region of lows east of Indonesia. In February, the trade winds are still weak in the Atlantic and Indian Oceans.

We will now look more quantitatively at the seasonal variation of the exchange coefficient by plotting its value integrated in zonal bands per ocean. The limits of the bands are 80°S; 55°S; 40°S; 10°S; 10°N; 40°N; 80°N. The Antarctic Ocean has been defined as being south of 40°S. The results are shown in Fig. 2 for each ocean.

In the Atlantic Ocean (Fig. 2a), the main variation is north of 40°N, a factor of nearly 4 between July and January. The variation in the zone

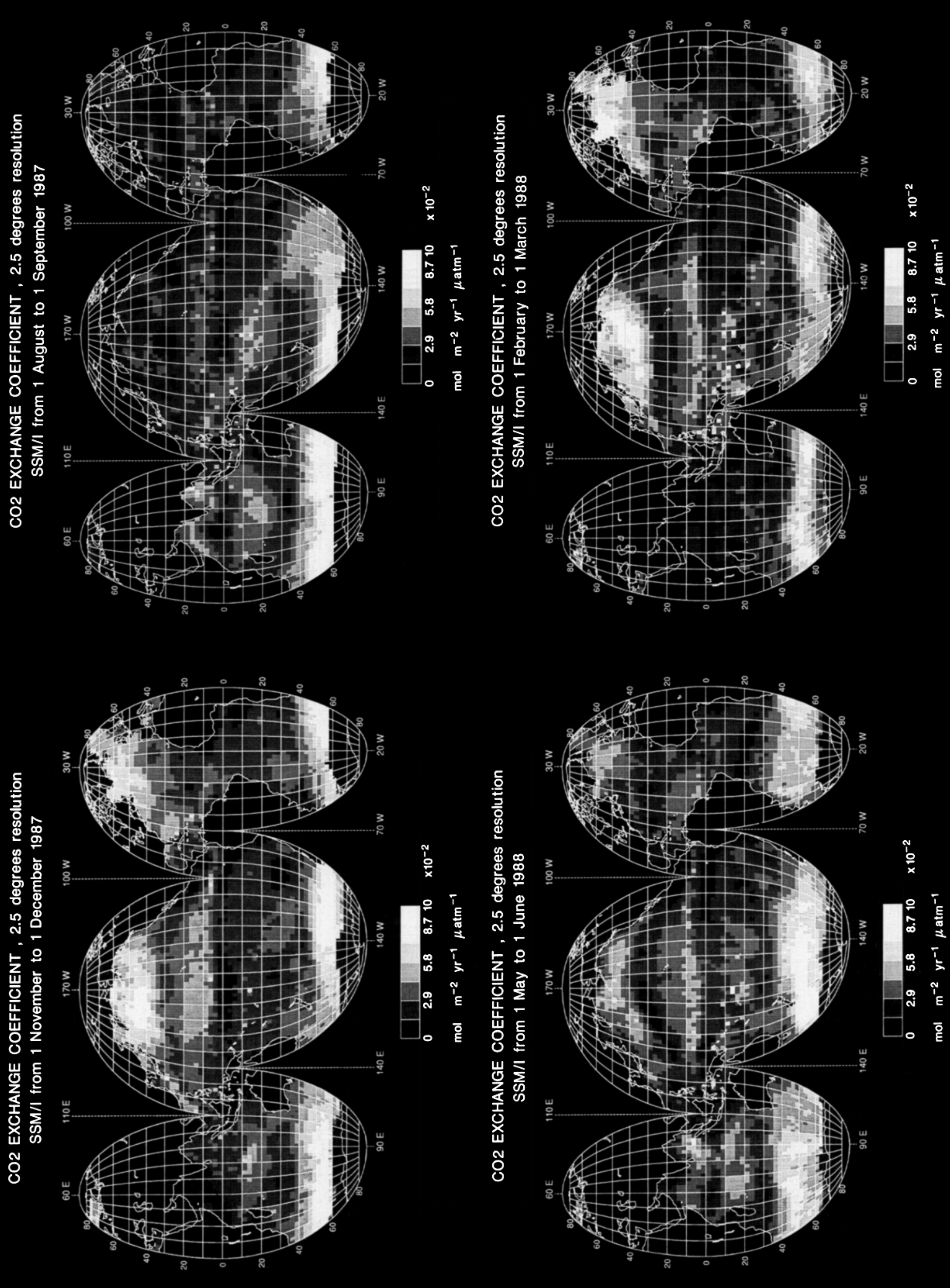

Fig. 1. Maps of the CO_2 exchange coefficient for August 1987, November 1987, February 1988 and May 1988.

