Fate and Effects of Disposed CO₂ for Scenarios in the North Pacific Ocean

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Abstract. Carbon dioxide (CO₂) introduced into the oceans, either directly or indirectly via the atmosphere, reacts with the dissolved carbonate ions in sea water or with the calcareous particles in the water column as well as the chalk deposits on the ocean floor, to become dissolved bicarbonate ions. In the world oceans, the carbonate ions, which account for the buffering capacity of sea water, are limited to 1,440 GtC equivalent in the oceans but only about 250 GtC in the upper ocean, now being depleted at about 2 GtC yr⁻¹. CO₂ disposed into the upper 250 m of the oceans will accelerate this depletion, and is undesirable. Replenishment of the oceanic buffering capacity comes from river supply of carbonate and redissolution of the calcareous deposits, which are very low at less than 1 per cent of the sediment in the North Pacific Ocean. The environmental effects of direct CO₂ disposal into the oceans are change in pH, dissolution of calcareous shells and bones, and changes in the evasion rate of CO₂ from the oceans into the atmosphere. Based on a scenario of direct disposal of 0.25 GtC yr⁻¹ near Japan, and continuously over 200 years, the amount of excess CO₂ retained by the ocean is practically 100% for a disposal depth of 3,000–5,250 m, while only 50% for depths shallower than 463 m. The controlling factors of the fate of the excess CO₂ are oceanographic processes of horizontal and vertical diffusion, advection and large-scale circulation, and the neutralizing chemical processes of reaction with carbonate ions and diffusive supply of carbonate ions from chalk deposits on the sea floor. In contrast with the Atlantic, Indian and the South Pacific Oceans, with high carbonate contents in the sediments and availability of isolated small basins, the North Pacific Ocean, with a single large basin and lack of carbonate in the sediment, has to rely on dilution by its enormous volume to minimize the effects of the directly disposed CO₂. The criteria for selection of sites are: (1) depth greater than 3,000 m, (2) isolated basin, (3) large quantity of chalk deposit in the sediment, (4) residence time of the water mass and (5) acceptability of environmental sensitivity. The Philippines Basin, the Caroline Basin and the Guatemala Basin mainly in the tropics satisfy the first three criteria. The southern part of Okhotsk Sea and Bering Sea are also possible sites. The Canada Basin in the Arctic Ocean has the advantage of well stratified water column, sealed by freshwater or ice cap, and the extremely low temperature favors gas hydrate formation in the deep waters. However, in all these cases, both environmental sensitivity and economic factors have to be considered before any large-scale direct CO₂ disposal is attempted.
1. Introduction

Carbon is partitioned among major geospheric reservoirs in different forms. Free CO$_2$ molecules are found mainly in the atmosphere, now at 750 GtC (gigatons of carbon). The dissolved inorganic carbon in the oceans totals about 36,000 GtC, consisting of 250 GtC of dissolved free CO$_2$ and carbonic acid (H$_2$CO$_3$), 35,400 GtC of bicarbonate ions (HCO$_3^-$) and 1,500 GtC of carbonate ions (CO$_3^{2-}$). In addition, the oceans contain about 2,000–3,000 GtC of dissolved organic carbon. (Wong, 1978; De Barr, 1992; Wong and Matear, 1993). Large quantities of carbon are deposited as carbonate rocks (67 million GtC), as organic carbon in sediment (25 million GtC) in sediments on earth (Revelle and Fairbridge, 1957), of which the marine sediments hold 1,000 GtC as petroleum and 12 million GtC as organic carbon, but a major fraction of 50 million GtC as calcium carbonate (De Barr, 1992).

