Recent increase of DIC in the western North Pacific

Tsuneo Ono a,*, Yutaka W. Watanabe b, Shuichi Watanabe c

a National Research Institute of Fisheries Science, 12-4 Hakuura, Kanazawa-ku, Yokohama, 236-8648, Japan
b National Research Institute for Resources and Environment, 16-3 Onogawa, Tsukuba, Ibaraki 305-8569, Japan
c Graduate School of Environmental Earth Sciences, Hokkaido University, Sapporo 060-0810, Japan

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Abstract

The temporal variation of the total dissolved inorganic carbon (DIC) content in the western North Pacific is investigated by comparing the DIC distribution obtained from the data sets of three different periods, the GEOSECS data observed in 1973, the CO2 dynamics Cruise data observed in 1982, and recent Japanese data sets observed during the early 1990s. The overall feature of the signal of temporal DIC change during 1973 and early 1990s agreed with that of former studies, and did not significantly change with the calculation scheme (the grid-selection method vs. the multiple regression method). The observed increase in DIC among the different time scales showed a good inner consistency, which also indicates the stability of the method used in the DIC change calculation. The apparent rate of increase of the DIC inventory in the upper 1000 m water column, however, differed significantly by the data set used for the calculation: It was 5.6 ± 2.4 g C/m2/year, based on the data comparison between 1982 and the early 1990s, while it became 7.6 ± 2.4 g C/m2/year when based on the data between 1973 and the early 1990s. This result provides us an information about the data-dependency on the former estimation of temporal DIC change. © 2000 Elsevier Science B.V. All rights reserved.

Keywords: North Pacific; Total dissolved inorganic carbon; Carbon cycles; Anthropogenic effects

1. Introduction

The most fundamental and reliable way to detect the distribution and transport of anthropogenic carbon in the ocean is to carry out precise monitoring studies on the long-term temporal changes in the global distribution of total dissolved inorganic carbon (DIC) with a suitable spatial/temporal resolution, as for atmospheric CO2 and/or O2 monitoring. Data on the temporal change of sea surface fugacity of CO2 in several ocean basins have recently been published (e.g., Wong and Chan, 1991; Wong et al., 1995; Inoue et al., 1995; Poisson et al., 1996; Takahashi et al., 1997; Nojiri et al., 1999). However, we do not yet have enough information on the temporal changes in the distribution of DIC and its related properties (titration alkalinity; TA, and pH) within the water column, except at a few point locations such as HOT (e.g., Winn et al., 1994, 1998), BATS (e.g., Bates et al., 1996), Station P (e.g., Wong and
2. Data and analysis

Data from four Japanese cruises conducted during 1992–1994 were analyzed in this latest data set, and hereafter are referred to as the 90s’ data (detailed cruise information is listed in Table 1). Data from the CO$_2$ Dynamics WNP Cruise (hereafter WNP) were obtained from the FTP site at CDIAC (Chen et al., 1988). Data of dissolved oxygen and nutrients in WNP were also obtained from NOAA PMEL (PMEL/GCCP, 1999). The following nine GEOSECS Stations located around the 90s’ Cruises stations were used for analysis (Broecker et al., 1982) after making the correction recommended in this reference: Stn. 214 (32.02°N, 176.98°W), Stn. 215 (37.48°N, 177.32°W), Stn. 217 (44.67°N, 177.05°W), Stn. 218 (50.43°N, 176.58°W), Stn. 221 (45.21°N, 149.42°E), Stn. 222 (40.17°N, 160.50°E), Stn. 223 (34.97°N, 151.00°E), Stn. 225 (32.62°N, 161.92°E), Stn. 226 (30.57°N, 170.60°E).

All data were interpolated linearly to the fixed 28 density surfaces from $\sigma_t = 24.80$ to 27.75. These density surfaces were chosen at 0.05 $\sigma_t$ intervals in the density range of 27.10 < $\sigma_t$ < 27.75, at 0.10 $\sigma_t$ intervals in the range of 26.60 < $\sigma_t$ < 27.10, and at 0.20 $\sigma_t$ intervals in the remaining range. Data above 100 m depth were deleted from the data sets to avoid any seasonal differences between the data sets. Some GEOSECS data were also deleted before the density-based data sets were made if they were obviously outlined in the vertical profile. The location of stations in each data set is shown in Fig. 1.

