

small fraction of the levels in 1990 (IPCC, 2000). The declines in Asia and Africa are driven by the depletion of forests, while trends in the Americas have the highest uncertainty given the extent of the forest resource.

Afforestation: Recent (since 1970) afforestation and reforestation as direct human-induced activities have not yet had much impact on the global terrestrial carbon sink. However, regional sinks have been created in areas such as China, where afforestation since the 1970s has sequestered 0.45 GtC (Fang et al., 2001). The largest effect of afforestation is not immediate but through its legacy.

Agricultural practices: Improvement of agricultural practices on carbon-depleted soils has created a carbon sink. For instance, the introduction of conservation tillage in the USA is estimated to have increased soil organic matter (SOM) stocks by about 1.4 GtC over the last 30 years. However, yearly increases in SOM can be sustained only for 50 to 100 years, after which the system reaches a new equilibrium (Cole et al., 1996; Smith et al., 1997). Moreover, modern conservation tillage often entails large inputs of chemicals and fertilizer, which are made using fossil fuels, reducing the CO₂ benefit from carbon sequestration in agricultural soils. The increase in soil carbon stocks under low-tillage systems may also be mostly a topsoil effect with little increase in total profile carbon storage observed, confounded by the fact that most studies of low-tillage systems have only sampled the uppermost soil layers.

7.3.3.1.6 Forest regrowth

Some studies suggest that forest regrowth could be a major contributor to the global land carbon sink (e.g., Pacala et al., 2001; Schimel et al., 2001; Hurtt et al., 2002). Forest areas generally increased during the 20th century at middle and high latitudes (unlike in the tropics). This surprising trend reflects the intensification of agriculture and forestry. Globally, more food is being grown on less land, reflecting mechanisation of agriculture, increased fertilizer use and adoption of high-yield cultivars, although in parts of Africa and Asia the opposite is occurring. Likewise, intensive forest management and agroforestry produce more fibre on less land; improved forest management favours more rapid regrowth of forests after harvest. These trends have led to carbon sequestration by regrowing forests. It should be noted, however, that industrialised agriculture and forestry require high inputs of fossil energy, so it is difficult to assess the net global effects of agricultural intensification on atmospheric greenhouse gases and radiative forcing.

Regional studies have confirmed the plausibility of strong mid-latitude sinks due to forest regrowth. Data from the eddy flux tower network show that forests on long-abandoned former agricultural lands (Curtis et al., 2002) and in industrial managed forests (Hollinger et al., 2002) take up significant amounts of carbon every year. Analysis of forest inventory data shows that, in aggregate, current forest lands are significant sinks for atmospheric CO₂ (Pacala et al., 2001). Few old growth forests remain at mid-latitudes (most forests are less than 70 years old), in part due to forest management. Therefore, forests in

these areas are accumulating biomass because of their ages and stages of succession. Within wide error bands (see Section 7.3.2.3), the uptake rates inferred from flux towers are generally consistent with those inferred from inverse methods (e.g., Hurtt et al., 2002). Stocks of soil carbon are also likely increasing due to replenishment of soil organic matter and necromass depleted during the agricultural phase, and changes in soil microclimate associated with reforestation; these effects might add 30 to 50% to the quantity of CO₂ sequestered (e.g., Barford et al., 2001). It is important to note that at least some of this sequestration is ‘refilling’ the deficits in biomass and soil organic matter, accumulated in previous epochs (see Figure 7.3), and the associated CO₂ uptake should be expected to decline in the coming decades unless sustained by careful management strategies designed to accomplish that purpose.

7.3.4 Ocean Carbon Cycle Processes and Feedbacks to Climate

7.3.4.1 Overview of the Ocean Carbon Cycle

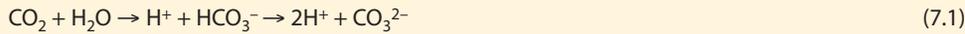
Oceanic carbon exists in several forms: as DIC, DOC, and particulate organic carbon (POC) (living and dead) in an approximate ratio DIC:DOC:POC = 2000:38:1 (about 37,000 GtC DIC: Falkowski et al., 2000 and Sarmiento and Gruber, 2006; 685 GtC DOC: Hansell and Carlson, 1998; and 13 to 23 GtC POC: Eglinton and Repeta, 2004). Before the industrial revolution, the ocean contained about 60 times as much carbon as the atmosphere and 20 times as much carbon as the terrestrial biosphere/soil compartment.

