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OPTIONS FOR ENHANCING THE STORAGE OF CARBON DIOXIDE IN THE OCEANS: A REVIEW

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ABSTRACT

The various proposed options for enhancing CO₂ storage are often ingenious but rarely feasible. Fertilizing the central gyres would require enormous amounts of N and P fertilizer. Addition of iron to the Southern Ocean has been rejected as inefficient. The deep sea disposal of 50-100% of fossil fuel CO₂ would suppress the predicted transient peak of atmospheric CO₂ but eventually the atmospheric CO₂ level will be the same as without, unless rapid buffering by dissolution of calcite is significant. Ocean disposal applies only to point sources representing ~30% of fossil fuel emissions; relocation of these plants to maritime regions may not be economical either. Also ocean disposal costs ~30-45% of the energy produced. Current research is focusing on production and oceanic mixing behaviour of the effluent stream (this volume). In this paper the fate of the disposed CO₂ is assessed. The various proposed forms appear chemically unstable relative to ambient seawater. However the higher CO₂ contents of seawater might lead to part of the CO₂ being buffered forever through enhanced dissolution of existing calcite deposits. The predicted impact depends on the selected rate constant of latter dissolution. Ocean dumping is a fallback option, in case the sensible options of energy conservation and shifting to other energy sources would not be implemented adequately. Latter options would require societal and economic adaptations but appear the most efficient, if not inevitable, policy for imminent reduction of CO₂ emissions.

INTRODUCTION

The low level of about 0.03 % carbon dioxide (CO₂) in our atmosphere has largely been achieved by the marine biota. Over hundreds of millions of years some 50x10¹⁸kg or 50 million Gigaton C (GtC) as limestone (CaCO₃) and another 12 million GtC organic matter has accumulated within deep sea sediments (Fig. 1). Another portion is stored in continental soils, partly also as fossilized coal. As a result the atmosphere contains only about 700 GtC which exchanges at a rate of ~40-100 GtC/y with the surface ocean. Over several geological time scales this small amount of CO₂ in the atmosphere has varied, often preceded by orbital forcing and closely followed (Shackleton and Pisias, 1985) by variations in global temperature (Barnola *et al.*, 1987; Berner, 1991). Mankind has burned increasing amounts of fossil fuels and the atmospheric CO₂ rapidly increased with ~2-3 GtC/y from about 0.028% (280 x 10⁻⁶ atm) in 1850 to about 0.036% today, very fast compared to the changes over geological time. The biosphere has lived through previous dramatic change events, but not without massive extinctions. Comparison with industrial production records has learned that only ~60% of the emissions remain in the atmosphere, the other ~40% apparently is taken up by the oceans (Goudriaan, 1990).

In temperate zone surface waters the CO₂-fixation by marine algae is limited severely by depletion of essential nutrients nitrogen (N) or phosphorus (P). In high latitude waters the productivity may be limited by a variety of other factors (Lancelot *et al.*, 1992): low ambient light, also due to ice cover and intense cloud cover; absence of iron (Fe) as required in minute amounts for algal growth; or intense grazing pressure keeping algal biomass hence productivity at minimal levels. The relative importance of and interaction between these various factors is not well understood. Nevertheless there is unanimous agreement that shifts of marine biological productivity, notably in (sub)polar waters, exert a major control on changes of atmospheric CO₂, hence climate (Broecker and Denton, 1989; Charles and Fairbanks, 1992).

The pathways for net transfer of CO₂ from the atmosphere into the deep ocean are several, all of them necessarily going through surface waters. These euphotic surface waters, where penetrating light is driving photosynthetic CO₂ fixation, are well separated from the deep waters by a strong thermal gradient (at ~200-600m depth) preventing vertical mixing between surface and deep ocean. Only in the polar regions sinking of cold waters occurs in the northern North Atlantic Ocean and the Antarctic Weddell Sea during the northern and austral winters respectively. The cold waters draw down quite some dissolved CO₂, as well as Dissolved Organic Carbon (DOC). In addition to these two pathways there is a worldwide raindown of

Fig. 1. The global carbon cycle with inventories in Gigaton C (1 GtC=10¹⁵ gram) and annual fluxes in GtC/y. Large bold arrows denote natural fluxes, small arrows fluxes caused by mankind. All values are approximations.

biological debris, settling particles containing both organic C as well as biogenic CaCO₃ tests. This export production is only a few percent of the gross photosynthetic C fixation which is mostly recycled again to CO₂ within the complex foodweb of the euphotic zone (De Baar and Stoll, 1989). Once in the deep ocean most of the exported particles are again mineralized to CO₂: consumed by bacteria, devoured by animals living on the seafloor, or slowly dissolving (the CaCO₃). However a very small percentage is buried into the sediments, eventually leading to the massive C reservoirs mentioned above.