The grid-selection method (Ono et al., 1998) was used to calculate the temporal DIC change between the data sets. This method determines which data pair between the two data sets is the most suitable to

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Table 1

<table>
<thead>
<tr>
<th>Cruise name</th>
<th>Sampling date</th>
<th>Sampling area</th>
<th>No. of stations</th>
<th>Depth</th>
<th>Reference</th>
</tr>
</thead>
</table>

*Thirty-six stations east of 170°W were used in this study.
calculate DIC differences, avoiding any artificial changes caused by geographical factors such as the difference of data-point location between the data sets or interannual drifting of water masses during observations. In this method, analyses are made on the density surface basis, and in the following discussion we use the term “grid” as the 3D-coordinate of each data (latitude, longitude, density).

For each data of a base data set (hereafter Grid A), all the grids of a different data set which were within the range of ±4° latitude and ±10° longitude and had the same density as the Grid A, were selected as the first candidates for DIC comparison (Grids B; see Fig. 2A). The salinity of each grid in the Grids B was then compared with that of the Grid A. The grids which had the same salinity as the Grid A, within a criterion range (Table 2), were selected from the Grids B and formed a new set, Grids C (see Fig. 2B).

In the next step, the apparent oxygen utilization value (AOU) of each grid in the Grids C was compared to that of the Grid A. The grids which had the same AOU with the Grid A, within a range of ±5 \( \mu \text{mol/kg} \), were selected from the Grids C and formed a final data set (Grids D, see Fig. 2C). If all grids in the Grids C had an AOU difference larger than 5 \( \mu \text{mol/kg} \) relative to Grid A, the three members of the Grids C which had the smallest AOU difference within the range of ±15 \( \mu \text{mol/kg} \) were selected as the Grids D. When all grids in the Grids C had an AOU difference larger than 15 \( \mu \text{mol/kg} \) relative to Grid A, the analysis for this grid A was canceled.

The temporal DIC differences between Grid A and each grid of Grids D were finally calculated using the following equation:

\[
\Delta \text{DIC} = \frac{(\text{DIC}_B - \text{DIC}_A) S_A}{35} - 0.768(\text{AOU}_B - \text{AOU}_A) \tag{1}
\]

\[
\Delta \text{DIC} = \text{average}(\Delta \text{DIC}) \tag{2}
\]

where \( \Delta \text{DIC} \) is the DIC difference between the Grid A and each grid of the Grids D, the \( \Delta \text{DIC} \) is the
The Grid Selection Method

(A) establishment of Grids B for a Grids A

(B) selection of Grids C from Grids B

(C) selection of Grids D from Grids C

Fig. 2. A schematic of the grid-selection method. (A) The Grids B are selected based on the relative location against the Grid A. (B) The Grids C are selected from the Grids B based on the salinity difference. In this figure, the solid circles represent the grids that pass the selection criteria, and the open circles represent those that do not pass the criteria. (C) The Grids D are finally selected from the Grids C based on the AOU difference.

The estimated temporal difference in DIC at the location of the Grid A and is defined by the average of $\delta$DIC. AOU$_A$ and NDIC$_A$ are the AOU and the DIC value of the Grid A normalized to a salinity of 35, respec-
These criteria are the same as those of Ono et al. 1998.

Table 2
Selection criteria for the extent of discrepancy in salinity between the corresponding data sets in the grid-selection method

<table>
<thead>
<tr>
<th>Density ($\sigma_r$)</th>
<th>Limit salinity difference</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\sigma_r \leq 25.40$</td>
<td>0.030</td>
</tr>
<tr>
<td>25.40 &lt; $\sigma_r$ ≤ 26.60</td>
<td>0.020</td>
</tr>
<tr>
<td>26.60 &lt; $\sigma_r$ ≤ 27.20</td>
<td>0.010</td>
</tr>
<tr>
<td>27.20 &lt; $\sigma_r$</td>
<td>0.005</td>
</tr>
</tbody>
</table>

These criteria are the same as those of Ono et al. (1998).