In the natural cycle, carbonate ions are entering the oceans by rivers at about 1 ± 0.5 GtC yr$^{-1}$ of which 13% is from silicate reaction and 87% by solution of terrestrial carbonate rocks (Revelle and Fairbridge, 1957). The ocean is already absorbing anthropogenic CO$_2$ indirectly via the atmosphere at a rate of 2.1 ± 0.7 GtC yr$^{-1}$ (Wong, 1978; Quay et al., 1992), i.e. about 30% of the total CO$_2$ emissions from the burning of fossil fuels and wood (IPCC, 1990). Direct ocean disposal of CO$_2$ collected from thermal power plants, as a means of mitigating global warming due to CO$_2$ was first proposed by Marchetti (1977), and disposal options were discussed in a recent IEA CO$_2$ Disposal Symposium, held in March, 1993 in Oxford, U.K. During the Symposium, the urgency to consider ocean disposal was spelt out. The Climate Campaign of Greenpeace International stated that the global warming concerns would place the acceptable total input into the atmosphere at about 300 GtC. The IPCC report (1990) has indicated that the cumulative input is already at 312 ± 40 GtC. If the world energy consumption of fossil fuels is proceeding at a “business as usual” pace of 2% yr$^{-1}$ increase as versus following the IPCC scenario D (business as usual to year 2010, then decreasing at 2% yr$^{-1}$, means has to be found to store the 1,200 GtC as the difference between the cumulative totals for the two scenarios, if the CO$_2$ is not released into the atmosphere. An overview by Ormerod et al. (1993) placed the total of the sinks of aquifers, exhausted gas and oil wells, enhanced oil recovery and global forest management, at 266–316 GtC, far short of the 1,200 GtC. Thus, the examination of the ocean disposal option (Ohsumi, 1993) has an urgency. There is uncertainty on the perception of the oceanic capacity to take up CO$_2$ in an environmentally acceptable way. Ormerod et al. (1993) suggested a capacity of 20 million GtC, based on the solubility of CO$_2$ in seawater, while Wong and Matear (1993) at 1,000 GtC based on neutralization of carbonate ions in the ocean. This paper will clarify this point and suggest criteria for sites in the North Pacific Ocean.
2. Thermodynamics of the Oceanic CO₂ System

For oceanic CO₂ disposal, the two most important parameters of the thermodynamics of the oceanic CO₂ system are the partial pressure of CO₂ ($p_{CO₂}$) and the carbonate ions ($CO₃^{2-}$). These quantities can be calculated from the equilibria of the chemical reactions of CO₂ in sea water:

\[
CO₂ (g) = \frac{K₀}{K₉} CO₂ (aq.)
\]

\[
CO₂ + H₂O = H₂CO₃ \quad \frac{K₉}{K₉'}
\]

\[
H₂CO₃ = H^+ + CO₃^-
\]

\[
HCO₃^- = H^+ + CO₃^{2-} \quad \frac{K₈'}{K₈'}
\]

where $K₀$ is the solubility coefficient of CO₂ in sea water, $K₉$ the hydration constant, and $K₉'$ and $K₈'$ the apparent first and second dissociation constants of carbonic acid. (UNESCO, 1987).

The partial pressure of CO₂ in sea water is defined by

\[
p_{CO₂} = \frac{[CO₂]}{K₀}
\]

where $[CO₂]$ is the concentration in sea water. In air-sea exchange of CO₂, the escape tendency of the gas from the sea surface into the atmosphere is governed by:

\[
F = K₉(p_{CO₂seawater} - p_{CO₂atmosphere}) = Δp_{CO₂}
\]

where $F$ is the flux of CO₂ from sea to air, $K₉$ is the gas exchange coefficient usually expressed as a function of wind speed, and $Δp_{CO₂}$ is the partial pressure difference between the sea surface and the atmosphere (Wong and Chan, 1991), and is the parameter determining whether a water body is a source or a sink of CO₂, important to the question of the effect of CO₂ disposed on the atmospheric CO₂ level.

The concentration of carbonate ions in sea water, can be calculated from the relationship (UNESCO, 1987):

\[
[CO₃^{2-}] = TA - \frac{T_B \cdot K'}{H^+ + K'} \cdot \frac{K₈'}{2 + H^+}
\]
where $T_A$ is the total alkalinity, $T_B$ the total boron, $K'$ the first dissociation constant of boric acid and $H^+$ the hydrogen ion concentration. The carbonate ion concentration in seawater is an important measure of the buffering capacity of the oceans. Carbonate ions neutralize the anthropogenic CO$_2$ either entering the oceans through air-sea exchange or directly through disposal, by the chemical reaction:

$$H_2O + CO_2 + CO_3^{2-} = 2HCO_3^-.$$

This reaction shows that the $p$CO$_2$ in sea water is inversely proportional to the carbonate ions. Thus, as the CO$_3^{2-}$ is being depleted by the neutralization, the $p$CO$_2$ goes up, thus increasing the escape tendency into the atmosphere, as illustrated in the diagram of Broecker and Peng (1982).