Intentional enhancement of oceanic CO₂ uptake has been suggested to remedy the expected increase of atmospheric CO₂ by fossil fuel burning. One concern here is the implicitly assumed right to solve an environmental problem in one compartment of the biosphere by tinkering with yet another compartment.

Leaving aside this important matter for the moment one would allow oneself to examine the feasibility of schemes for enhancing oceanic CO₂ storage. The various schemes rely either on intentional fertilization or on bypassing both the atmosphere and the surface ocean by collecting fossil fuel CO₂ directly at its source and pumping it into the deep sea.

INTENTIONAL FERTILIZATION.

Macronutrients.

The addition of N and P to nutrient starved central gyres of the oceans in temperate zones would extend the current coastal eutrophication into the open ocean. Recently the possible effect of eutrophication in the North Sea on the atmospheric CO₂ budget was assessed. Despite major uncertainties it was deemed trivial: enhancement of gross photosynthesis is rapidly mineralized again i.e. no net increase of storage of C in the sediments (Hoppema and De Baar, 1991). The shallow North Sea is very well mixed vertically, there is no deep water layer isolated from exchange with the atmosphere. On the other hand the central ocean surface waters are well separated from the underlying deep waters. Settling biogenic particles transport C down into the deep waters where they are mineralized again but the ensuing dissolved CO₂ is out of contact with the atmosphere. This biological pump already operates but might be enhanced. Marine algae take up C, N and P in fairly uniform proportions C:N:P=106:16:1 (Redfield *et al.*, 1963). For each additional 1 GtC/y to be fixed some 0.15 GtN/y and 0.01 GtP/y are needed. When N and P are recycled less efficiently than C in the surface ocean ecosystem they would be exported downward more rapidly; then the proportions of added N and P would have to be raised versus C to be removed. Clearly the amounts of N and P required would be enormous, as is the challenge of spreading this evenly over large and remote areas. The one initiative to work out such scheme apparently was abandoned (Broecker, 1977).

Fe fertilization.

In 1987 Fe enrichment experiments were done in the sub-Arctic North Pacific Ocean, where results hinted at possible Fe limitation (Martin and Fitzwater, 1988). Latter conclusion was challenged (Banse, 1990). On the other hand an independent experiment run simultaneously also supports the Fe limitation concept for this region (Coale, 1991). From this work in the subArctic it was extrapolated repeatedly that Fe would be the limiting factor for all of the Southern Ocean (Martin *et al.*, 1988; 1989; Martin, 1990), but direct evidence was lacking. Esoteric schemes for spreading Fe over all the Southern Ocean followed (Anonymous, 1990).

Fig. 2. Stimulation of phytoplankton growth (Chlorophyll a) upon addition of 1 nM, 1 nM + daily extra 1nM and 10nM Fe to seawater collected at 59°S, 49°W in the Weddell/Scotia Seas. Note that untreated controls also outgrow the initial field conditions (DeBaar *et al.*, 1990).

The first Antarctic Fe enrichment experiments were executed in 1988 in the Weddell-Scotia Seas (De Baar *et al.*, 1989; 1990). Here Fe was found to stimulate biomass, however the untreated controls also outgrew the field (Fig. 2). Latter observation was ascribed to the exclusion of larger grazers from the experiments (Burna *et al.*, 1991), i.e. in the field the algal biomass would be kept low by intense grazing. In 1989 very similar results were obtained at 3 stations in the Ross Sea but this led to the opposite conclusion, i.e. Fe presumably being the major growth limiting factor (Martin *et al.*, 1990a). Latter conclusion was challenged by Dugdale and Wilkerson (1990) whom argued in favor of biological loss terms, i.e. grazing or sedimentation. Unfortunately no good information was given on the plankton community. However in the same study a

fourth experiment in far offshore waters provided some convincing evidence of the sought after Fe limitation (Martin *et al.*, 1990a). In offshore Antarctic waters the Fe levels may well be very low (e.g. Drake Passage, Martin *et al.*, 1990b) and limiting. Studies in offshore waters are scheduled by our group for late 1992.