3. Results and discussions

3.1. DIC increase during 1973–1993 in the temperate/subpolar North Pacific

Vertical profiles of $\Delta$DIC between the 90s’ data and GEOSECS (hereafter as $\Delta_{93-73}$) was calculated at each station. Although there seemed to be some systematic difference in the $\Delta_{93-73}$ profiles among the stations, such station-to-station variability was rather small so that we were unable to distinguish whether this had a significant meaning. In the following discussion, we thus concentrate on the area averaged profiles of $\Delta_{93-73}$ for the stations south of 40°N (subtropical average) and north of 40°N (subpolar average), which are shown in Fig. 3a and b.

In the averaged profiles, the standard deviation of $\Delta_{93-73}$ in the waters below the density surface of $\sigma_r = 27.50$ (i.e., the lowest five points of each profile in Fig. 3a) was 3.8 µmol/kg for the subtropical average and 5.7 µmol/kg in the subpolar average. If we assume that these density surface are deep enough so that the $\Delta_{93-73}$ values in these waters were constantly zero, the uncertainty of our $\Delta_{93-73}$ calculation in the deep layers was estimated from these variability as about 5 µmol/kg.

There are several potential sources of systematic errors in the $\Delta$DIC calculation. The selection criteria of salinity are decided by rather practical reasons; i.e., they are decided so that at least five grids remain in Grids C against any density surface of a GEOSECS station. Similarly, the selection criteria of AOU are decided so that at least three grids remain in Grids D on 75% of the density surfaces in a GEOSECS station. Thus, sometimes the water masses may not be the same between Grid A and Grids D even after the grid selection is done. In this case, artificial DIC differences between the water masses may occur for the following two reasons: (1) difference in the pre-formed DIC value and (2) difference in the initial DO saturation. In addition, we should consider one more source of error when AOU differences are large, and that is: (3) difference in the biological stoichiometry from the value used in Eq. (1), i.e., 0.768.

The pre-formed DIC value typically changes by 5 µmol/kg per 1°C potential temperature change (e.g., Takahashi et al., 1999). On the other hand, the salinity difference of 0.03 psu corresponds to the potential temperature difference of 0.3–0.4°C on the $\sigma_r = 24.80$ density surface (i.e., the smallest density in Fig. 3a). Thus, the maximum artificial difference in DIC due to the pre-formed value difference is estimated to be about 2 µmol/kg. Since the sensitivity of water density to temperature changes becomes
larger at lower temperature, the pre-formed DIC difference due to the salinity difference become smaller in the deeper layers.

The difference of initial DO saturation value among the water masses is a bit difficult to estimate. If this value differs between the water masses, an artificial difference in the AOU value occurs and consequently causes the artificial DIC difference while calculating Eq. (1). Najjar and Keeling (1997) reported the climatological distribution of oxygen anomaly against its saturation value in the world ocean. Although their distribution is somewhat different from that of the sea surface temperature (e.g., Levitus and Boyer, 1994) in winter, the area variability of oxygen anomaly seems not to exceed 5 μmol/kg within the area with a 1°C of temperature variation. This implies that the artificial AOU difference between the water masses does not exceed 2 μmol/kg even if there is 0.03 psu of salinity difference (i.e., 0.3–0.4°C of the potential temperature difference) between the grids. Thus, the artificial DIC difference by such artificial AOU difference does not exceed 1.5 μmol/kg.

Finally, we must consider the uncertainty of the AOU coefficient in Eq. (1). There are several estimation of the coefficient of DIC change against AOU change during the remineralization in the open ocean (e.g., Redfield et al., 1963; Takahashi et al., 1985; Anderson and Sarmiento, 1994). Among these, Anderson and Sarmiento (1994) reported the lowest coefficient value of 0.688. On the other hand, Takahashi et al. (1985) reported the highest coefficient value of 0.814. If we use these different coefficient values instead of 0.768 in Eq. (1), the ΔDIC can differ by 1.2 μmol/kg from the original calculation results when the AOU value differs by 15 μmol/kg between the two grids.