3. Neutralization of CO$_2$ by the Carbonate in Sediment

On the sea floor, CO$_2$ will be neutralized by the calcium carbonate deposits in the sediment, by the dissolution reaction:

$$CO_2 + H_2O + CaCO_3 \text{ (solid phase)} = Ca^{2+} + 2HCO_3^{-}.$$

The calcareous particles in the water column, such as that collected in sediment traps (Wong, 1989) in the Pacific Ocean, undergo the same dissolution reaction. The amount of CaCO$_3$ in deep sea sediment available for neutralization is estimated as:

$$\text{CaCO}_3 \text{ for neutralization} = \text{dry wt. density} \times \text{area} \times \text{burrowing mixed depth (9 cm)} \times \text{fraction of CaCO}_3/\text{fraction of non-CaCO}_3$$

by Broecker and Takahashi (1977), who derived at 4,920 GtC for the world ocean sediment (see Table 1).

Biogenic CO$_2$ in the North Pacific Ocean is high due to high productivity and has already neutralized the CaCO$_3$ in the ocean sediment. The Carbonate Compensation Depth (the equivalent of snow-line, below which little or no carbonate accumulates in the sediment) is sloping upward from 5 km. at the equator to about 3 km. in the Aleutian Island area (Seibold and Berger, 1982). The North Pacific sediment is poor in CaCO$_3$ with the majority of the area at less than 1% (Kennett, 1982). The North Pacific Ocean is rather flat and disposal of CO$_2$ into the deep waters is not expected to be neutralized by more than a fraction of the 108 GtC available for sediment depths between the lysocline and the compensation depth, i.e. probably close to 2 GtC (using about 1% for the North Pacific and 50% for the S. Pacific CaCO$_3$ contents in deep sediment), an insignificant amount in the neutralization process for disposal rate at the GtC yr$^{-1}$ range. In other words, the
Table 1. Available CO$_3$ in deep sea sediment in GtC equivalent (adapted from Broecker and Takahashi, 1977).

<table>
<thead>
<tr>
<th>Region</th>
<th>Atlantic</th>
<th>Indian</th>
<th>Pacific</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>Shelf break to Ridge crest (200-2,800m)</td>
<td>84</td>
<td>36</td>
<td>72</td>
<td>192</td>
</tr>
<tr>
<td>Ridge to Lysocline</td>
<td>2,160</td>
<td>1,104</td>
<td>1,080</td>
<td>4,368</td>
</tr>
<tr>
<td>Lysocline to Compensation depth *</td>
<td>156</td>
<td>96</td>
<td>108</td>
<td>360</td>
</tr>
<tr>
<td>Total</td>
<td>2,400</td>
<td>1,236</td>
<td>1,285</td>
<td>4,920</td>
</tr>
</tbody>
</table>

*Note: The lysocline and compensation depths vary widely in different oceans e.g. from 4.5 to 6 km in the Atlantic, 4 to 5 km in the Indian, and 3.5 to 5 km in the Pacific Ocean. The lysocline varies from 4.7 km in the N. Atlantic to less than 3 km in the N. Pacific.

direct CO$_2$ disposal in the North Pacific Ocean would not be buffered by the sediment carbonate unless the sites are selected near the tropical Pacific areas where more carbonate (1–30% range) is available.

4. Neutralization of CO$_2$ by the Carbonate Ions in the Ocean

The North Pacific Ocean has high content of total CO$_2$ due to high production rates in the upper ocean, with oxidation of organic matter at the oxygen minimum to produce biogenic CO$_2$. This biogenic CO$_2$ produces high pCO$_2$ in the intermediate waters in the N. Pacific at depth where remineralization occurs. Figure 1 shows the pCO$_2$ distribution in the Pacific Ocean on the HUDSON-70 from 63°S to 57°N along 150°W, with the pCO$_2$ calculated from total alkalinity and total CO$_2$. Figure 2 shows the carbonate ions for the same section. The carbonate ions, based on data of GEOSECS, INDOPAC, HUDSON-70, TRANSPAC-72, HUDSON-81 and GEMS-91, are summarized in Table 2. In the upper ocean, the waters are supersaturated with respect to calcite and aragonite. The calcite saturation in the Pacific Ocean is shown in Fig. 3. The upper waters of the tropical Pacific Ocean are highly supersaturated with calcite, to over 400% in 30°N to 30°S, and the degree of saturation decreases with latitude towards the north, where the deep undersaturated waters upwell. In the Southern Hemisphere, the 100% saturation line stays at about 1000 m depth, shallowed gradually to less than 100 m at 57°N. The deeper waters in the North Pacific are undersaturated with respect to both calcite and aragonite. This undersaturation is caused by high
Fig. 1. Distribution of \( \rho \)CO\(_2\) in the Pacific Ocean from 63°S to 57°N along 150°W during the Canadian HUDSON-70 Expedition (\( \rho \)CO\(_2\) in atm. units).