Even when assuming that all nutrients in Antarctic surface waters can be converted to organic matter (e.g. by Fe fertilization), the predicted drawdown of CO₂ is modest (Peng and Broecker, 1991; Joos *et al.*, 1991). As soon as the action, e.g. fertilization with Fe, would be abandoned, the extra CO₂ taken up would rapidly be released again into the atmosphere. At a special symposium in February 1991 all evidence was presented (Chisholm and Morel, 1991) and discussed with the conclusion that intentional Fe fertilization would not likely have an effect, i.e. not to be pursued any longer (ASLO, 1991). On the other hand the role of Fe as one of several factors affecting marine productivity was clearly recognized.

DISPOSAL OF CO₂ IN THE DEEP OCEAN

The various technical reports have been reviewed elsewhere (Marland, 1986; De Baar and Stoll, 1989; De Baar, 1990, 1992). Briefly one remarkable first publication (Marchetti, 1977) led to one published feasibility study (Baes *et al.*, 1980) where several options had been suggested:

- a. High density solution in seawater; the reject stream would sink to the ocean floor.
- b. Pure, liquefied CO₂ emitted at depths greater than 3000m for negative buoyancy.
- c. Dumping solidified blocks of CO₂-hydrate at depths below 500m as to avoid bubbling.
- d. Dumping solidified CO₂ (dry ice), at depths below 500m as to avoid bubbling.

For a power plant at the sea surface it is possible in principle to produce any of the above four states. The energy requirements for pressurization are not the issue here, although they would be quite significant, in the order of 30% of the energy produced when assuming a pressure of 33 atm is adequate, i.e. disposal at about 330m depth. Unfortunately the publication is very brief (and occasionally inconsistent) on the chemical stability of the material during its downfall as well as when residing at the seafloor.

Solubility of CO₂ in seawater

The partial pressure or vapor pressure of a given gas is one of the master variables. For example in our atmosphere the partial pressure of CO₂ is only about 360×10^{-6} atm at a total pressure of 1 atm. Only in the case of a pure gas is its vapor pressure (fugacity) the same as the total pressure. The distinction between vapor and total pressure appears obvious but it has sometimes been confused. For equilibrium of CO₂ between two phases α and β the chemical potentials in both phases ought to be equal

$$\mu^{\alpha}_{\text{CO}_2} = \mu^{\beta}_{\text{CO}_2}$$

This can be written out as:

$$\mu^{\alpha,0}_{\text{CO}_2} + RT \ln a^{\alpha}_{\text{CO}_2} = \mu^{\beta,0}_{\text{CO}_2} + RT \ln a^{\beta}_{\text{CO}_2}$$

where a_{CO_2} is the activity of CO₂ in the given phase and $\mu^{\alpha,0}_{\text{CO}_2}$ and $\mu^{\beta,0}_{\text{CO}_2}$ are the standard chemical potentials of CO₂ in both phases. The first problem is to assess latter parameters for each phase. Next some assessment of the activity a is required. In general the activity of a gas equals its fugacity and under ideal conditions both are also the same as its partial pressure. Under non-ideal conditions, notably at high overall pressures or deviating temperatures, this is no longer true and one defines an activity coefficient to relate partial pressure with activity.

The assumed standard temperature, salinity and depth of the deep ocean site are 2°C, S=35 at 4000m depth and represent median values (Mc Lellan, 1974). The solubility of CO₂ is a function of temperature and salinity (Weiss, 1974) at 1 atm ambient (i.e. total) pressure. At (T=2°C, S=35, p=1atm) this solubility or Henry's law constant k is $0.05822 \text{ mol.kg}^{-1}.\text{atm}^{-1}$. For an atmospheric partial pressure of 360×10^{-6} atm this leads to the dissolved (CO₂)_{aqueous} being $20.96 \text{ } \mu\text{mol.kg}^{-1}$ in the surface water. This is furthermore in equilibrium with major dissolved forms, namely bicarbonate (HCO₃⁻ at about $1800\text{--}2200 \text{ } \mu\text{mol.kg}^{-1}$) and carbonate (CO₃²⁻ at about $50\text{--}200 \text{ } \mu\text{mol.kg}^{-1}$) as well as a negligible amount of H₂CO₃ ($\sim 0.02 \text{ } \mu\text{mol.kg}^{-1}$). The sum of concentrations of these various forms of CO₂ is known as ΣCO_2 :

$$\Sigma\text{CO}_2 = [\text{CO}_2]_{\text{aqueous}} + [\text{H}_2\text{CO}_3] + [\text{HCO}_3^-] + [\text{CO}_3^{2-}]$$