To sum up the above three possibilities, the potential error for ΔDIC values associated with the grid-selection method itself is estimated to 5 μmol/kg at maximum. This is the same value as that of the scatter in the data observed in Fig. 3a and b.
The apparent penetration depth of $\Delta u_{93-73}$, the maximum depth where $\Delta u_{93-73}$ exceeds the theoretical uncertainty of 5 $\mu$mol/kg, was 770 m (or $\sigma_u = 26.96$ in density) for the subtropical average and 910 m (or $\sigma_u = 27.60$) for the subpolar average, respectively (Fig. 3a and b). The 5 $\mu$mol/kg isoline of $\Delta u_{93-73}$ occurred around 650 m in the eastern temperate North Pacific (Slansky et al., 1997). Ono et al. (1998) also found that the apparent penetration depth of $\Delta u_{93-73}$ was around the $\sigma_u = 27.00$ density surface, which corresponded to a depth of around 700 and 800 m in the eastern and western temperate North Pacific, respectively. Our results on the detection limit of $\Delta u_{93-73}$ in the temperate North Pacific thus agree closely with the former analysis. We should note, however, that the standard deviation of $\Delta u_{93-73}$ is rather large relative to its absolute value and thus the small signal of DIC increase in the deep layers may be blurred by such noise. CFCs in the western temperate North Pacific were found to penetrate to deeper than 1000 m at the time of the 90s’ Cruises (Watanabe et al., 1994; Tokieda et al., 1996; Warner et al., 1996). The ‘true’ penetration depth of $\Delta u_{93-73}$ should be deeper than that of the CFCs, as the intrusion of CFCs into the ocean started far later than that of anthropogenic carbon. (Note that the $\Delta u_{93-73}$ signal does not imply that anthropogenic carbon penetrated after 1973. If the age of a water mass was 1900 during the GEOSECS cruises, a signal of DIC increase corresponding to the period from 1900 to 1920 would be observed in the $\Delta u_{93-73}$.)

We should also consider that if there is some ‘true’ $\Delta u_{93-73}$ signal even in the waters below the $\sigma_u = 27.50$ density surface, the calculated $\Delta u_{93-73}$ value should be smaller than the true one because we have corrected the systematic difference of DIC values between the data set based on the data of these deep waters. If we correct the DIC difference based on the data below the $\sigma_u = 27.60$ density surface only, however, the calculated $\Delta$DIC values decreased by 0.6 $\mu$mol/kg. This suggests to us that the $\Delta u_{93-73}$ signal below the $\sigma_u = 27.50$ density surface is smaller than the analytical errors. Thus, even if the signal existed, we can ignore its effect on the uncertainty of $\Delta$DIC values.

The $\Delta u_{93-73}$ detection limit in the subpolar region seemed deeper than that observed in the subtropical region, although the scatter of the profiles made the information blurred (Fig. 3a and b). This is the first report of detailed $\Delta$DIC profile observed in the subpolar Pacific, and more data are required to evaluate further the variation in the distribution of anthropogenic tracers in this region. Above the detection limit, the subpolar $\Delta u_{93-73}$ values were consistently larger than the subtropical one for each density surface from $\sigma_u = 27.3$ to $\sigma_u = 26.9$ (Fig. 3a). This made the depth-based vertical profile of subpolar $\Delta u_{93-73}$ very similar to the temperate one (Fig. 3b), despite the fact that the averaged density structure of the water columns were significantly different between these areas.