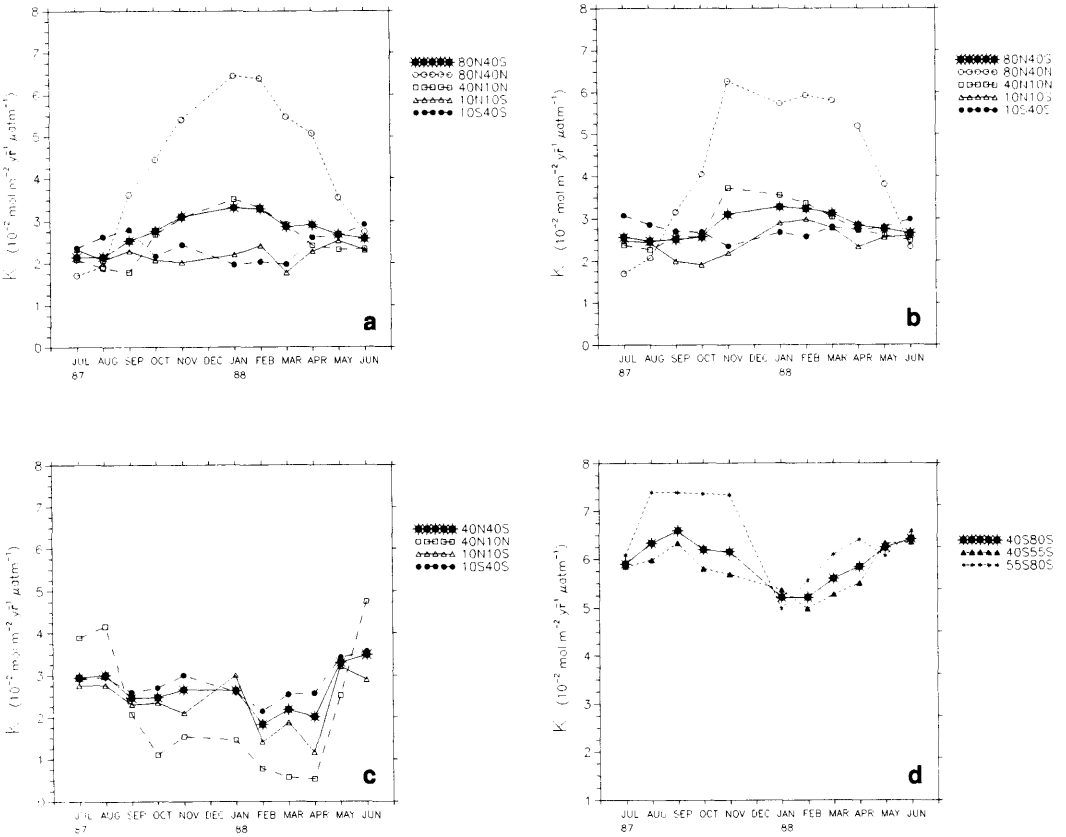

Fig. 2. Seasonal variation of the CO_2 exchange coefficient per latitudinal bands in each ocean: (a) Atlantic ocean; (b) Pacific ocean; (c) Indian ocean; (d) Antarctic ocean.

10°N–40°N is 2 while it is only about 1.5 at the corresponding south latitudes. In the equatorial band, the variation is small. Over the whole Atlantic Ocean (north of 40°S), the seasonal variation is a factor of 1.6, maximum in winter, with an annual average exchange coefficient of 2.77 10^{-2} mol m^{-2} yr^{-1} μatm^{-1} for an area of 68 10^6 km^2.

In the Pacific Ocean (Fig. 2b), the seasonal variation is similar to that in the Atlantic Ocean, but of slightly smaller amplitude, about 20% less. For the whole Pacific Ocean (north of 40°S) the seasonal variation is a factor of 1.3, compared to 1.6 in the Atlantic Ocean, with an annual mean of 2.83 10^{-2} mol m^{-2} yr^{-1} μatm^{-1} for an area of 144 10^6 km^2.