The equilibria between these various forms are governed by equilibrium constants which in themselves are functions of ambient temperature and pressure. For simplicity it is assumed the deep water at the disposal site reflects equilibrium with today's atmosphere and contains above $[\text{CO}_2]_{\text{aq.}}$ at $20.96 \text{ } \mu\text{mol.kg}^{-1}$. Latter value applies to surface conditions (1 atm). Readjustment to the ambient pressure of 400 atm does not affect

ΣCO_2 but will slightly shift the equilibria such that $[\text{CO}_2]_{\text{aq}}$ decreases to an in situ concentration of $\sim 20 \mu\text{mol.kg}^{-1}$ ($T=2^\circ\text{C}$, $S=35$, $p=400 \text{ atm}$).

Disequilibrium versus ambient seawater in the deep ocean.

a. High density solution in seawater. Seawater is equilibrated with pressurized (33 atm) pure CO₂-gas (Baes *et al.*, 1980). The activity coefficient of CO₂ is about 0.75 (2°C , 33 atm) and its activity becomes $0.75 \times 33 \sim 25 \text{ atm}$. Next the above solubility of $0.05822 \text{ mol.kg}^{-1} \cdot \text{atm}^{-1}$ at standard conditions ($T=2^\circ\text{C}$, $S=35$, $p=1 \text{ atm}$) needs adjustment to conditions with a total pressure of 33 atm. At higher ambient pressure the solubility of CO₂ increases (Millero, 1983). For $P=33 \text{ atm}$ the solubility is $0.062 \text{ mol.kg}^{-1} \cdot \text{atm}^{-1}$ ($T=2^\circ\text{C}$, $S=35$, $p=33 \text{ atm}$). The dissolved concentration then would be $[\text{CO}_2]_{\text{aq}} \sim 1.5 \text{ mol.kg}^{-1}$ similar to values shown by Baes *et al.* (1980). The amount of CO₂ dissolved in the seawater will be three orders of magnitude more than the original ΣCO_2 and five orders of magnitude above the original dissolved $[\text{CO}_2]_{\text{aq}}$. When bringing this water down to 4000m depth the solubility would increase to $0.1048 \text{ mol.kg}^{-1} \cdot \text{atm}^{-1}$ at ($T=2^\circ\text{C}$, $S=35$, $p=400 \text{ atm}$). As such the dissolved $[\text{CO}_2]_{\text{aq}} \sim 1.5 \text{ mol.kg}^{-1}$ would not have the tendency to escape. However the surrounding untreated seawater would have exactly the same solubility ($0.1048 \text{ mol.kg}^{-1} \cdot \text{atm}^{-1}$) but contains only $[\text{CO}_2]_{\text{aq}} \sim 20 \mu\text{M.kg}^{-1}$. The concentration gradient of five orders of magnitude will drive the disposed CO₂ into surrounding oceanic waters.

Baes *et al.* (1980) seemingly equate the partial pressure of CO₂ with the hydrostatic pressure, i.e. "the minimum depth of injection without bubble formation". Obviously *within* the waste stream itself there would indeed be no tendency for CO₂ bubbles to escape. However this should not be interpreted as the waste stream being chemically stable. Even if bubbles would form there would be immediate dissolution in the surrounding seawater which contains only $[\text{CO}_2]_{\text{aq}} \sim 21 \mu\text{mol.kg}^{-1}$. At any depth (whether above or below the suggested depth level) the injected $[\text{CO}_2]_{\text{aq}}$ will tend to diffuse into the surrounding waters.

b. Pure liquefied CO₂. Negative buoyancy of the liquid CO₂ stream can be achieved at a temperature between 0°C and 10°C and vapor pressures greater than 300 atm, i.e. injection at depths greater than 3000m (Baes *et al.*, 1980). Chemical stability should properly be assessed from comparison of chemical potential versus that of ambient seawater ($T=2^\circ\text{C}$, $S=35$, $p=400 \text{ atm}$). Here the activity of CO₂ in both the seawater and the pure liquefied CO₂ is important. Given the concentration $[\text{CO}_2]_{\text{aq}} \sim 20 \mu\text{mol.kg}^{-1}$ in ambient seawater and an activity coefficient of ~ 1.23 (Markham and Kobe, 1941) the activity would be $(\text{CO}_2)_{\text{aq}} \sim 25 \mu\text{mol.kg}^{-1}$. In general the activity of a pure liquid is taken as unity at standard pressure (1 atm). The effect of pressure (400 atm) needs assessment by proponents of this scheme, but the activity of the pure liquid CO₂ likely will be orders of magnitude higher than in ambient seawater. More simple it might be stated that the partial or vapor pressure versus the liquid CO₂ would be 300 atm as compared to $360 \times 10^{-6} \text{ atm}$ in the ambient seawater. Hence the effluent stream would likely be grossly oversaturated versus ambient seawater, i.e. the liquefied CO₂ would in essence be unstable and dissolve into the seawater, where (under proper conditions) the below hydrates may act as transient intermediates (Sakai *et al.*, 1990).