Seawater can, through inorganic processes, absorb large amounts of CO₂ from the atmosphere, because CO₂ is a weakly acidic gas and the minerals dissolved in the ocean have over geologic time created a slightly alkaline ocean (surface pH 7.9 to 8.25: Degens et al., 1984; Royal Society, 2005). The air-sea exchange of CO₂ is determined largely by the air-sea gradient in pCO₂ between atmosphere and ocean. Equilibration of surface ocean and atmosphere occurs on a time scale of roughly one year. Gas exchange rates increase with wind speed (Wanninkhof and McGillis, 1999; Nightingale et al., 2000) and depend on other factors such as precipitation, heat flux, sea ice and surfactants. The magnitudes and uncertainties in local gas exchange rates are maximal at high wind speeds. In contrast, the equilibrium values for partitioning of CO₂ between air and seawater and associated seawater pH values are well established (Zeebe and Wolf-Gladrow, 2001; see Box 7.3).

In addition to changes in advection and mixing, the ocean can alter atmospheric CO₂ concentration through three mechanisms (Volk and Hoffert, 1985), illustrated in Figure 7.10: (1) absorption or release of CO₂ due to changes in solubility of gaseous CO₂ (‘solubility pump’); (2) changes in carbon fixation to POC in surface waters by photosynthesis and export of this carbon through sinking of organic particles out of the surface layer (‘organic carbon pump’) – this process is limited to first order by availability of light and nutrients (phosphate, nitrate, silicic acid and micronutrients such as iron); and (3) changes in

Box 7.3: Marine Carbon Chemistry and Ocean Acidification

The marine carbonate buffer system allows the ocean to take up CO_2 far in excess of its potential uptake capacity based on solubility alone, and in doing so controls the pH of the ocean. This control is achieved by a series of reactions that transform carbon added as CO_2 into HCO_3^- and CO_3^{2-} . These three dissolved forms (collectively known as DIC) are found in the approximate ratio $\text{CO}_2:\text{HCO}_3^-:\text{CO}_3^{2-}$ of 1:100:10 (Equation (7.1)). CO_2 is a weak acid and when it dissolves, it reacts with water to form carbonic acid, which dissociates into a hydrogen ion (H^+) and a HCO_3^- ion, with some of the H^+ then reacting with CO_3^{2-} to form a second HCO_3^- ion (Equation (7.2)).



Therefore, the net result of adding CO_2 to seawater is an increase in H^+ and HCO_3^- , but a reduction in CO_3^{2-} . The decrease in the CO_3^{2-} ion reduces the overall buffering capacity as CO_2 increases, with the result that proportionally more H^+ ions remain in solution and increase acidity.

This ocean acidification is leading to a decrease in the saturation state of CaCO_3 in the ocean. Two primary effects are expected: (1) the biological production of corals as well as calcifying phytoplankton and zooplankton within the water column may be inhibited or slowed down (Royal Society, 2005), and (2) the dissolution of CaCO_3 at the ocean floor will be enhanced (Archer, 2005). Aragonite, the meta-stable form of CaCO_3 produced by corals and pteropods (planktonic snails; Lalli and Gilmer, 1989), will be particularly susceptible to a pH reduction (Kleypas et al., 1999b; Hughes et al., 2003; Orr et al., 2005). Laboratory experiments under high ambient CO_2 with the coccolithophore species *Emiliania huxleyi* and *Gephyrocapsa oceanica* produce a significant reduction in CaCO_3 production and a stimulation of POC production (Riebesell et al., 2000; Zondervan et al., 2001). Other species and growth under other conditions may show different responses, so that no conclusive quantification of the CaCO_3 feedback is possible at present (Tortell et al., 2002; Sciandra et al., 2003).

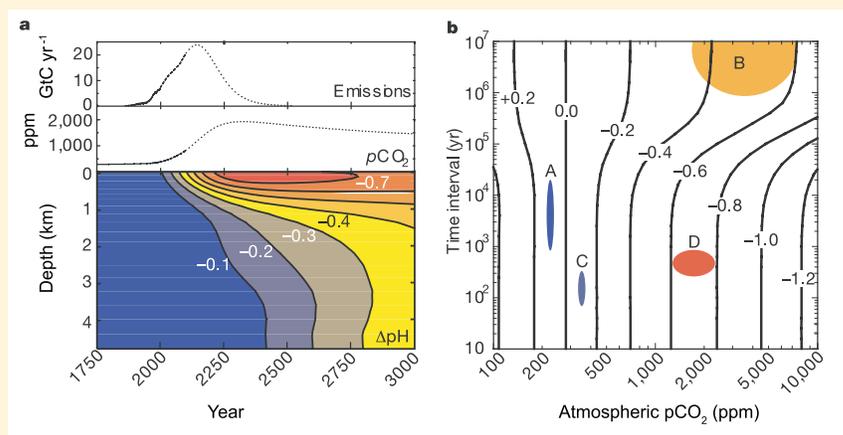
The sinking speed of marine particle aggregates depends on their composition: CaCO_3 may act as an efficient ballast component, leading to high sinking speeds of aggregates (Armstrong et al., 2002; Klaas and Archer, 2002). The relatively small negative feedback of reduced CaCO_3 production to atmospheric pCO_2 may be compensated for by a change in the ballast for settling biogenic particles and the associated shallowing of re-mineralization depth levels in the water column for organic carbon (Heinze, 2004). On the other hand, production of extracellular organic carbon could increase under high CO_2 levels and lead to an increase in export (Engel et al., 2004).