In the Indian Ocean (Fig. 2c), the seasonal variation is much more marked than in the two oceans above. This is due to the strong seasonality of the trade winds and monsoon regimes in this ocean. The 10°N–40°N band, which in fact involves a small region south of 25°N, shows a variation by a factor of 7 with a maximum between June and August and a minimum in April. The corresponding southern band varies only by a factor of 1.7, which is also less than the equatorial band which varies by a factor of 2.8. The whole Indian Ocean varies by a factor of 1.9 with a minimum between February and April, with an annual mean of 2.6 10^{-2} mol m^{-2} yr^{-1} μatm^{-1} over an area of 49 10^6 km^2.

In the Antarctic Ocean (Fig. 2d), the seasonal variation is weak: a factor of 1.6 south of 55°S with a minimum in January, and 1.3 between 40°S and

55° S, resulting in a factor of 1.3 for the whole ocean. The average annual exchange coefficient is $5.7 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}, twice as large as in any other ocean, for an area depending on the seasonal ice cover, of about $65 \cdot 10^6$ km^2.

Fig. 3a shows, in the same latitude bands, the seasonal variation of the exchange coefficient in the global ocean while Fig. 3b shows the corresponding areas. The largest seasonal variation, a factor 3.7, is north of 40° N, since it is of the same order in the Atlantic and Pacific Oceans. Elsewhere, the seasonal variations vary between 1.3 and 1.6, with opposite phase in both hemispheres, resulting in a global exchange coefficient, the variation of which is only 1.14 with an annual averaged value of $3.34 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}. In all regions, the seasonal variation tends to be larger than that found by Heimann and Monfray (1990). For instance, in the band north of 40° N, their seasonal variation is a factor of 2.5 compared to 3.7. This must be due to the fact that they use a climatology which tends to smooth the variations, while we study a particular year.

Despite its very high exchange coefficient, twice as large as for the other oceans, the Antarctic Ocean increases the exchange coefficient of the other three oceans by only 25%, because its area south of 55° S, where the exchange coefficient is large, is rather small, with a strong seasonal variation of 1.9 as can be seen in Fig. 3b, the area being minimal in winter when the exchange coefficient is the strongest. It can be seen from Fig. 3b that $\frac{2}{3}$ of the world ocean is between 40° S and 40° N where the exchange coefficient very seldom exceeds $6 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}. The sea ice coverage significantly changes the oceanic area only in the southern hemisphere, the seasonal change being about $13 \cdot 10^6$ km^2, while it changes by only $3 \cdot 10^6$ km^2 in the northern hemisphere, as compared to a global free ocean area of about $320 \cdot 10^6$ km^2.

4.2. CO_2 flux

The aim of computing monthly values of the exchange coefficient is to be able to compute CO_2 fluxes in which the temporal and spatial variability of K, as well as of the partial pressure difference, will be taken into account. Etcheto and Merlivat (1988, 1989) stressed the importance of taking into account the variability of both K and ΔP_{CO_2} and showed, using specific examples, that errors as large as a factor 8 can arise if it is omitted.

Unfortunately, at present, not enough is known about the partial pressure difference on a global scale. To get an order of magnitude of the CO_2 flux exchanged between ocean and atmosphere, we will combine two annually averaged maps of the CO_2 pressure gradient: the one of Broecker et al. (1986) derived from the available measurements and the one of Bacastow and Maier-Reimer (1990) derived from a three-dimensional ocean model, including the effect of the ocean biota (Fig. 14 of their paper) with the above maps of the exchange coefficient, averaged over 1 year. Since 1 month of data is missing (December), we first average by season and then average the seasons to obtain a yearly average. The resulting global net flux is $-0.89 \cdot 10^{15}$ g C yr^{-1} and $-1.39 \cdot 10^{15}$ g C yr^{-1},

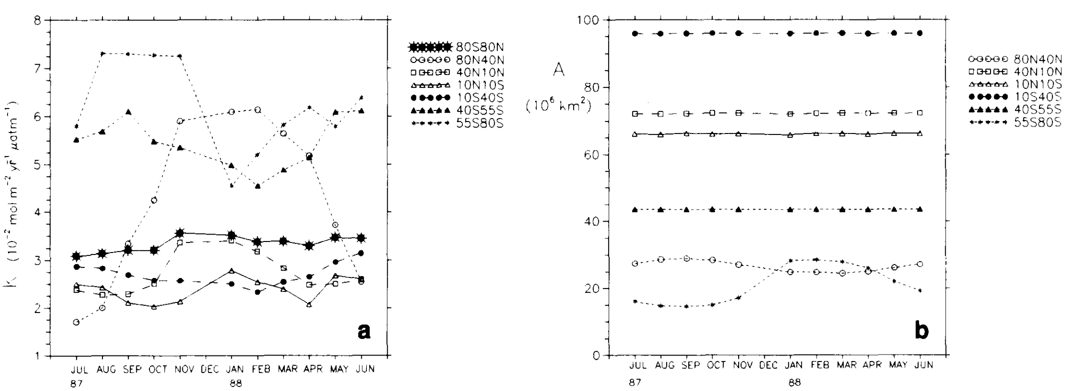

Fig. 3. Seasonal variation: (a) of the CO_2 exchange coefficient par latitudinal bands in the global ocean; (b) of the corresponding oceanic area.