c. Solid CO₂-hydrate. The CO₂-hydrate consists of a cubic unit cell of 46 water molecules forming 8 cages into which the CO₂ can fit, also known as a clathrate structure. The atmosphere of Mars consists for about 95 % of CO₂ with a partial pressure of $6.5 \times 10^{-3} \text{ atm}$. and the low temperatures ($\sim 120^\circ\text{C}$) at the Martian ice cap would be suitable for the existence of CO₂-hydrates in the ice (Miller and Smythe, 1970). The upper critical decomposition temperature was cited to be 10.2°C at a pressure of 44.5 atm above which CO₂-hydrate is unstable relative to liquid CO₂ and H₂O (Herreillers, 1936; Larson, 1955). This relates to CO₂-hydrate versus a system of pure CO₂ and (sea)water, for example as observed at submarine gas vents (Sakai *et al.*, 1990). Apparently this led Baes *et al.* (1980) to suggest that at hydrostatic pressures greater than $\sim 50 \text{ atm}$ (depths $> 500 \text{ m}$) and temperature less than 12°C no more gas bubbles would form from the decomposition of the CO₂-hydrate. As with above option (a) this may not be incorrect, however this should not be interpreted as the CO₂-hydrate being stable. On the contrary the *partial* pressure of CO₂ in the ambient seawater would still be only about $360 \times 10^{-6} \text{ atm}$, i.e. five orders of magnitude below the $\sim 44.5 \text{ atm}$ partial pressure of CO₂ as required for the CO₂-hydrate to be stable. Again this should be properly calculated through comparison of chemical potentials, where the activity of CO₂-hydrate is likely in the order of unity. Presumably both during settling as well as on the seafloor the CO₂-hydrate would tend to decompose into the surrounding seawater, as recognized by Baes *et al.* (1980). Similarly those submarine formations of CO₂-hydrate in direct contact with overlying seawater should be seen as transient rather than permanent features (Sakai *et al.*, 1990).

Some inconsistency about partial pressure and total pressure is apparent when reading the option of storing CO₂-hydrate at the South Pole. Here it was suggested that "the decomposition pressure of CO₂-hydrate does not fall below one atmosphere until it is cooled to about -54°C " according to (Baes *et al.*, 1980). This is consistent with the findings of Miller and Smythe (1970) for CO₂-hydrate versus the *partial* pressure of CO₂ being 1 atm at -54°C (219°K). However for the surface of the earth the partial pressure of CO₂ is only $360 \times 10^{-6} \text{ atm}$ at a *total* pressure of 1 atm.

d. Solid CO₂ (dry ice). Given the activity (CO₂)_{aq} ≈ 24 μmol.kg⁻¹ in ambient seawater it is obvious that the activity in the pure solid CO₂ is orders of magnitude higher. At standard pressure (1 atm) the activity of solid CO₂ would be near unity, the exact value at 400 atm needs assessment. However the vapor pressures for stability of dry ice are hundreds of atmospheres and would suggest the effluent stream to be grossly oversaturated versus ambient seawater. Also the temperature of dry ice is about -80°C or lower. The solidified CO₂ would be unstable as it tends to dissolve into the seawater by diffusion of both CO₂ and heat. As a matter of fact the likely reactivity with seawater has been recognized by Baes *et al.* (1980) mentioning possible reaction intermediates as CO₂-hydrate or CO₂-rich solution.

Vertical stratification.

Summarizing it is not impossible, when disregarding the expenditure, to produce any of the four physical states by selecting appropriate temperatures and pressures. This is feasible at a platform at the sea surface but also applies to natural CO₂ emissions from ocean floor vents (Sakai *et al.*, 1990). However all four physical states are (a) or appear (b,c,d) very unstable with respect to chemical dissolution in ambient seawater. The rate of dissolution depends on the rate of turbulent mixing between the disposal and the deep ocean waters. The turbulence occurring during downward transport of the plume has been studied (Baes *et al.*, 1980) and is the focus also of ongoing research (Herzog *et al.*, 1991; Zemba *et al.*, 1992). Upon arrival at the seafloor the stockpile of CO₂ in any form would remain there due to this higher density, however at its interface with overlying waters erosion would occur continuously. The rate of erosion is controlled by turbulence caused by shear stress with the lateral flow of deep ocean currents. For normal ocean basins the disposed CO₂ can therefore be modeled along with the regular deep water renewal and eventually the CO₂ would return to the atmosphere (Hoffert *et al.*, 1979). Depending on the disposal site the injected CO₂ would re-equilibrate with the atmosphere after periods in the order of ~50-500 years.