Ecological changes due to expected ocean acidification may be severe for corals in tropical and cold waters (Gattuso et al., 1999; Kleypas et al., 1999a; Langdon et al., 2003; Buddemeier et al., 2004; Roberts et al., 2006) and for pelagic ecosystems (Tortell et al., 2002; Royal Society, 2005). Acidification can influence the marine food web at higher trophic levels (Langenbuch and Pörtner, 2003; Ishimatsu et al., 2004).

Since the beginning of the industrial revolution, sea surface pH has dropped by about 0.1 pH units (corresponding to a 30% increase in the H ion concentration). The expected continued decrease may lead within a few centuries to an ocean pH estimated to have occurred most recently a few hundred million years before present (Caldeira and Wickett, 2003; Key et al., 2004; Box 7.3, Figure 1).

According to a model experiment based on the IPCC Scenarios 1992a (IS92a) emission scenario, bio-calcification will be reduced by 2100, in particular within the Southern Ocean (Orr et al., 2005), and by 2050 for aragonite-producing organisms (see also Figure 10.24). It is important to note that ocean acidification is not a direct consequence of climate change but a consequence of fossil fuel CO_2 emissions, which are the main driver of the anticipated climate change.

Box 7.3, Figure 1. (a) Atmospheric CO_2 emissions, historical atmospheric CO_2 levels and predicted CO_2 concentrations from the given emission time series, together with changes in ocean pH based on horizontally averaged chemistry. The emission time series is based on the mid-range IS92a emission scenario (solid line) prior to 2100 and then assumes that emissions continue until fossil fuel reserves decline. (b) Estimated maximum change in surface ocean pH as a function of final atmospheric CO_2 pressure, and the transition time over which this CO_2 pressure is linearly approached from 280 ppm. A: Glacial-interglacial CO_2 changes; B: slow changes over the past 300 Myr; C: historical changes in ocean surface waters; D: unabated fossil fuel burning over the next few centuries. Source: Caldeira and Wickett (2003). Reprinted with permission from Macmillan Publishers Ltd: Nature, Caldeira and Wickett (2003), copyright (2003).



the release of CO_2 in surface waters during formation of CaCO_3 shell material by plankton (' CaCO_3 counter pump').

Organic particles are re-mineralized (oxidized to DIC and other inorganic compounds through the action of bacteria) primarily in the upper 1,000 m of the oceanic water column, with an accompanying decrease in dissolved O. On the average, CaCO_3 particles sink deeper before they undergo dissolution: deep waters are undersaturated with respect to CaCO_3 . The remainder of the particle flux enters marine sediments and is subject to either re-dissolution within the water column or accumulation within the sediments. Although the POC reservoir is small, it plays an important role in keeping DIC concentrations low in surface waters and high in deep waters. The loop is closed through the three-dimensional ocean circulation: upwelling water brings inorganic carbon and nutrients to the surface again, leading to outgassing and biogenic particle production. Dissolved organic carbon enters the ocean water column from rivers and marine metabolic processes. A large fraction of DOC has a long ocean residence time (1–10 kyr), while other fractions are more short-lived (days to hundreds of years; Loh

et al., 2004). The composition of dissolved organic matter is still largely unknown.

In conjunction with the global ocean mixing or overturning time of the order of 1 kyr (Broecker and Peng, 1982), small changes in the large ocean carbon reservoir can induce significant changes in atmospheric CO_2 concentration. Likewise, perturbations in the atmospheric pCO_2 can be buffered by the ocean. Glacial-interglacial changes in the atmospheric CO_2 content can potentially be attributed to a change in functioning of the marine carbon pump (see Chapter 6). The key role for the timing of the anthropogenic carbon uptake by the ocean is played by the downward transport of surface water, with a high burden of anthropogenic carbon, into the ocean's interior. The organic carbon cycle and the CaCO_3 counter pump modulate, but do not dominate, the net marine uptake of anthropogenic carbon.