respectively, a ratio of 1.6. Most of the absorption occurs in the Antarctic ocean while most of the outgassing occurs in the equatorial Pacific Ocean, for both CO_2 pressure gradient fields. In all regions, Broecker et al. (1986) give more outgassing and less absorption than Bacastow and Maier-Reimer (1990), except in the Antarctic Ocean and in the northern Atlantic Ocean.

These results can be compared to those obtained by Etcheto and Merlivat (1988), using the SEASAT scatterometer wind speeds and the Broecker et al. (1986) partial pressure gradient. The ratio of the average global exchange coefficient is 1.07, while the ratio of the global net flux is 1.35. This shows again the importance of the spatial and temporal variability of K: the SEASAT measurements were made during 3 months (July, August, September) of a normal year while the SSM/I measurements were made during an El Niño year, using a different instrument.

A recent estimate of the CO_2 flux at the atmosphere-ocean interface has been made (Tans et al., 1990). The authors show that the experimental data are still sparse at the ocean global scale. They report that the data set which is available today allows us to obtain an estimate of the mean value and of the seasonal variation of ΔP_{CO_2} for two periods of the year in the northern and equatorial areas. No such information exists for the southern Ocean, with respect to the mean annual value as well as to the seasonal variation of ΔP_{CO_2}. It is only when such information will be available that our exchange coefficient seasonal variation will be usable to derive a global net flux in a proper way.

5. Discussion and conclusion

Beyond the seasonal variation of the exchange coefficient, we should now assess more carefully the accuracy of the absolute value of this quantity. This accuracy depends on the validity of the Liss and Merlivat (1986) relationship and the reliability of the wind speed retrieval.

The Liss and Merlivat (1986) relationship was originally based on measurements made in Rockland lake by Wanninkhof et al. (1985) up to 8 m s^{-1}, extended to higher wind speeds by using the results of wind tunnel observations made by Broecker and Siems (1984) and Merlivat and Memery (1983). Recent additional measurements, including observations of wind speeds up to 22 m s^{-1} are in good agreement with this relation: measurements in Cornish lakes by Upstill-Goddard et al. (1990), measurements in the North Sea by Watson et al. (1991), and measurements in the very large Delft wind tunnel by Wanninkhof and Bliven (1991). Fig. 4 shows these new measurements superimposed on the original Liss and Merlivat relationship: the measurements are in good agreement with the curve. Several determinations of the global average exchange coefficient were performed using this relation. In this paper, we obtain a mean annual value of the exchange coefficient for the whole ocean for the year 1987–1988, equal to $3.34 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}. For the sake of comparison, we note that this figure was equal to $3.63 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1} for the 3 months of SEASAT data (July–September 1978) (Etcheto and Merlivat, 1988). In recent work, Heimann and Monfray (1990) compute a value of K based on a wind climatology equal to $3.56 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}. All these values differ by 10% at most. However, they are signifi-

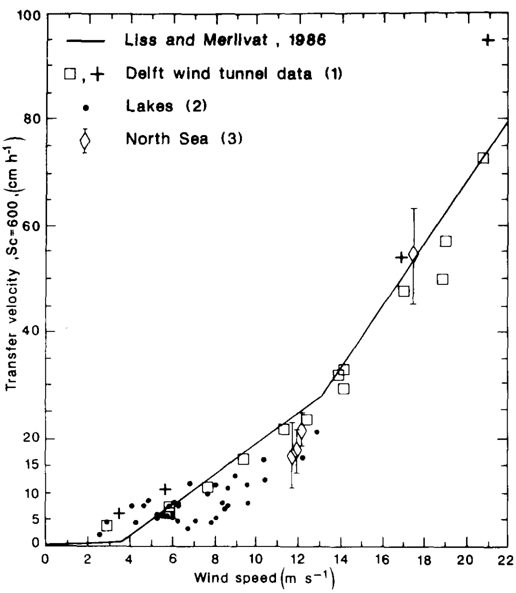

Fig. 4. CO_2 transfer velocity at 20°C versus wind speed at 10 m height above the surface. Recent measurements are superimposed over the Liss and Merlivat (1986) original relationship (solid line): (1) Wanninkhof and Bliven (1990); (2) Upstill-Goddard et al. (1990); (3) Watson et al. (1991).