\[ \text{CO}_3^{2-} \text{ (m mool kg}^{-1} \text{)} \]

LATITUDE

CO\(_3^{2-}\) (m mool kg\(^{-1}\))

S

N

DEPTH (m)

0

1000

2000

3000

4000

5000

60

50

40

30

20

10

0

0.08

0.10

0.20

0.25

0.30

< 0.10
Fig. 2. Distribution of $\text{CO}_3^{2-}$ in the Pacific Ocean from 63°S to 57°N along 150°W during the Canadian HUDSON-70 Expedition ($\text{CO}_3^{2-}$ in $\mu$M kg$^{-1}$).
Fig. 3. Calcite saturation in the Pacific Ocean from 63°S to 57°N along 150°W during the Canadian HUDSON-70 Expedition (in % saturation).
Table 2. Carbonate ions concentration (CO$_3^{2-}$) in the Pacific Ocean available for neutralization of CO$_2$ (in GtC).

<table>
<thead>
<tr>
<th>LATITUDINAL ZONE</th>
<th>DEPTH INTERVALS</th>
<th>TOTAL</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0-250m</td>
<td>250-1000m</td>
</tr>
<tr>
<td>60°N-40°N</td>
<td>4.0</td>
<td>5.9</td>
</tr>
<tr>
<td>40°N-15°N</td>
<td>18.7</td>
<td>23.1</td>
</tr>
<tr>
<td>15°N-0°N</td>
<td>17.0</td>
<td>17.9</td>
</tr>
<tr>
<td>0°S-10°S</td>
<td>18.3</td>
<td>20.2</td>
</tr>
<tr>
<td>10°S-40°S</td>
<td>24.1</td>
<td>38.5</td>
</tr>
<tr>
<td>40°S-60°S</td>
<td>8.5</td>
<td>19.3</td>
</tr>
</tbody>
</table>

| 60°N-0°N         | 364.6            | |
| 60°S-0°S         | 462.7            | |
| 60°N-60°S        | 827.3            | |

biogenic CO$_2$ generated by detritus flux of organic carbon raining into the deeper waters. Directly disposed CO$_2$ into the deeper water of the North Pacific Ocean has less carbonate ions to interact with, again less buffered when upwelled back into the upper ocean, thus has smaller tendency to remain in the ocean.

This neutralizing capacity of the ocean is finite. Atmospheric CO$_2$ from the burning of fossil fuels and wood is entering the upper ocean through air-sea exchange at a rate of 2.1 GtC yr$^{-1}$ (Quay et al., 1992). This will interact with the buffering zone of carbonate ions at the same rate to convert the CO$_3^{2-}$ to HCO$_3^-$ using Broecker and Peng's (1982) estimate of 1,440 GtC of CO for the world ocean, the upper 250 m of water would provide about 254 GtC in 1970s. For the North Pacific Ocean, the atmospheric CO$_2$ is being absorbed into the upper ocean at about 0.3 GtC yr$^{-1}$ (Wong, unpublished data) based on $p$CO$_2$ measurements on ships of opportunity. Thus, in the North Pacific Ocean, the carbonate ions at about 40 GtC from Table 2, will be neutralized at about 1% per year. Directly disposed CO$_2$ at 0.25 GtC yr$^{-1}$ into the surface North Pacific Ocean will consume the buffering capacity at almost doubling the present rate of uptake in the North Pacific, at almost 2% per year.