Basins with restricted deep circulation might be preferable as storage sites. For example the Deep Black Sea is stagnant (Murray *et al.*, 1991), where the inflowing high salinity Mediterranean water over the shallow Bosphorus sill may not be unsuitable for CO₂ injection. Currently most of this inflowing Mediterranean water mixes with low salinity subsurface waters and flows out again at the surface of the Bosphorus, yet part of it might, at least in principle, be diverted directly into the deep (>2000m) Black Sea. However in Black Sea bottom waters there is strong evidence for deep convection driven by geothermal heating (Öszozy *et al.*, 1991), i.e. considerable vertical mixing (Brunt-Vaisala frequencies < 0.01 h⁻²; Murray *et al.*, 1991).

CALCITE DISSOLUTION

At higher total CO₂ content (ΣCO₂) seawater in general shows a lower concentration [CO₃²⁻] of the carbonate ion. As a result the deep and bottom waters of the ocean basins are undersaturated versus crystalline calcite and below the Calciumcarbonate Compensation Depth (CCD) the seafloor is devoid of calcareous sediments. Uptake of fossil fuel CO₂ in the ocean will enhance dissolution of calcite, i.e. the CCD horizon would shoal. The overall reaction may be written as

and the fossil fuel CO₂ will eventually be buffered, i.e. permanently stored. The rate of dissolution relative to the rate of fossil fuel emission is crucial. In the natural ocean the time scale of calcite dissolution is poorly known and may range from 10³ to 10⁵ years (Sundquist, 1986). At ~10⁴ years (Sundquist, 1988) there will be virtually no impact towards smoothing the predicted transient peak in the atmosphere (De Baar and Stoll, 1989).

Deep ocean injection of CO₂ may affect the dissolution rate and at time constants less than ~10³ years significant buffering, i.e. permanent storage, of CO₂ may occur (Wilson, 1992). On the one hand laboratory studies of well-mixed suspensions show rate dependency on the fourth power of the decrease in [CO₃²⁻] ion concentration (Keir, 1980). On the other hand dissolution of in situ sediments is diffusion limited, where upon CO₂ disposal the (eddy) diffusion would likely diminish as bioturbation would stop once benthic fauna disappears due to the low pH of the CO₂-rich effluent. Intense vertical mixing at the benthic boundary would favor rapid buffering by calcite dissolution, this in contrast with the above search for a stratified regime (no vertical mixing) required for preserving the dense CO₂-rich phases (a,b,c,d) versus overlying seawater. There is another paradox, the conservative injection model (without sediment dissolution; Hoffert *et al.*, 1979) would favor disposal at great depth, yet this generally is below the CCD horizon such that there is no CaCO₃ sediment to react with. Upon better understanding of the actual dissolution rate further non-conservative modeling might find disposal at intermediate depth (where existing CaCO₃-sediments are available for dissolution) to be favorable. With or without deep CO₂-injection, the calcite dissolution rate needs to be known more accurately for predicting the rate at which the oceans will buffer the fossil fuel CO₂.

SUMMARY

Large uncertainty remains about the likelihood of global warming, but the CO₂ increase as such already warrants implementation of curtailing measures. Reduction of fossil fuel CO₂ emissions is most imminent and will have to be achieved primarily by realistic options of both energy conservation as well as shifting to other energy sources such as biomass (De Baar, 1990; Hall *et al.*, 1991). Further research into enhancing oceanic CO₂ storage would appear justified only for developing an emergency fallback option for the future in case the realistic options have not been adequately implemented. Deep sea injection requires ~30-45% of the energy produced (DeBaar and Stoll, 1989), is applicable only to point sources representing ~30% of all fossil fuel emissions, also assuming all these plants will be relocated to the seashore. Eventually the injected CO₂ may either re-equilibrate with the atmosphere over 50-500 years, or taken up more permanently when assuming buffering by rapid dissolution of calcite sediments (Wilson, 1992).

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