cantly lower, by a factor of 1.6, than the carbon 14 based values of $6.1 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}, either natural or artificial ^{14}C (Broecker et al., 1986). The reason for this discrepancy is not clear. Possible explanations are proposed below.

In the case of the natural ^{14}C calculation, the fact that the natural ^{14}C distribution varies largely with latitude at the surface of the ocean (Toggweiler et al., 1989a, b) is not taken into account. However, the ensuing error is not expected to be more than 10%.

The calculation of the exchange coefficient based on the artificial carbon 14 oceanic inventories is not totally straightforward either, as within a column of water, the ^{14}C distribution is the result of a balance between the input function of $^{14}CO_2$ from the atmosphere and its redistribution in the interior of the ocean by the oceanic circulation. The accuracy with which the global ^{14}C inventory in the ocean is known, is difficult to evaluate. However, this knowledge is crucial to the understanding of the 1.6 factor between the ^{14}C based mean gas exchange coefficient and the values derived through the Liss and Merlivat relationship.

It is worth noting that in 3D models designed to compute the absorption of CO_2 by the oceans, the value of the global uptake is rather insensitive to the value of the gas exchange coefficient (Toggweiler et al., 1989a, b; Maier-Reimer and Hasselmann, 1988), the flux transfered between air and sea being mainly controlled by the transport inside the ocean. Thus, these models cannot be used to get information on the exchange coefficient.

We will now discuss the accuracy of the wind speeds retrieved from the brightness temperatures measured by the SSM/I, as Boutin and Etcheto (1991) showed that they are the dominant errors in using satellite wind speeds, well above sampling or integration errors. The algorithms were calibrated by careful comparison with buoys' measurements located close to the East and West Coast of North America and in the Gulf of Mexico. Moreover, a tuning of the global average wind speed with the results from the SEASAT scatterometer was performed, ensuring that the average of the wind speeds retrieved from the SEASAT microwave radiometer and scatterometer would be equal (Wentz et al., 1986). Yet, Boutin and Etcheto (1990) showed that strong regional biases could exist in the wind speeds retrieved from the two instruments. A true validation would require a global reference wind field, which is not available. We will nevertheless evaluate an upper limit for the error of the retrieved wind speeds.

The accuracy of the retrieved wind speeds is claimed to be 2 m s^{-1} (Wentz, 1989), and should result in a very small error on K after averaging in space and time, since it is random and should appear only via the non-linearity of relations (1). Of more serious consequences are possible biases.

One bias linked to sea surface temperature (SST) was identified by Boutin and Etcheto (1990), who proposed a correction. The SSM/I data we used were corrected for the SST bias, using another relation between wind speed and SST (Wentz, personnal communication). To get an order of magnitude for the error resulting from an incorrect correction of the SST bias, we compared the exchange coefficient obtained using each of the two correction laws during 4 days of SSM/I data in August 1987, assuming that the difference represents an estimate of the uncertainty. We found that the wind speeds used in this work give a global average transfer velocity underestimated by about 3 cm h^{-1} (corresponding to underestimating the exchange coefficient by $8.7 \cdot 10^{-3}$ mol m^{-2} yr^{-1} μatm^{-1}), with much larger values, in the regions of high wind speed and cold waters (about 4°C) south of 50°S, and still large deviations between 35°S and 50°S, where the SST is around 10°C. A possible influence of this kind of error on the seasonal variation of the exchange coefficient would derive from the seasonal variation of the SST and wind speed. The underestimation of the transfer velocity by the SSM/I is larger for lower SST (mainly below 13°C) and higher wind speeds (above 13 m s^{-1}). Thus, the tendency would be to decrease the seasonal variation at high latitudes. In order to quantify this error, we used an equation similar to eq. (3) to compute the averaged SST (analyzed in the way described in Reynolds, 1988) and wind speed, in each of the latitudinal bands used for the study in Section 4. From the averaged SST, we derived an estimate of the wind speed error (assumed to be equal to the differece between the two correction laws at this temperature) and from the averaged wind speed and eqs. (1) and (2), we converted it into an error in the exchange coefficient. We did that for August 1987 and February 1988. The difference between the 2 errors is considered as an estimate on the seasonal variation error. In all cases for which there is an effect, it is an underestimation of the seasonal cycle. Outside the equatorial band, the error is of the order of $7 \cdot 10^{-3}$ mol m^{-2} yr^{-1} μatm^{-1}, probably less between 10° N and 40° N. The error is larger between 40° N and 80° N, about 2 × the value found in the southern ocean. The error on the seasonal cycle of the global average exchange coefficient should be small, since the errors in both hemisphere roughly compensate. Since none of these two laws for correcting wind speed dependence on SST is sure, these estimates should be taken only as an indication of the order of magnitude of the uncertainty, not reliable enough to use to correct our results.