5. Circulation of the North Pacific Ocean

Carbon dioxide discharged directly into the surface mixed layer will equilibrate very fast with the atmosphere within about one year. Thus, to achieve any delay in returning to the atmosphere or to make use of the very large buffering
capacity of the intermediate and deep ocean, discharge at greater depths is necessary. The return of the deeper water and its CO$_2$ contents would be controlled by the oceanographic processes of advective and vertical diffusive transports, as discussed in Wong and Matear (1993). In the North Pacific Ocean, Ekman vertical transport (Gill, 1982) is produced by cyclonic wind stress on the air-sea boundary layer, causing the surface water to move away from the cyclonic gyre, with deeper water being sucked up into the surface mixed layer as replacement. The Ekman vertical transport varies according to the zonal wind stress. In the subarctic North Pacific waters, such a transport is rather weak (Gargett, 1991) at about 15 m yr$^{-1}$ (Miller et al., 1991). Acting in the opposite direction, is the barrier created by the “estuarine” circulation of highly stratified water masses with an outflow of freshwater from coastal drainage in the summer from Alaska and British Columbia, over the deeper saline waters (Royer, 1982). A strong permanent density gradient results from a combination of the salinity gradient above and a strong shallow seasonal thermocline, thus locking the deeper water from effective exchange with the surface waters. The diffusive flux of water through this barrier is weak with a vertical displacement of only 8 m yr$^{-1}$. The total advective and diffusive vertical transport into the upper subarctic waters is about 23 m yr$^{-1}$. In the western Pacific Ocean, the surface circulation is dominated by the warm Kuroshio Current, with volume transport of over 88 Sv (10$^6$ m$^3$s$^{-1}$) as estimated by Worthington and Kawai (1972). The very strong shear of the Kuroshio favors downward mixing into the deeper ocean. The subtropical gyre area has strong down-welling of water with velocity up to 20 m yr$^{-1}$ downwards (Gargett, 1991). The CO$_2$ penetration absorbed from the atmosphere has already taking place, now mixed down into 1,200 m depths in the western boundary current and subtropical gyre, also evident in the C-13/C-12 ratios in Quay et al. (1992). The intermediate waters, formed by the intermittent sinking of winter cold waters of high salinity off the Kuril Island and the Kamchatka Peninsula, would carry the CO$_2$ disposed into the sinking water mass, to depths of 200–1,200 m, spreading eastwards under the subarctic gyre and southwards to waters under the North Pacific Drift (Reid, Jr., 1965; Talley and Joyce, 1992). The intermediate water will ventilate along the 26.8 sigma-T density surface in the N.E. Pacific Ocean and outcrops in the Alaska Gyre (vanScy et al., 1991). The time-scale of the ventilation is about 20 years, thus not effective for using intermediate water CO$_2$ disposal to achieve delay time.

The deep and bottom waters of the North Pacific Ocean are relatively poorly known (Tabata, 1975). Chemical tracer study such as $^3$He indicates a spread from the hydrothermal vent site in the Eastern Pacific Rise northwards and westwards in depths of 2,000–3,000 m. The deep Pacific Ocean has only one source water, the Antarctic Bottom Water formed originally in the Weddell Sea shelf and with minor components from the Ross Sea and Amery Ice Shelf in the Antarctica, mixed also with the Antarctic Circumpolar waters. The Antarctic Bottom Water flows northward along the western side of the Pacific Basin, along the deep trench east of New Zealand at velocity of about 8–25 Sv (Reid, 1968; Bolin and Stommel,
averaged to about 15 Sv along a width of 100 km at depths 2,500 to 4,500 m flowing to as far north as offshore Japan and then spreading into the eastern part of the deep Pacific Ocean (Tabata, 1975). Deep zonal flows also occur near Hawaii in an eastward direction, then turning north (Edmond et al., 1971), and along the depressions on the fracture zones along 10°N and 5°S as zonal eastward flows (Wong, 1972). Between the deep waters of the Pacific and the intermediate water, there is a return flow southwards at 2,000 m indicated in the silicate distribution (Talley and Joyce, 1992). On the global scale, the North Pacific Ocean is at the end of the “oceanic conveyor belt” of CO₂ and water originated in the region of deep water formation area in the Greenland and Labrador Seas (Broecker and Peng, 1982). Thus, the disposed CO₂ into the deep Pacific waters would be mixed into the deep water reservoir, then into the southward return flow and upwelled and ventilated with the atmosphere. The time-scale is of the order 500 years, based on replacement of the deep Pacific waters by Circumpolar waters and a general Pacific Ocean upwelling of 25 Sv (Warren, 1973). For disposed CO₂ in the northern end of the Pacific Basin, the delay time in returning to the upper ocean is probably of the order of 200–300 years.