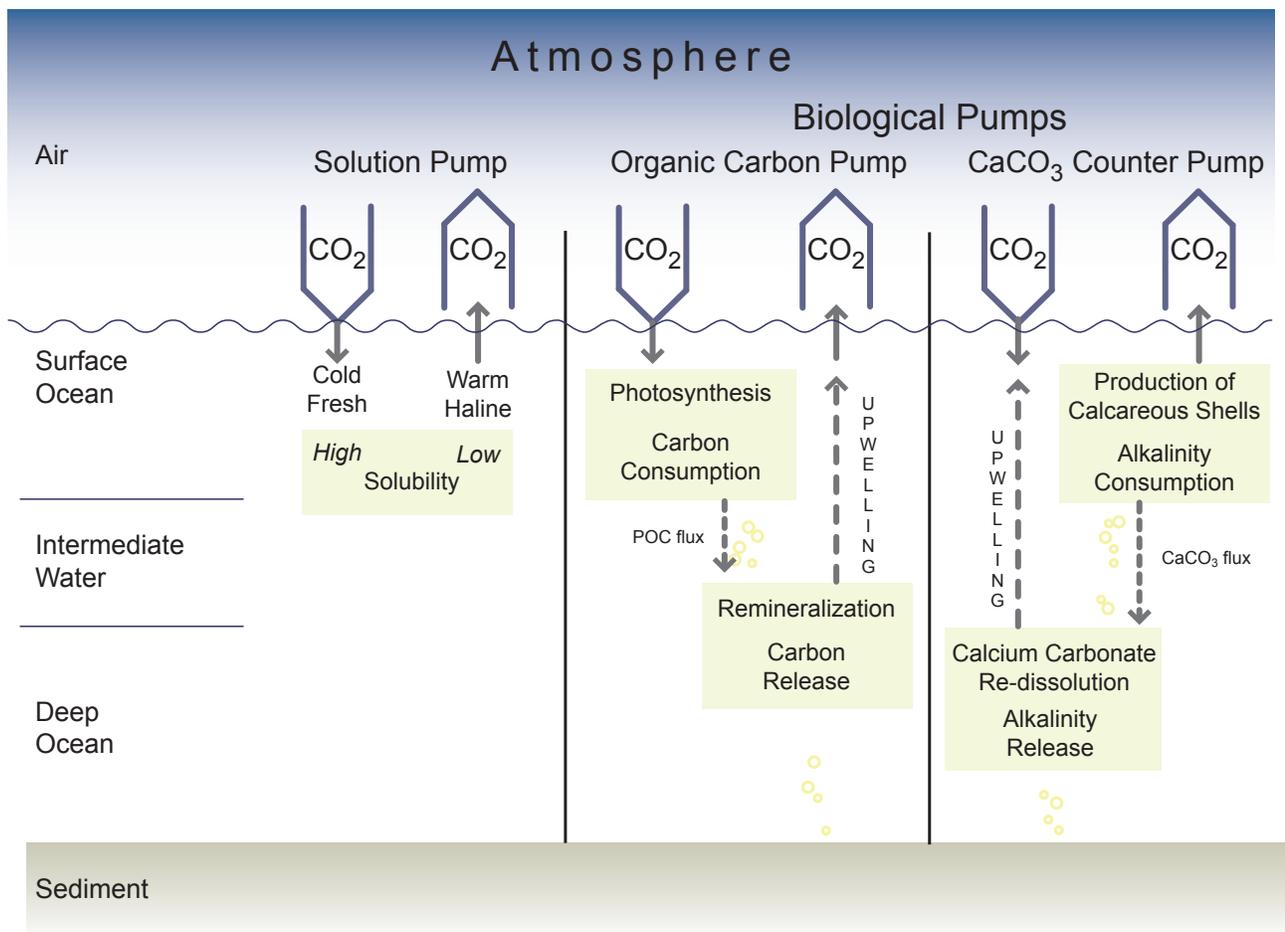


Figure 7.10. Three main ocean carbon pumps govern the regulation of natural atmospheric CO_2 changes by the ocean (Heinze et al., 1991): the solubility pump, the organic carbon pump and the CaCO_3 'counter pump'. The oceanic uptake of anthropogenic CO_2 is dominated by inorganic carbon uptake at the ocean surface and physical transport of anthropogenic carbon from the surface to deeper layers. For a constant ocean circulation, to first order, the biological carbon pumps remain unaffected because nutrient cycling does not change. If the ocean circulation slows down, anthropogenic carbon uptake is dominated by inorganic buffering and physical transport as before, but the marine particle flux can reach greater depths if its sinking speed does not change, leading to a biologically induced negative feedback that is expected to be smaller than the positive feedback associated with a slower physical downward mixing of anthropogenic carbon. Reprinted with permission, copyright 1991 American Geophysical Union.