To derive an error estimate of the absolute value of the exchange coefficient, it is necessary to check satellite wind speeds against in situ data. First, we compared the SSM/I wind speeds with ship data. We used ship data from the National Climatic Data Center (NCDC) on the tropical Atlantic Ocean in the region 20° S to 30° N and 15° W to 60° W, for years 1987 and 1988.

The comparison was made for 10 months of data between August 1987 and June 1988 (in October and December 1987, the SSM/I data were not available at that time). We first binned, for each month, the ship data in 2.5° × 2.5° squares after removing wind speeds larger that 30 m s^{-1} and kept only the squares containing at least 30 measurements, after verifying that they are evenly distributed in time. It results in about 70 2.5° × 2.5° squares per month, this number varying between 45 and 110, depending on the month. Then the SSM/I measurements made in the same square during the same month are averaged. Then, the two averages are compared, plotting the SSM/I wind speed against the ship wind speed. We made a scatter plot of all the data obtained during 10 months (Fig. 5), obtaining 789 couples of averaged wind speeds. A regression line fitted to these points, using a principal component analysis, explains 89% of the variance and gives the following fit:

$$U_{\mathrm{SSMI}} = 0.89\ U_{\mathrm{NCDC}} + 0.08. \tag{4}$$

The average distance of the points to the fit is 0.7 m s^{-1}, giving 95% confidence limits of 1.4 m s^{-1}. The average difference between the ship wind speeds and the SSM/I wind speeds is 0.66 m s^{-1}, the SSM/I being low. When converted into an exchange coefficient, it results in the SSM/I underestimating the exchange coefficient by $5 \cdot 10^{-3}$ mol m^{-2} yr^{-1} μatm^{-1} (1.8 cm h^{-1} for the transfer velocity).

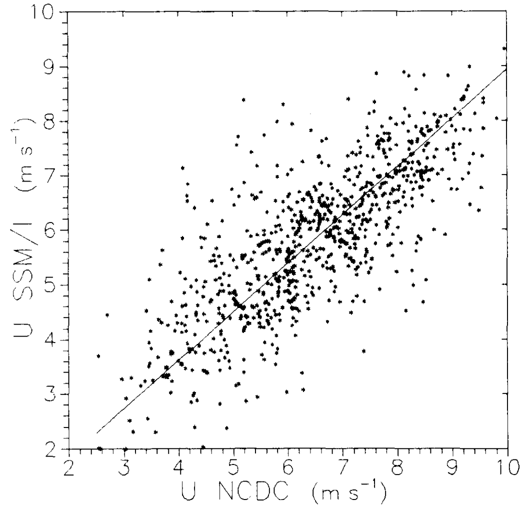

Fig. 5. Monthly average of SSM/I wind speed measurements in 2.5° × 2.5° squares against the corresponding ships measurements, for 10 months during 1987–1988.

If we assume that this is representative of an error on a global scale, we find a global exchange coefficient of about $4 \cdot 10^{-2}$ mol m^{-2} yr^{-1} μatm^{-1}, slightly above the values previously obtained from wind speed fields of other origins, during other periods.

In conclusion, using satellite wind speed measurements, we have been able to determine the seasonal variation (for a particular year) of the CO_2 exchange coefficient everywhere on the global ocean. This variation is large and should be taken into account when computing the net flux exchanged between ocean and atmosphere. It is vital to point out that the space and time variability of ΔP can be very large too. The next step is now to study the interannual variation of both the average exchange coefficient and the seasonal variation, since the year studied in this paper was an anomalous one, and to improve the spatial and temporal observational record for Δp.

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