The column inventory of $\Delta u_{93-73}$ above the detection limit was calculated by integrating the $\Delta u_{93-73}$ vertical profile (Fig. 3b) and yielded an average of 152 ± 46 gC/m² for the subtropical and 167 ± 56 gC/m² for the subpolar, respectively. In this calculation, the $\Delta u_{93-73}$ above 100 m depth were assumed to be constant and the same as the average of the top three layers in each profile. The errors were estimated by multiplying the penetration depth to the estimated maximum potential uncertainty of $\Delta u_{93-73}$, i.e., 5 $\mu$mol/kg. The $\Delta u_{93-73}$ inventory observed in the temperate North Pacific in this study agreed well with former observations (Tsunogai et al., 1993; Ono et al., 1998). In addition, our study suggested that the DIC inventory in the subpolar North Pacific region had also been increased with an almost same rate as that in the temperate North pacific region.

The average value of the uppermost three layers (corresponding to the layers of 100–160 m for subtropical and 100–251 m for subpolar profile, respectively, Fig. 3b) was 18 ± 6 $\mu$mol/kg for subtropical profile and 13 ± 2 $\mu$mol/kg for subpolar profile. Assuming that the $\Delta u_{93-73}$ in the surface waters is thermodynamically proportional to the atmospheric pCO2 increase (from 330 to 357 ppm at Mauna Loa, Keeling and Whorf, 1997), the theoretically expected $\Delta u_{93-73}$ is around 14–17 $\mu$mol/kg for <40°N area and 11–14 $\mu$mol/kg at >40°N area, respectively, when using the carbonate dissociation constants of Dickson and Millero (1987) and Dickson (1990) for boric acid. The $\Delta u_{93-73}$ values observed in the near-surface layers agreed with these theoretical DIC increase within the observed uncertainty.

Wallace (1995) had established the multiple regression approach to calculate $\Delta$DIC, which virtu-
ally reconstruct the present DIC profile by using linear-regression equation of present DIC against several water properties (temperature, salinity, AOU, and nutrients). Since their method has been widely used in a recent ΔDIC calculation (e.g., Slansky et al., 1997 in the North Pacific), it is interesting to compare our results with that obtained by the multiple regression method. The multiple linear regression equations of present DIC distribution were generated from the 90s’ data according to the Slansky et al.’s (1997) formulation as follows:

\[
\text{DIC} = -7.250 T + 13.1 S + 0.587 \text{AOU} + 7.3 P \\
+ 0.588 S_i + 1668.4 \text{ (south of } 40^\circ \text{N)} \tag{3}
\]

\[
\text{DIC} = -15.60 T + 49.2 S + 0.787 \text{AOU} - 11.4 P \\
+ 0.013 S_i + 534.2 \text{ (north of } 40^\circ \text{N)} \tag{4}
\]

where \( T \), \( P \), and \( S_i \) are the potential temperature (°C), the phosphate concentration (\( \mu \text{mol/kg} \)), and the silicate concentration (\( \mu \text{mol/kg} \)), respectively.

The precision was ±3.9 \( \mu \text{mol/kg} \) for Eq. (3) and ±4.7 \( \mu \text{mol/kg} \) for Eq. (4), respectively.

The DIC profiles of the early 1990s at to each GEOSECS station were reconstructed using Eqs. (3) and (4), and \( \Delta_{95-73} \) was calculated by subtracting the observed DIC profile. The results averaged for the subtropical region (\(< 40^\circ \text{N}\)) and subpolar region (\(> 40^\circ \text{N}\)) are shown in Fig. 4a and b, respectively. Overall, patterns of \( \Delta_{95-73} \) distributions were very similar between the two methods, both in south of 40°N (Fig. 4a) and in north of 40°N (Fig. 4b). This result suggests that the dependence of ΔDIC on the calculation scheme is very small. This also indicates that the data scattering observed in the \( \Delta_{95-73} \) profiles (Figs. 3 and 4) is not caused by the potential uncertainties of each calculation scheme but due mainly by the scattering of the data itself.

In summary, our \( \Delta_{95-73} \) results showed good agreement with former studies both in the overall pattern of profile, penetration depth, near-surface values, and inventory. Fig. 4 also showed that the

![Fig. 4. Averaged profiles of \( \Delta_{95-73} \) for the GEOSECS stations (a) south of 40°N and (b) north of 40°N, respectively, plotted against the potential density. The open circles and x represent the results obtained by the grid-serection method and the multi-regression method, respectively.](image-url)
grid-selection method gives consistent results when compared with former procedures. In Section 3.2, we made an assessment about the data-set dependency of the \( \Delta \text{DIC} \); i.e., the distribution of \( \Delta \text{DIC} \) value between WNP and 90s’ Cruises in the North Pacific (hereafter \( \Delta_{93,82} \)) was analysed and compared with the \( \Delta_{93,73} \) results.

### 3.2. DIC increase during 1973–1982 and 1982–1993 in the temperate North Pacific

The \( \Delta_{93,82} \) profile was calculated for each WNP station just the same way as \( \Delta_{93,73} \) profile, and only the average profile for the area south of 40°N is shown in Fig. 5a and b.

The grid-selection method allows direct calculation of the \( \Delta \text{DIC} \) between GEOSECS and WNP (hereafter \( \Delta_{92,73} \)) to be made. This may be an advantage of our method, because the standard deviation of DIC predicted by the regression equation made from the WNP data set was ±16.1 \( \mu \text{mol/kg} \) and the \( \Delta_{92,73} \) could not be calculated at any station by the multi regression method. Even by our method, however, the coarse geographical resolution of both data sets restricted direct comparison only to two locations in the temperate region (around 30°N, 165°E and 35°N, 165°E) where stations were relatively concentrated in both data sets. The results of each \( \Delta \text{DIC} \) vertical profile around 30°N, 165°E are shown as an example in Fig. 6. Although there was significant scatter, the sum of \( \Delta_{92,73} \) and \( \Delta_{93,82} \) matched well to the \( \Delta_{93,73} \) on each density surfaces (Fig. 6). This inner consistency of the independently calculated three properties clearly indicated the stability of the output of our method, although we were unable to calculate the area-averaged \( \Delta_{92,73} \) profile directly because we could calculate it only at two locations.

The inner-consistency performed at the test stations encouraged us to calculate the average profile of \( \Delta_{92,73} \) indirectly, by subtracting the average \( \Delta_{93,82} \) value from the average \( \Delta_{93,73} \) for each density surface (results also shown in Fig. 5). On the other hand, we were unable to obtain any profile of \( \Delta_{92,73} \) directly for the area north of 40°N and we could not...
determine the reliability of the indirect Δ93.82 calculation in this area. Thus, in the following section, we concentrate our discussion on the results for the <40°N area only.

The Δ93.82 showed roughly constant values of around 2–8 μmol/kg on the density surfaces of σθ ≥ 27.00 (Fig. 5a) or above 700 m depth (Fig. 5b). The analytical precision of DIC in the WNP Cruise was about 5 μmol/kg (Chen et al., 1988), and the potential uncertainty related to the grid-selection method is also considered to be about 5 μmol/kg (see previous discussion). The standard deviation of the Δ93.82 values below σθ = 27.10, however, was far smaller than the predicted maximum uncertainty of our calculation, i.e., ±1.6 μmol/kg. The uncertainty of the Δ93.82 value, which is a statistically presumed value estimated as the average of the several δDICs (see Eq. (1)), thus seems far smaller than the simple analytical precision or theoretical uncertainty of each δDIC value (= 5 μmol/kg). As a consequence, the signal of positive Δ93.82 in the waters above σθ = 27.00 is also statistically significant.

During the 90s’ Cruises, the CFC age of the water mass on σθ = 26.00 and 27.00 density surfaces was determined as 1985 ± 4 and 1968 ± 3, respectively, in the temperate North Pacific (Watanabe et al., 1994). The time of formation of the same water masses are thus expected to have been about 1958 and about 1975, respectively, at the time of WNP Cruise. If we assume that the DIC increase in each water mass is thermodynamically proportional to the atmospheric pCO2 increase during the period of these CFC ages (9 ppm during 1958–1968 and 15 ppm during 1975–1985, Keeling and Whorf, 1997), the theoretical value of the average Δ93.82 becomes about 3–5 μmol/kg and about 6–8 μmol/kg for σθ = 26.00 and 27.00, respectively, depending on the potential temperature. The observed Δ93.82 values were distributed within the range of these theoretical values for these density surfaces. Watanabe et al. (2000) made more thorough calculation for the theoretical ΔDIC distribution during 1980s for each density surface from σθ = 27.8 to =25.0 over a wide area of the temperate North Pacific, and its average vertical profile showed a close agreement with that of our direct observations (Fig. 5a). This suggests to us that the observed distribution of Δ93.82 is reliable, and that the temporal DIC change during the 1980s observed in the intermediate layer (density range of 26.0 < σθ < 27.0) in the western North Pacific is well explained by the simple solubility model. These results also indicate that the Δ93.82 in these water masses is not significantly diluted after its ventilation from the mixed layer. Watanabe et al. (1995) found that the CFC content of the water above the density surface of σθ = 26.8 in the temperate North Pacific agreed well with its saturation value estimated from its CFC age, and concluded that these water masses are not significantly diluted by the ‘older’ waters in which no CFC are dissolved.

The Δ82.73 also distributed around the theoretical ΔDIC vertical profile below σθ = 26.4, but the scatter of data was much larger than that of Δ93.82 (Fig. 5a). In addition, the Δ82.73 values observed on the
Table 3
Column inventory of ΔDIC (upper column, g C/m²) and the rate of DIC increase (lower column, g C/m²/year) during each period

| Depth range (m) | Δ93-73 | Δ93-82 | Δ93-73
<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>300–1000</td>
<td>81 ± 4.2</td>
<td>31 ± 1.5</td>
<td>50 ± 5.9</td>
</tr>
<tr>
<td>0–1000</td>
<td>4.0 ± 2.0</td>
<td>3.4 ± 1.7</td>
<td>4.5 ± 5.3</td>
</tr>
<tr>
<td>152 ± 60</td>
<td>50 ± 2.2</td>
<td>102 ± 5.3</td>
<td>9.3 ± 7.6</td>
</tr>
<tr>
<td>7.6 ± 3.0</td>
<td>5.6 ± 2.4</td>
<td>9.3 ± 7.6</td>
<td></td>
</tr>
</tbody>
</table>

The σb = 26.00 density surface occurred around 300 m depth both in 1973 and 1982 (Fig. 5b). We could not calculate Δ93-82 above this density surface because all 90’s data were refused to be in Grids D of any WNP station as the results of the grid selection. The Δ93-73 and Δ93-82 inventories above 300 m depth was thus calculated assuming that these waters had the same ΔDIC value as that of the σb = 26.00 density surface. The Δ93-73 inventories above 100 m depth were also calculated in similar way. The uncertainty of Δ93-82 and Δ93-73 inventories was estimated based on the overall uncertainty of observed ΔDIC concentration, which was indicated by the standard deviation of ΔDIC values below the σb = 27.10 density surface, i.e., ±5 μmol/kg for Δ93-73, ±2 μmol/kg for Δ93-82, and ±7 μmol/kg for Δ93-73.

4. Concluding remarks

In this paper, the temporal DIC increase during the periods from 1973 to the early 1990s was examined both in the subpolar and the temporal North Pacific. The column inventory of Δ93-73 above the detection limit was 152 ± 46 gC/m² for the subtropical region and 167 ± 56 g C/m² for the subpolar region, respectively. The calculation results of Δ93-73 and Δ93-82, however, suggested that the current ΔDIC estimations were strongly influenced by the choice of data sets. The 30% difference in the rate of increase of DIC derived from Δ93-73 and Δ93-82 provided information about the extent of such data-set dependency. Our result provides an information about the potential uncertainty in the recent estimations of DIC inventory change based on the various subtractive methods (Tsunogai et al., 1993; Wallace, 1995; Watanabe et al., 1996; Slansky et al., 1997; Ono et al., 1998).

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