7.3.4.2 Carbon Cycle Feedbacks to Changes in Atmospheric Carbon Dioxide

Chemical buffering of anthropogenic CO₂ is the quantitatively most important oceanic process acting as a carbon sink. Carbon dioxide entering the ocean is buffered due to scavenging by the CO₃²⁻ ions and conversion to HCO₃⁻, that is, the resulting increase in gaseous seawater CO₂ concentration is smaller than the amount of CO₂ added per unit of seawater volume. Carbon dioxide buffering in seawater is quantified by the Revelle factor ('buffer factor', Equation (7.3)), relating the fractional change in seawater pCO₂ to the fractional change in total DIC after re-equilibration (Revelle and Suess, 1957; Zeebe and Wolf-Gladrow, 2001):

$$\text{Revelle factor (or buffer factor)} = \frac{(\Delta[\text{CO}_2] / [\text{CO}_2])}{(\Delta[\text{DIC}] / [\text{DIC}])} \quad (7.3)$$

The lower the Revelle factor, the larger the buffer capacity of seawater. Variability of the buffer factor in the ocean depends mainly on changes in pCO₂ and the ratio of DIC to total alkalinity. In the present-day ocean, the buffer factor varies between 8 and 13 (Sabine et al., 2004a; Figure 7.11). With respect to atmospheric pCO₂ alone, the inorganic carbon system of the ocean reacts in two ways: (1) seawater re-equilibrates, buffering a significant amount of CO₂ from the atmosphere depending on the water volume exposed to equilibration; and (2) the Revelle factor increases with pCO₂ (positive feedback; Figure 7.11). Both processes are quantitatively important. While the first is generally considered as a system response, the latter is a feedback process.

The ocean will become less alkaline (seawater pH will decrease) due to CO₂ uptake from the atmosphere (see Box 7.3).

The ocean's capacity to buffer increasing atmospheric CO₂ will decline in the future as ocean surface pCO₂ increases (Figure 7.11a). This anticipated change is certain, with potentially severe consequences.

Increased carbon storage in the deep ocean leads to the dissolution of calcareous sediments below their saturation depth (Broecker and Takahashi, 1978; Feely et al., 2004). The feedback of CaCO₃ sediment dissolution to atmospheric pCO₂ increase is negative and quantitatively significant on a 1 to 100 kyr time scale, where CaCO₃ dissolution will account for a 60 to 70% absorption of the anthropogenic CO₂ emissions, while the ocean water column will account for 22 to 33% on a time scale of 0.1 to 1 kyr. In addition, the remaining 7 to 8% may be compensated by long-term terrestrial weathering cycles involving silicate carbonates (Archer et al., 1998). Due to the slow CaCO₃ buffering mechanism (and the slow silicate weathering), atmospheric pCO₂ will approach a new equilibrium asymptotically only after several tens of thousands of years (Archer, 2005; Figure 7.12).

Elevated ambient CO₂ levels appear to also influence the production rate of POC by marine calcifying planktonic organisms (e.g., Zondervan et al., 2001). This increased carbon fixation under higher CO₂ levels was also observed for three diatom (siliceous phytoplankton) species (Riebesell et al., 1993). It is critical to know whether these increased carbon fixation rates translate into increased export production rates (i.e., removal of carbon to greater depths). Studies of the nutrient to carbon ratio in marine phytoplankton have not yet shown any significant changes related to CO₂ concentration of the nutrient utilisation efficiency (expressed through the 'Redfield ratio' – carbon:nitrogen:phosphorus:silicon) in organic tissue (Burkhardt et al., 1999).