For CO₂ disposed into the depths below the upper ocean, it will be mixed downward into the deep ocean water at a rate of 30 m yr⁻¹ (Wong and Matear, 1993), inferred from C-13/C-12 and C-14/C-12 changes in the Pacific Ocean between 1970 and 1990 (Quay et al., 1992). The anthropogenic CO₂ absorbed by the Pacific Ocean from the atmosphere also illustrates the CO₂ penetration into the ocean’s interior. The anthropogenic CO₂ has penetrated to about 1,200 m in the western subtropical Pacific and 400 m in the eastern side (Chen, 1980) i.e. approximately 800 m on the average during the period of 1950–1975. The inferred penetration downwards is 33 m yr⁻¹. Thus, the dilution of 30 m yr⁻¹ of deeper water, with its carbonate ions will provide an additional neutralizing capacity of 13 GtC yr⁻¹. CO₂ disposed into the 1,000 m depth range will eventually reach the sea floor through this downward mixing and diffusive process.


A North Pacific Carbon Cycle Model (Matear, 1993) made use of the integrated meridional transport in the North Pacific by adopting the velocity field of the Hamburg Large-Scale Geostrophic Ocean Circulation-Biogeochemical Model of Bacastow and Maier-Reimer (1990) in a 18-box model with the boxes defined by zonal sections of 10°N–24°N, 24°N–47°N and 47°N–65°N, and by 6 layers of photic zone (0–100 m), surface water (100 m to density surface of sigma \( \theta \) 26.8), North Pacific Intermediate Water (sigma \( \theta \) 26.8 to 27.3), Antarctic Water (sigma \( \theta \) 27.7 to 27.8) and Antarctic Bottom Water (sigma \( \theta \) 27.8 to sea floor), as shown in Fig. 4. The carbon dioxide parameters were obtained using the thermodynamic relations for the oceanic CO₂ chemical systems. The annual mean ocean circulation from the Hamburg Large-Scale Geostrophic ocean circulation model
Fig. 4. Water masses in boxes defined in the Matear Model of the Carbon Cycle of the N. Pacific Ocean.
along with the monthly convective events were used to define the global ocean circulation on a 5 × 5 degree grid with 15 layers (Matear, 1993). The added carbon is treated as a perturbation from the steady state model, allowing this carbon to be treated as a passive tracer. This carbon tracer is transported using the positive definite advection scheme of Smolarkiewicz (1983) with a time step of one week. The advection scheme consists of an upwind advection with an additional step to partially correct for the numerical diffusion of the upwind advection scheme. The numerics of the scheme introduces a diffusion like term into the advection equation. In the model, carbon is also exchanged with the atmosphere at every time step using the partial pressure difference of CO₂ between the ocean and atmosphere and the monthly gas exchange coefficients of Merlivat. Additional carbon transport is performed monthly to simulate the convective mixing that was required in the Hamburg model to ensure density stability. These convective events were saved when the physical model was run and they are applied in the transport of the carbon tracer. When required, the convective mixing cause uniform mixing of two adjacent layers. Such a process could also be modelled as a diffusive process which causes the deepening of the mixed layer.

Using this circulation field, a reference run was performed starting with an atmospheric concentration of 350 μatm. and increasing it by 2 μatm yr⁻¹. The reference model reflects how the content in the ocean is increased by the additional amount of carbon in the atmosphere. Next, carbon was added to the ocean at 32°N and 157°E at several different depths at a rate of 0.25 GtC yr⁻¹ (approximately one-half of the total carbon output of Japan) in this disposal scenario. This added carbon is transported by the ocean with some carbon eventually reaching the surface waters. This additional carbon in the surface water increases the pCO₂ of the surface ocean, which affects the air-sea CO₂ flux. The effect of the CO₂ disposal on the natural uptake of CO₂ by the ocean is determined by the fraction of excess CO₂ remaining in the ocean, defined as the ratio of the (ocean carbon content in the disposal run minus ocean carbon content in the reference model)/total carbon disposed into the ocean up to that time.

In this hypothetical case, the continuous discharge from a point source at 32°N and 157°E at 0.25 GtC yr⁻¹ is made at depths of 425 m, 1,200 m, 3,000 m, and 5,250 m to obtain an output of fractions of excess CO₂ still remaining in the ocean to the total amount of CO₂ disposed into the North Pacific, for a period of 200 years as summarized in Table 3 and in Fig. 5. For disposal into shallow depth of 425 m, the disposed CO₂ will reach the surface in about 20 years and 23 GtC of the 50 GtC disposed will be released back into the atmosphere in 200 years. At 1,200 m, about 25% will be returned to the atmosphere. Disposal at depths greater than 3,000 m will have virtually no CO₂ escaping back to the upper ocean. Thus, disposal into the deep Pacific Common Water at 0.25 GtC yr⁻¹ does not have impact on the atmospheric CO₂ level for a 200 years time-scale, i.e. the same as the time-scale of ventilation of this deep water.
Table 3. Fraction of excess CO₂ for disposed CO₂ introduced at different depths continuously over different time periods.