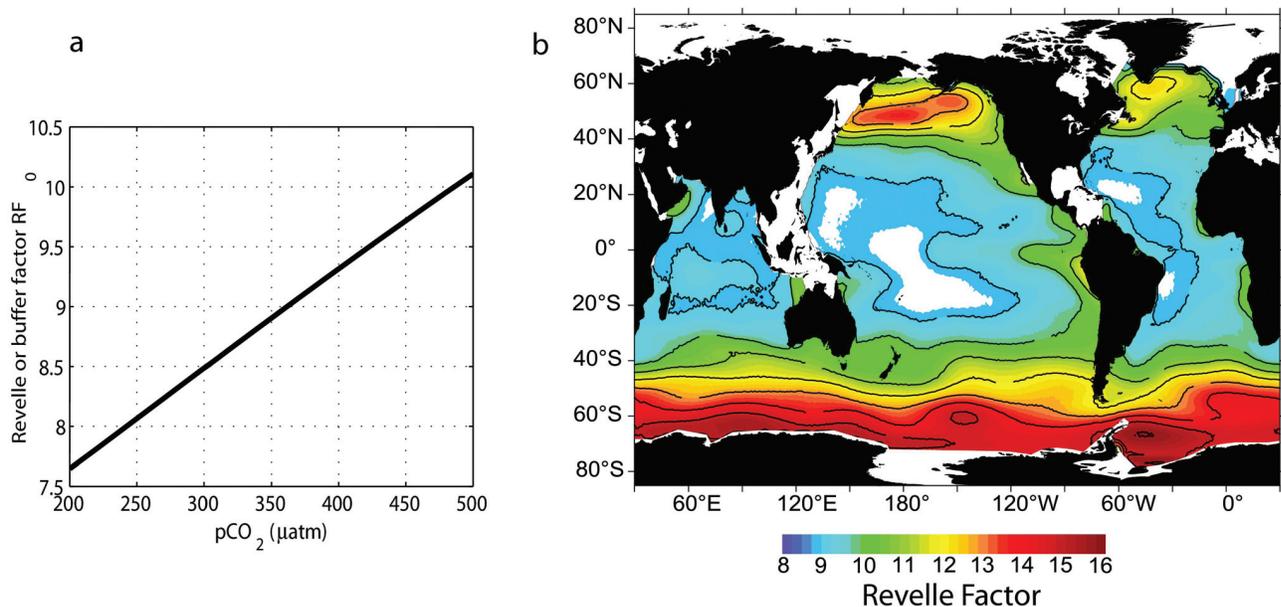


Figure 7.11. (a) The Revelle factor (or buffer factor) as a function of CO₂ partial pressure (for temperature 25°C, salinity 35 psu, and total alkalinity 2,300 μmol kg⁻¹) (Zeebe and Wolf-Gladrow, 2001, page 73; reprinted with permission, copyright 2001 Elsevier). (b) The geographical distribution of the buffer factor in ocean surface waters in 1994 (Sabine et al., 2004a; reprinted with permission, copyright 2004 American Association for the Advancement of Science). High values indicate a low buffer capacity of the surface waters.

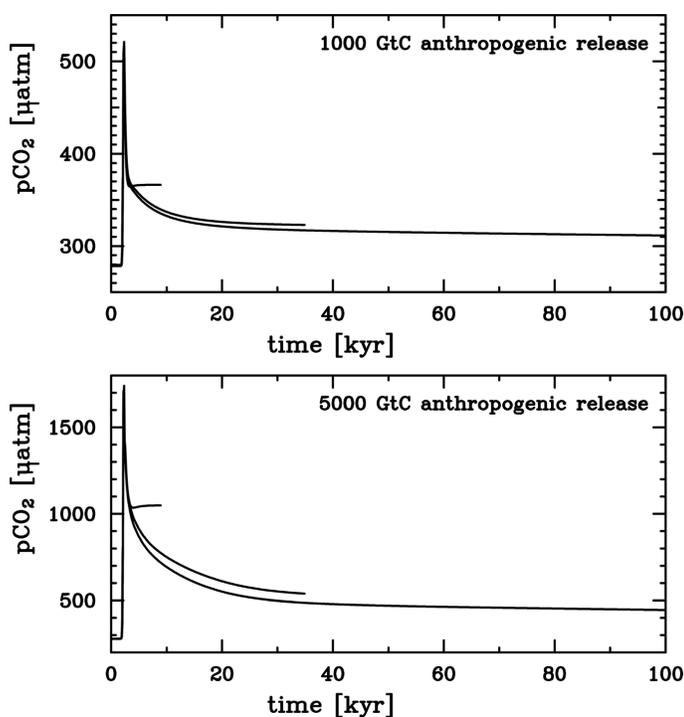


Figure 7.12. Model projections of the neutralization of anthropogenic CO_2 for an ocean-only model, a model including dissolution of CaCO_3 sediment and a model including weathering of silicate rocks, (top) for a total of 1,000 GtC of anthropogenic CO_2 emissions and (bottom) for a total of 5,000 GtC of anthropogenic CO_2 . Note that the y-axis is different for the two diagrams. Without CaCO_3 dissolution from the seafloor, the buffering of anthropogenic CO_2 is limited. Even after 100 kyr, the remaining pCO_2 is substantially higher than the pre-industrial value. Source: Archer (2005).

7.3.4.3 Carbon Cycle Feedbacks to Changes in Physical Forcing

A more sluggish ocean circulation and increased density stratification, both expected in a warmer climate, would slow down the vertical transport of carbon, alkalinity and nutrients, and the replenishment of the ocean surface with water that has not yet been in contact with anthropogenic CO_2 . This narrowing of the ‘bottleneck’ for anthropogenic CO_2 invasion into the ocean would provide a significant positive feedback to atmospheric greenhouse gas concentrations (Bolin and Eriksson, 1959; see also the carbon cycle climate model simulations by Cox et al., 2000; Friedlingstein et al., 2001, 2006). As long as the vertical transfer rates for marine biogenic particles remain unchanged, in a more sluggish ocean the biological carbon pump will be more efficient (Boyle, 1988; Heinze et al., 1991), thus inducing a negative feedback, which is expected to be smaller than the physical transport feedback (Broecker, 1991; Maier-Reimer et al., 1996; Plattner et al., 2001; see Figure 7.10). However, a modelling study by Bopp et al. (2005) predicts a decrease in vertical particle transfer and hence shallower depths of re-mineralization of particulate organic carbon resulting in a positive CO_2 feedback. Further changes in plankton community structure including the role of N_2 -fixing organisms can feed

back to the carbon cycle (Sarmiento et al., 2004; Mahaffey et al., 2005). Changes in ocean circulation can affect the regional circulation of shelf and coastal seas, leading either to increased export of nutrients plus carbon from the shallow seas into the open ocean or to increased upwelling of nutrients plus carbon onto the shelf and towards coastal areas (Walsh, 1991; Smith and Hollibaugh, 1993; Chen et al., 2003; Borges, 2005). A reduction in sea ice cover may increase the uptake area for anthropogenic CO_2 and act as a minor negative carbon feedback (ACIA, 2005). The physical ‘bottleneck’ feedback dominates over biological feedbacks induced by circulation change, resulting in an anticipated overall positive feedback to climate change. Both feedbacks depend on details of the future ocean circulation and model projections show a large range.

The solubility of CO_2 gas in seawater and the two dissociation constants of carbonic acid in seawater depend on temperature and salinity (Weiss, 1974; Millero et al., 2002). A 1°C increase in sea surface temperature produces an increase in pCO_2 of 6.9 to 10.2 ppm after 100 to 1,000 years (Heinze et al., 2003; see also Broecker and Peng, 1986; Plattner et al., 2001). Warming may increase the biological uptake rate of nutrients and carbon from surface waters, but the net effect on export and DIC is uncertain. Laws et al. (2000) proposed that export efficiency increases with net photosynthesis at low temperatures, which implies a positive feedback to warming. In addition, DOC may be degraded more quickly at higher temperatures.

7.3.4.4 Carbon Cycle Feedbacks Induced by Nutrient Cycling and Land Ocean Coupling

Rivers deliver carbon (DIC, DOC) and nutrients to the ocean. Rising CO_2 levels in the atmosphere and land use may lead to increased chemical and physical weathering, resulting in increased carbon and alkalinity loads in rivers (Clair et al., 1999; Hejzlar et al., 2003; Raymond and Cole, 2003; Freeman et al., 2004). Depending on the lithology and soil composition of the catchment areas, increased levels of alkalinity, DIC or DOC can lead to local positive or negative feedbacks. Mobilisation of silicate carbonates from soils and transfer to the ocean would lead to a negative feedback to atmospheric CO_2 on long time scales (Dupre et al., 2003). Variations in nutrient supply can lead to species shifts and to deviations from the large-scale average Redfield ratios mainly in coastal waters, but also in the open ocean (Pahlow and Riebesell, 2000). Nutrient supply to the ocean has been changed through increased nitrate release from land due to fertilizer use as well as nitrogen deposition from the atmosphere in highly polluted areas (De Leeuw et al., 2001; Green et al., 2004).

Dust deposition to the ocean provides an important source of micronutrients (iron, zinc and others, e.g., Frew et al., 2001; Boyd et al., 2004) and ballast material to the ocean. Areas where iron is not supplied by aeolian dust transport in sufficient amounts tend to be iron-limited. A warmer climate may result on the average in a decrease of dust mobilisation and transport (Werner et al., 2002; Mahowald and Luo, 2003) although increased dust loads may result as well due to changes in land

use (Tegen et al., 2004) and in vegetation cover (Woodward et al., 2005). A decrease in dust loads could result in a net positive feedback, further increasing CO₂ through a weakening of marine biological production and export of aggregates due to clay ballast (Haake and Ittekkot, 1990; Ittekkot, 1993). Changes in plankton species composition and regional shifts of high production zones due to a changing climate could lead to a series of further feedbacks. Light absorption due to changes in bio-optical heating may change and induce a respective temperature change in ocean surface water (Sathyendranath et al., 1991; Wetzel et al., 2006). An increase in blooms involving calcifying organisms as indicated for the high northern latitudes (Broerse et al., 2003; Smyth et al., 2004) can temporarily increase surface ocean albedo, though the effect on the radiation budget is small (Tyrell et al., 1999).