<table>
<thead>
<tr>
<th>Number of Years</th>
<th>Integrated Amount of CO₂ Introduced (GtC)</th>
<th>Fraction remaining for depths of introduction</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>425 m.</td>
</tr>
<tr>
<td>50</td>
<td>12.5</td>
<td>0.94</td>
</tr>
<tr>
<td>100</td>
<td>25.0</td>
<td>0.78</td>
</tr>
<tr>
<td>150</td>
<td>37.5</td>
<td>0.64</td>
</tr>
<tr>
<td>200</td>
<td>50.0</td>
<td>0.54</td>
</tr>
</tbody>
</table>

DUMPING of 0.25 Gt C Per YEAR

Fig. 5. Fraction of disposed CO₂ still remaining in the ocean, as a function of depth of disposal and time, for a scenario of continuous discharge of 0.25 GtC yr⁻¹ off Japan.
7. Disposal Site and Strategy

Ideally, CO₂ disposal strategy would include utilization of the following characteristics: (1) Pacific deep water greater than 3,000 m depth to make use of the deep water capacity and dilution and optimization of the delay time, (2) availability of sediment carbonate to neutralize the CO₂, (3) isolated basin if specialized storage, e.g. liquified CO₂ pool, or to obtain longer delay time, (4) long residence time of the water mass, and (5) acceptability of the environmental sensitivity. The Pacific Ocean has low carbonate contents in the sediment, relatively shorter residence of the deep water before in contact with the atmosphere and lack of isolated basins. For disposal at a rate of 0.25 GtC yr⁻¹, criterion (1) is sufficient to achieve the objective. For larger quantities, disposal into isolated tropical basins with more carbonate available in the sediment is more desirable to
use basins with criteria (2) and (3). The Philippines Basin, the Caroline Basin and the Guatemala Basin, all deeper than 4,000 m, with 1–30% carbonate contents in the sediment are shown in Fig. 6. In the subarctic waters in the North Pacific, waters deeper than 3,000 m are found in the southern part of Okhotsk Sea near the Kuril Islands and in the western part of the Bering Sea. In the polar seas, the deeper waters are well stratified and the surface ice and freshwater provide an effective lock of deep water exchange with the atmosphere. Little excess CO₂ from the atmosphere would penetrate into the deep polar waters, e.g. no excess CO₂ reaching 500–5,500 m depth in the Weddell Sea waters (Poisson and Chen, 1987). The Canada Arctic Basin north of Beaufort Sea, with similar oceanographic properties in waters deeper than 3,000 m, is an interesting water body with high concentrations of gas hydrates due to the extreme cold. The waters also have a long residence time of 500 years (Anderson and Dyrrsen, 1989) Actual disposal at these sites requires further scrutiny of the environmental consequence, the feasibility of the disposal technology and the economic factors.

8. Conclusions

(1) The North Pacific Ocean has low carbonate contents in the deep-sea sediments north of the tropical region, but its large volume offers dilution and utilization of the carbonate ions as buffering agent.

(2) Direct CO₂ disposal into the Pacific deep waters at depths below 3,000 m will not have effects on the atmospheric CO₂ level, if the quantity is kept low at 0.25 GtC yr⁻¹ over a period of 200 years, as shown in a model study.

(3) The criteria for disposal sites are: (a) disposal depths greater than 3,000 m, (b) availability of high carbonate contents in the sediment, (c) isolated basin and (d) long residence time of the water mass receiving the CO₂ disposal.

(4) Few sites satisfy all the criteria, but their usefulness depends on the method of disposal. Basins with advantages are the Philippines Basin, Caroline Basin, Guatemala Basin, Okhotsk Sea and Bering Sea. Outside the North Pacific Ocean, the basins in the South Pacific with high carbonate contents in the sediment and deep waters in the Canada Basin in the Arctic satisfy more of these criteria.

(5) Surface waters and intermediate water dispositions have the disadvantage in not being able to delay CO₂ return to the atmosphere.

(6) The upper ocean has a limited buffering capacity to absorb atmospheric CO₂. It is being depleted already by excess CO₂ from the atmosphere, at a rate of 2 GtC yr⁻¹ globally. Disposal into the surface waters should be avoided to preserve this buffering layer.

(7) Disposal should not proceed without international consensus on the environmental effects, economic factors of disposal and feasibility of disposal technology.
REFERENCES


**DISCUSSION**

**SuginoHarra:**
What type of model is used for your CO₂ budget in the North Pacific, particularly the model for circulation. You mentioned the Hamburg Model?

**Wong:**
The model is essentially set up of box model. And then we essentially use the velocity output of the Hamburg Model to put it in the model to generate CO₂ circulation.

**SuginoHarra:**
And so I think the distribution of CO₂ highly depends upon the effect of eddy diffusivity and circulation. So what value is used for the eddy diffusivity.

**Wong:**
Essentially there was estimate of different diffusivity between different boxes. And there would be also horizontal diffusivity estimated as well. But the basic point is that the horizontal diffusivity is much larger. Probably an order of 100 larger than the vertical diffusivity. So whatever you inject in the mid depth it would go sideways before it goes up and down.

**Morishita:**
I would like to ask your comments regarding the criteria for the ocean disposal. You presented several basins in the northern Pacific Ocean for candidate sites. How do you evaluate the acceptability of the environmental sensitivity in the deep ocean? Is it just based on the pH value or ...?
Wong:

There would be several questions to ask. I presume it is mainly the question on the water itself. The change in pH is probably not large enough to cause problem. But if you dispose CO₂ in the form of liquified CO₂, in the site of ocean bottom, then the impact may be different and also we do not know much about deep sea biology, either. But I am not an expert in deep sea biology, but for water itself, a quarter of Gt per year probably will be okay in a hundred or two hundred years time scale.

Question:

I would like to try and enlarge the discussion a little bit. It seems that we have two types of work going on. One is on a macro scale which Dr. Wong talk about massive ocean circulation. And then we got talk about disposing into much more isolated sites. Now, the characteristics of these isolated sites might, sort of, gets around some of the problems Dr. Wong talked about. Because he really is presenting a fairy-like picture in that he is saying that even you inject to a depth of 3000 meters, we are talking about perhaps a maximum retention time 200 years. Is this the case on much more localized scale? Can not there be a local site where it might not be the case?

Suginohara:

As an oceanographer, we should adhere to much larger scale. For localized scale, I guess the basic question is to define what kind of disposal technique, that the so-called site-specific approach, and I guess that probably fall in a different category, for example, if it is within 200 miles limit of fish resource, management ... then I guess ... probably will be handled by individual country.

Evans:

I have a question for Dr. Wong. Your conclusion reads “Direct CO₂ disposal into the Pacific deep waters at depths below three thousand meters will not have effects on the atmospheric CO₂ level if the quantity is kept low at 0.25 GtC per year over a period of 200 years, as shown in a study model.” I wonder if you could rephrase that for me, please. So the sea would act as a repository for 200 years, is what you are saying?

Wong:

In a general sense, yes. Except that that kind of time scale is actually based on an old carbon-14 study done in the 70s. So and if I recall carbon-14 age and a number of carbon-14 measurements in the North Pacific is not numerous in the North Pacific, that is probably a kind of a generalized conclusion one can have based on essentially pretty old data set. So possibly again after more measurement I presume some carbon-14 age will be done at Woods Hole, then perhaps we can have a better estimate at that time.
Evans:
I see. Would you say these are fairly conservative figures? I have heard of much longer time period for it.

Wong:
I don’t think it would be much longer in the sense that the Stuiver’s recalculation in 1983 based on GOSECS carbon-14 data then actually ... time of 500 years for the North Pacific.

Question:
You showed a vertical profile of isotopic ratio of carbon in comparison between 1970 and 1990, and I think your inventory of the penetration of CO2 in the ocean is the based on the difference in the 1970 and 1990 data. So did you consider the physical condition in the 1990 and 1970. So how did you determine the same physical situation in the 1990 and 1970. For example, you compared with the density, the potential density of seawater or not?

Wong:
To derive at that kind of situation, it is essentially the kind of generalized figure would be based on eight different stations selected in the Pacific where we have four 1970 data on the Canadian Hudson and similar data were available on the North to 1990. In fact, those with in those stations they are not really uniform but when you sort up data and generalize them, you obtained this figure, i.e. 35 meter penetration per year. But the confidence in that was also confirmed by looking at carbon-14 isotope and it was close to 26 meter per year. So based on both, I guess we do arrive at that 30 meters per year, for the Pacific. But that is the only experimental data we have.