7.3.4.5 Summary of Marine Carbon Cycle Climate Couplings

Couplings between the marine carbon cycle and climate are summarised in Table 7.3 and below.

7.3.4.5.1 Robust findings

- A potential slowing down of the ocean circulation and the decrease of seawater buffering with rising CO₂ concentration will suppress oceanic uptake of anthropogenic CO₂.
- Ocean CO₂ uptake has lowered the average ocean pH (increased acidity) by approximately 0.1 since 1750. Ocean acidification will continue and is directly and inescapably coupled to the uptake of anthropogenic CO₂ by the ocean.
- Inorganic chemical buffering and dissolution of marine CaCO₃ sediments are the main oceanic processes for neutralizing anthropogenic CO₂. These processes cannot prevent a temporary buildup of a large atmospheric CO₂ pool because of the slow large-scale overturning circulation.

7.3.4.5.2 Key uncertainties

- Future changes in ocean circulation and density stratification are still highly uncertain. Both the physical uptake of CO₂ by the ocean and changes in the biological cycling of carbon depend on these factors.
- The overall reaction of marine biological carbon cycling (including processes such as nutrient cycling as well as ecosystem changes including the role of bacteria and viruses) to a warm and high-CO₂ world is not yet well understood. Several small feedback mechanisms may add up to a significant one.
- The response of marine biota to ocean acidification is not yet clear, both for the physiology of individual organisms and for ecosystem functioning as a whole. Potential impacts are expected especially for organisms that build CaCO₃ shell material ('bio-calcification'). Extinction thresholds will likely be crossed for some organisms in some regions in the coming century.

7.3.5 Coupling Between the Carbon Cycle and Climate

7.3.5.1 Introduction

Atmospheric CO₂ is increasing at only about half the rate implied by fossil fuel plus land use emissions, with the remainder being taken up by the ocean, and vegetation and soil on land. Therefore, the land and ocean carbon cycles are currently helping to mitigate CO₂-induced climate change. However, these carbon cycle processes are also sensitive to climate. The glacial-interglacial cycles are an example of tight coupling between climate and the carbon cycle over long time scales, but there is also clear evidence of the carbon cycle responding to short-term climatic anomalies such as the El Niño-Southern Oscillation (ENSO) and Arctic Oscillation (Rayner et al., 1999; Bousquet et al., 2000; C. Jones et al., 2001; Lintner, 2002; Russell and Wallace, 2004) and the climate perturbation arising from the Mt. Pinatubo volcanic eruption (Jones and Cox, 2001a; Lucht et al., 2002; Angert et al., 2004).

Previous IPCC reports have used simplified or 'reduced-form' models to estimate the impact of climate change on the carbon cycle. However, detailed climate projections carried out with Atmosphere-Ocean General Circulation Models (AOGCMs) have typically used a prescribed CO₂ concentration scenario, neglecting two-way coupling between climate and the carbon cycle. This section discusses the first generation of coupled climate-carbon cycle AOGCM simulations, using the results to highlight a number of critical issues in the interaction between climate change and the carbon cycle.

7.3.5.2 Coupled Climate-Carbon Cycle Projections

The TAR reported two initial climate projections using AOGCMs with interactive carbon cycles. Both indicated positive feedback due largely to the impacts of climate warming on land carbon storage (Cox et al., 2000; Friedlingstein et al., 2001), but the magnitude of the feedback varied markedly between the models (Friedlingstein et al., 2003). Since the TAR a number of other climate modelling groups have completed climate-carbon cycle projections (Brovkin et al., 2004; Thompson et al., 2004; N. Zeng et al., 2004; Fung et al., 2005; Kawamiya et al., 2005; Matthews et al., 2005; Sitch et al., 2005) as part of C⁴MIP. The 11 models involved in C⁴MIP differ in the complexity of their components (Friedlingstein et al., 2006), including both Earth System Models of Intermediate Complexity and AOGCMs.

The models were forced by historical and Special Report on Emission Scenarios (SRES; IPCC, 2000) A2 anthropogenic CO₂ emissions for the 1850 to 2100 time period. Each modelling group carried out at least two simulations: one 'coupled' in which climate change affects the carbon cycle, and one 'uncoupled' in which atmospheric CO₂ increases do not influence climate (so that the carbon cycle experiences no CO₂-induced climate change). A comparison of the runs defines the climate-carbon cycle feedback, quantified by the feedback factor: