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Ocean Acidification: The Other CO₂ Problem?

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OCEAN ACIDIFICATION: THE OTHER CO₂ PROBLEM[∇]

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KEY WORDS: biogeochemistry, calcification, carbon dioxide, climate change, coral, ecosystem

ABSTRACT: Rising atmospheric carbon dioxide (CO₂), primarily from human fossil fuel combustion, reduces ocean pH and causes wholesale shifts in seawater carbonate chemistry. The process of ocean acidification is well documented in field data, and the rate will accelerate over this century unless future CO₂ emissions are curbed dramatically. Acidification alters seawater chemical speciation and biogeochemical cycles of many elements and compounds. One well-known effect is the lowering of calcium carbonate saturation states, which impacts shell-forming marine organisms from plankton to benthic molluscs, echinoderms, and corals. Many calcifying species exhibit reduced calcification and growth rates in laboratory experiments under high-CO₂ conditions. Ocean acidification also causes an increase in carbon fixation rates in some photosynthetic organisms (both calcifying and noncalcifying). The potential for marine organisms to adapt to increasing CO₂ and broader implications for ocean ecosystems are not well known; both are high priorities for future research. Although ocean pH has varied in the geological past, paleo-events may be only imperfect analogs to current conditions.

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INTRODUCTION

Over the past 250 years, atmospheric carbon dioxide (CO₂) levels increased by nearly 40%, from preindustrial levels of approximately 280 ppmv (parts per million volume) to nearly 384 ppmv in 2007 (Solomon et al. 2007). This rate of increase, driven by human fossil fuel combustion and deforestation, is at least an order of magnitude faster than has occurred for millions of years (Doney & Schimel 2007), and the current concentration is higher than experienced on Earth for at least the past 800,000 years (Lüthi et al. 2008). Rising atmospheric CO₂ is tempered by oceanic uptake, which accounts for nearly a third of anthropogenic carbon added to the atmosphere (Sabine & Feely 2007, Sabine et al. 2004), and without which atmospheric CO₂ would be approximately 450 ppmv today, a level of CO₂ that would have led to even greater climate change than witnessed today. Ocean CO₂ uptake, however, is not benign; it causes pH¹ reductions and alterations in fundamental chemical balances that together are commonly referred to as ocean acidification. Because climate change and ocean acidification are both caused by increasing atmospheric CO₂, acidification is commonly referred to as the “other CO₂ problem” (Henderson 2006, Turley 2005).

Ocean acidification² is a predictable consequence of rising atmospheric CO₂ and does not suffer from uncertainties associated with climate change forecasts. Absorption of anthropogenic CO₂, reduced pH, and lower calcium carbonate

1. pH: a measure of ocean acidity and hydrogen ion H⁺ concentration; pH = -log₁₀[H⁺]

2. Ocean acidification: the addition of carbon dioxide in seawater that causes a reduction in ocean pH and shifts in carbonate speciation

(CaCO₃) saturation in surface waters, where the bulk of oceanic production occurs, are well verified from models, hydrographic surveys, and time series data (Caldeira & Wickett 2003, 2005; Feely et al. 2004, 2008; Orr et al. 2005; Solomon et al. 2007). At the Hawaii Ocean Time-Series (HOT) station ALOHA the growth rates of surface water pCO₂ and atmospheric CO₂ agree well³ (Takahashi et al. 2006) (Figure 1), indicating uptake of anthropogenic CO₂ as the major cause for long-term increases in dissolved inorganic carbon (DIC)⁴ and decreases in CaCO₃ saturation state.⁵ Correspondingly, since the 1980s average pH measurements at HOT, the Bermuda Atlantic Time-Series Study, and European Station for Time-Series in the Ocean in the eastern Atlantic have decreased approximately 0.02 units per decade (Solomon et al. 2007). Since preindustrial times, the average ocean surface water pH has fallen by approximately 0.1 units, from approximately 8.21 to 8.10 (Royal Society 2005), and is expected to decrease a further 0.3–0.4 pH units (Orr et al. 2005) if atmospheric CO₂ concentrations reach 800 ppmv [the projected end-of-century concentration according to the Intergovernmental Panel on Climate Change (IPCC) business-as-usual emission scenario].

Fossil fuel combustion and agriculture also produce increased atmospheric inputs of dissociation products of strong acids (HNO₃ and H₂SO₄) and bases (NH₃) to the coastal and open ocean. These inputs are particularly important close to major source regions, primarily in the northern hemisphere, and cause decreases in surface seawater alkalinity, pH, and DIC (Doney et al. 2007). On a global scale, these anthropogenic inputs (0.8 Tmol/yr reactive sulfur and 2.7 Tmol/yr reactive nitrogen) contribute only a small fraction of the acidification caused by anthropogenic CO₂, but they are more concentrated in coastal waters where the ecosystem responses to ocean acidification could be more serious for humankind.

Seawater carbon dioxide measurements have been conducted since the beginning of the nineteenth century

3. pCO₂: carbon dioxide partial pressure

4. Dissolved inorganic carbon (DIC): sum of aqueous CO₂ gas, carbonic acid, bicarbonate, and carbonate ions

5. Saturation state: thermodynamic condition of seawater that describes the degree of supersaturation or undersaturation with respect to the particular phase of the CaCO₃ mineral

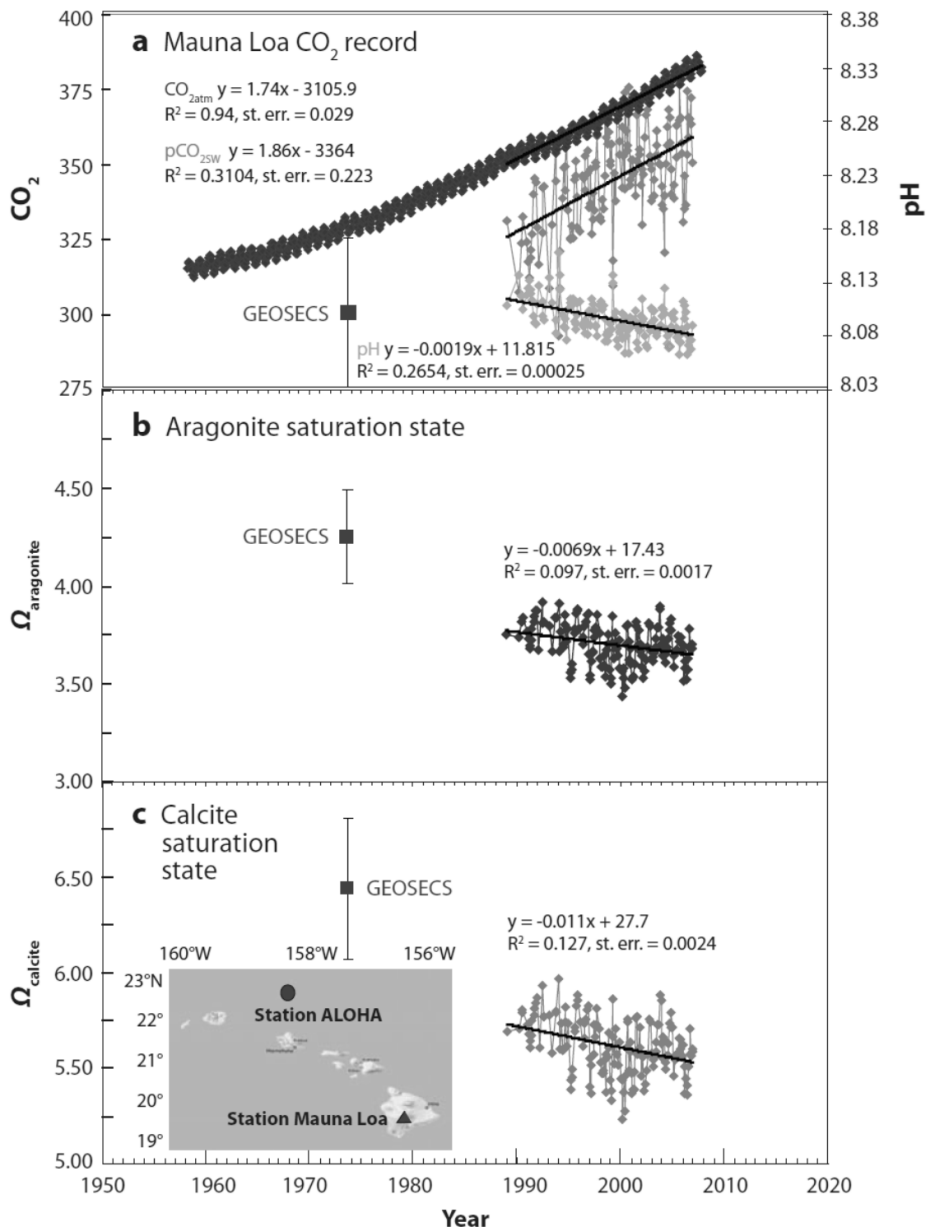
(Krogh 1904) but were sparse until the middle of the twentieth century (Keeling et al. 1965, Takahashi 1961) and particularly until the Geochemical Sections (GEOSECS)⁶ (1973–1979) (Craig & Turekian 1976, 1980) and Transient Tracers in the Ocean (TTO) (1981–1983) (Brewer et al. 1985) programs. Even so, the GEOSECS and TTO measurements were significantly less precise than those of today. Although researchers recognized that the concentration of carbon dioxide in the surface ocean was more or less in equilibrium with overlying atmosphere CO₂, they largely dismissed the potential impact on the ocean biota because calcite (the assumed CaCO₃ mineralogy of most calcifying organisms) would remain supersaturated in the surface ocean.

Since then, multiple studies revealed several issues that elevate ocean acidification as a threat to marine biota: (a) the calcification⁷ rates of many shell-forming organisms respond to the degree of supersaturation (e.g., Smith & Buddemeier 1992, Kleypas et al. 1999); (b) aragonite,⁸ a more soluble CaCO₃ mineral equally important in calcifying organisms, may become undersaturated in the surface ocean within the early 21st century (Feely & Chen 1982, Feely et al. 1988, Orr et al. 2005); and (c) the biological effects of decreasing ocean pH reach far beyond limiting calcification.

6. Geochemical Sections (GEOSECS): a global-scale 1970s chemical oceanography survey

7. Calcification: a biological process that uses dissolved ions to form calcium carbonate minerals for shells and skeletal components

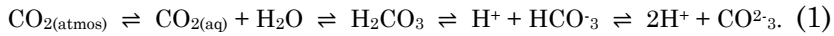
8. Aragonite: a relatively soluble mineral form of calcium carbonate found in corals, pteropods and other molluscs, and a variety of other invertebrates and algae

**Figure 1**

Time series of: (a) atmospheric CO₂ at Mauna Loa (in parts per million volume, ppmv) (red), surface ocean pH (cyan), and pCO₂ (µatm) (tan) at Ocean Station ALOHA in the subtropical North Pacific Ocean; and (b) aragonite saturation (dark blue) and (c) calcite saturation (gray) at Station ALOHA. Note that the increase in oceanic CO₂ over the past 17 years is consistent with the atmospheric increase within the statistical limits of the measurements. Mauna Loa data courtesy of Dr. Pieter Tans, National Oceanic and Atmospheric Administration/Earth System Research Laboratory (<http://www.esrl.noaa.gov/gmd/ccgg/trends>); Hawaii Ocean Time-Series (HOT)/ALOHA data courtesy of Dr. David Karl, University of Hawaii (<http://hahana.soest.hawaii.edu>). Geochemical Ocean Section Study (GEOSECS) data are from a station near Station ALOHA collected in 1973; GEOSECS data from Takahashi et al. (1980).

OCEAN CARBONATE SYSTEM

Seawater carbonate chemistry is governed by a series of chemical reactions:



Air-sea gas exchange equilibrates surface water CO₂ to atmospheric levels with a timescale of approximately one year. Once dissolved in seawater, CO₂ gas reacts with water to form carbonic acid (H₂CO₃), which can then dissociate by losing hydrogen ions to form bicarbonate (HCO₃⁻) and carbonate (CO₃²⁻) ions. The seawater reactions are reversible and near equilibrium (Millero et al. 2002); for surface seawater with pH of ~8.1, approximately 90% of the inorganic carbon is bicarbonate ion, 9% is carbonate ion, and only 1% is dissolved CO₂. Adding CO₂ to seawater increases aqueous CO₂, bicarbonate, and hydrogen ion concentrations; the latter lowers pH because $\text{pH} = -\log_{10}[\text{H}^+]$. Carbonate ion concentration declines, however, because of the increasing H⁺ concentrations. The projected 0.3–0.4 pH drop for the 21st century is equivalent to approximately a 150% increase in H⁺ and 50% decrease in CO₃²⁻ concentrations (Orr et al. 2005).

Over century and longer timescales, the ocean's ability to absorb atmospheric CO₂ depends on the extent of CaCO₃ dissolution in the water column or sediments:



The mineral CaCO₃ derives from shells and skeletons of marine organisms, including plankton, corals and coralline algae, and many other invertebrates. In pelagic environments, carbonates fall through the water column and are either dissolved or deposited in shallow or deep-sea sediments (Berelson et al. 2007, Feely et al. 2004). CaCO₃ formation and dissolution rates vary with saturation state (Ω), defined as the ion product of calcium and carbonate ion concentrations:

$$\Omega = [\text{Ca}^{2+}][\text{CO}_3^{2-}] / K_{\text{sp}}. \quad (3)$$

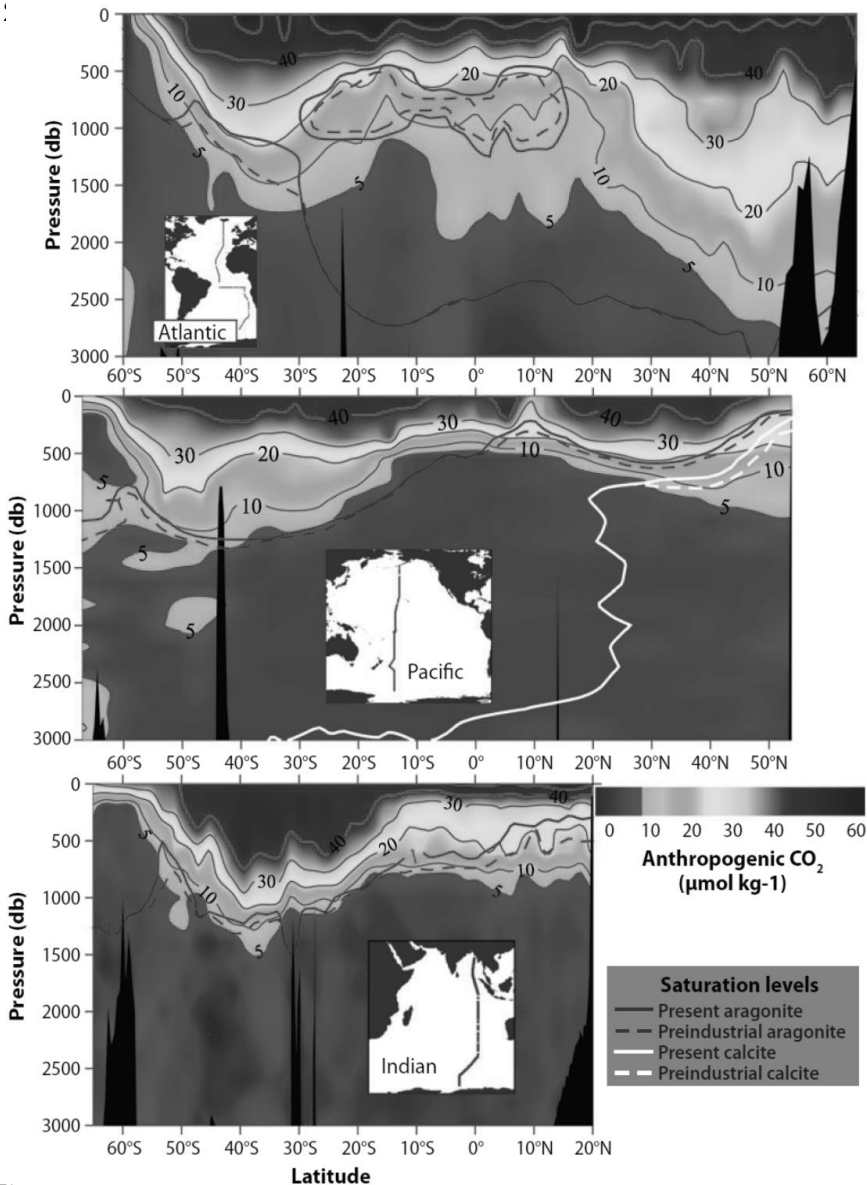


Figure 2

Vertical distributions of anthropogenic CO₂ concentrations in μmol kg⁻¹ and the saturation state $\Omega = 1.0$ horizons for aragonite (red) and calcite (white) for present (solid line) and preindustrial (dashed line) conditions along north-south transects in the (a) Atlantic, (b) Pacific, and (c) Indian Oceans as in Feely et al. (2004). Adapted with permission from AAAS.

The apparent solubility product K'_{sp} depends on temperature, salinity, pressure, and the particular mineral phase; aragonite is approximately 50% more soluble than calcite⁹ (Mucci 1983). Because $[Ca^{2+}]$ is closely proportional to

9. Calcite: a mineral form of calcium carbonate found in many marine plankton and

salinity, Ω is largely determined by variations in $[\text{CO}_3^{2-}]$, which can be calculated from DIC and total alkalinity data.¹⁰ Shell and skeleton formation generally occurs where $\Omega > 1.0$ and dissolution occurs where $\Omega < 1.0$ (unless the shells or skeletons are protected, for example, by organic coatings).

Saturation states are highest in shallow, warm tropical waters and lowest in cold high-latitude regions and at depth, which reflects the increase in CaCO_3 solubility with decreasing temperature and increasing pressure (Feely et al. 2004) (Figure 2). The aragonite and calcite saturation horizons ($\Omega = 1$) are shallower in the Indian and Pacific Oceans than in the Atlantic Ocean because of the longer deep-water circulation pathways and thus accumulation of more DIC from respired CO_2 (Broecker 2003). Anthropogenic CO_2 penetration into the ocean is concentrated in the upper thermocline (Sabine et al. 2004), and over time this has contributed to the shoaling of saturation horizons by 30–200 m from the preindustrial period to the present; evidence of aragonite undersaturation in thermocline waters in the North Pacific and Indian Oceans (Feely et al. 2002, Sabine et al. 2002); seasonal upwelling of seawater corrosive to aragonite ($\Omega_{\text{arag}} < 1.0$) onto the western continental shelf of North America, approximately 40 years earlier than predicted by models (Feely et al. 2008); and an increase in areal extent of shallow undersaturated regions in the eastern tropical Atlantic (Chung et al. 2003, 2004).

BIOLOGICAL RESPONSES TO ACIDIFICATION

Early studies investigated responses of calcifying organisms to altered seawater carbonate chemistry (Agegian 1985, Borowitzka 1981, Fabry 1990, Smith & Roth 1979), but usually for reasons unrelated to anthropogenic CO_2 . Several ground-breaking studies, specifically designed to test atmospheric CO_2 impacts, revealed potentially dramatic responses in corals (Gattuso et al. 1998, Langdon et al. 2000, Marubini & Atkinson 1999, Marubini & Davies 1996), coral reef communities (Langdon et al. 2000,

invertebrates that is less soluble than high-magnesium calcite, which is found in some marine taxa such as echinoderms and coralline algae

10. Total alkalinity: a conserved thermodynamic measure of seawater acid-base chemistry that is equal to the charge difference between conservative cations and anions

2003; Leclercq et al. 2000), and planktonic organisms (Bijma 1991, Riebesell et al. 2000). Numerous other studies mostly verified these results but also revealed the complicated nature of calcification responses, for example to Mg/Ca ratios (Ries 2005, Ries et al. 2006, Stanley et al. 2005) or CO₂ in combination with changing temperature (Reynaud et al. 2003) or nutrients (Langdon & Atkinson 2005, Sciandra et al. 2003).

Calcifying organisms exert a variable degree of control over biomineralization, which generally involves passive and active ion movement in and out of a calcification compartment isolated from ambient seawater (Weiner & Dove 2003). Reduced calcification rates are observed following acidification for a variety of calcareous organisms even when aragonite or calcite $\Omega > 1.0$ (Royal Society 2005, Kleypas et al. 2006, Fabry et al. 2008). The degree of sensitivity varies among species, however, and some taxa may show enhanced calcification at CO₂ levels projected to occur over the 21st century (Iglesias-Rodríguez et al. 2008, Ries et al. 2008). However, calcification-CO₂ response studies exist for a limited number of species in many calcifying groups, and we currently lack sufficient understanding of calcification mechanisms to explain species-specific differences observed in manipulative experiments.

Thus far, most of the elevated CO₂ response studies on marine biota, whether for calcification, photosynthesis, or some other physiological measure, have been short-term experiments that range from hours to weeks. Chronic exposure to increased pCO₂ may have complex effects on the growth and reproductive success of calcareous plankton and could induce possible adaptations that are not observed in short-term experiments. Nevertheless, such laboratory experiments enable testing of single or multiple environmental variables in highly controlled settings and are critically needed to identify species' preadapted sensitivities to increasing CO₂.

Shallow-Water Tropical Corals and Coral Reefs

Scleractinian (stony) corals evolved in the Triassic period more than 200 million years ago (Mya), and their ability to produce large quantities of CaCO₃ has undoubtedly contributed to their evolutionary success (Wood 2001). Scleractinian corals are an important group of hyper-calcifiers, organisms with the capacity to produce massive quantities of CaCO₃ but in which calcification rates vary under different environmental conditions (Stanley & Hardie 1998).

A dramatic example is the work by Fine & Tchernov (2007) in which two species of corals grown in highly acidified water completely lost their skeletons (Figure 3), then regrew them after being returned to seawater of normal pH. The study highlights three points: (a) coral calcification rates can vary greatly in response to changes in pH and aragonite saturation state, (b) the naked, anemone-like coral polyps remained healthy, but (c) the fitness of organisms overall would change because of the loss of the protective skeleton. These results also support the paleontologically sudden appearance of scleractinian corals some 14 million years (Ma)¹¹ after the Permian extinction event; that is, corals may have continued to exist as “naked corals” until ocean chemistry became favorable for skeletal formation (Stanley & Fautin 2001).

Many laboratory studies on a variety of coral species, indeed almost every study published to date (Figure 4), confirm that coral calcification rates decrease in response to decreasing aragonite saturation state. Analyses of cores from massive coral colonies of the Great Barrier Reef show that calcification rates declined 21% between 1988 and 2003, although this decrease exceeds that expected from lowered saturation state alone and probably reflects the composite effects of a suite of changing environmental conditions (e.g., saturation state, temperature, nutrients) (Cooper et al. 2008).

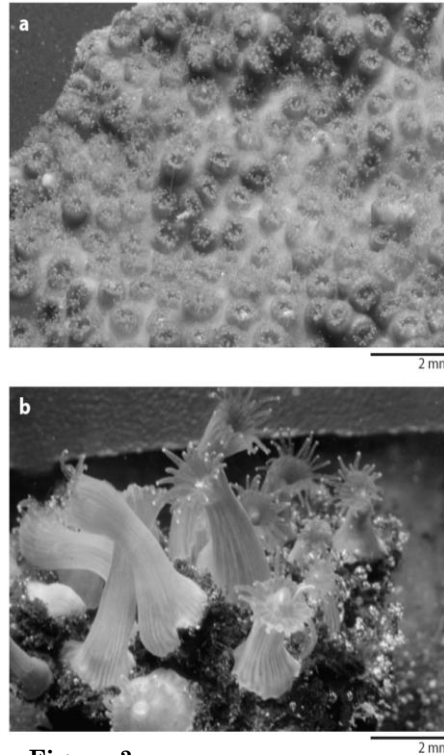
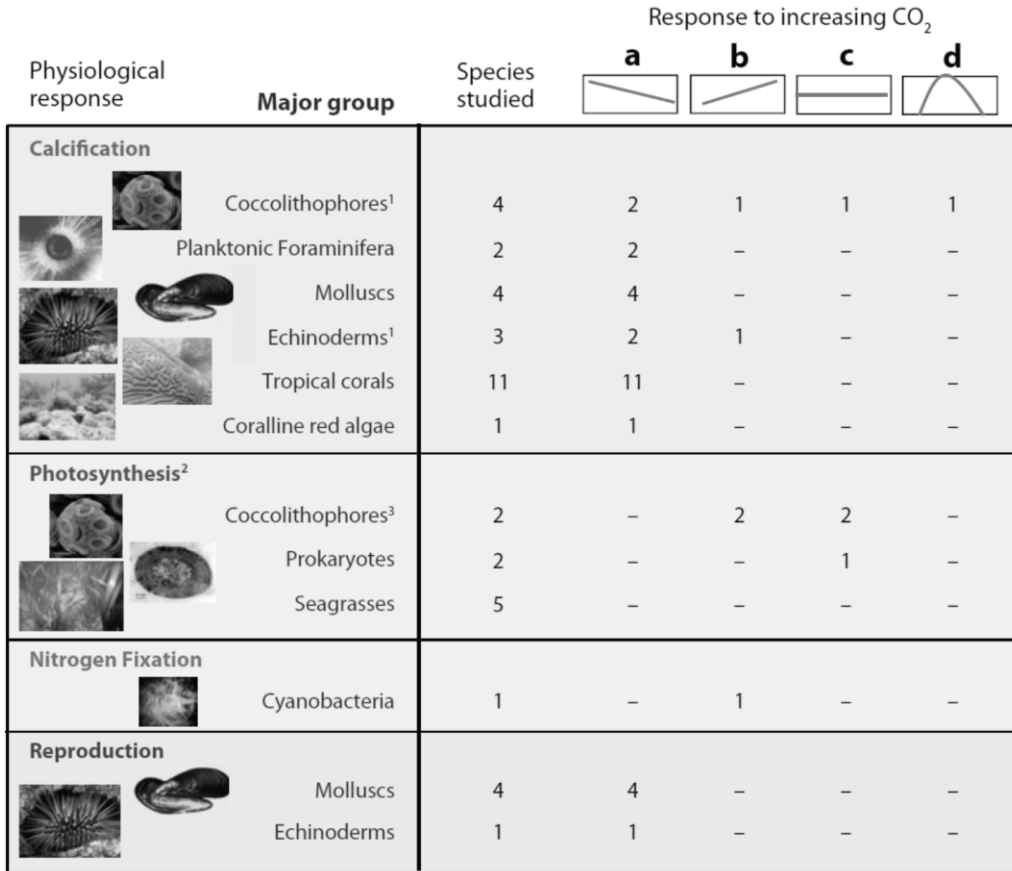


Figure 3

Photos of scleractinian coral *Oculina patagonica* after being maintained for 12 months in (a) normal seawater (pH = 8.2) and (b) acidified seawater (pH = 7.4). From Fine & Tchernov (2007). Reprinted with permission from AAAS.

11. Ma: millions of years

Many other benthic calcifying taxa are also both biogeochemically and ecologically important, including calcifying green algae and coralline red algae in particular. The



1) Increased calcification had substantial physiological cost; 2) Strong interactive effects with nutrient and trace metal availability, light, and temperature; 3) Under nutrient replete conditions.

Figure 4

Representative examples of impacts of ocean acidification on major groups of marine biota derived from experimental manipulation studies. The response curves on the right indicate four cases: (a) linear negative, (b) linear positive, (c) level, and (d) nonlinear parabolic responses to increasing levels of seawater pCO₂ for each of the groups. Note that in some cases strains of the same species exhibited different behavior in different experiments (cf. Fabry et al. 2008; Guinotte & Fabry 2008).

contribution of calcifying green algae in the genus *Halimeda* to the global net CaCO₃ production may rival that of coral reefs (Milliman & Droxler 1996, Rees et al. 2007). Coralline red algae are widespread, globally significant, but often overlooked benthic marine calcifiers (Foster 2001). A recent study on a common crustose coralline alga in Hawaii showed that both calcification rates and recruitment rates decline at lower carbonate saturation state (Kuffner et al. 2008), but relatively few studies have been conducted on either green or red algae.

Field measurements of reef calcification at the community scale (Bates 2002, Broecker & Takahashi 1966, Gattuso et al. 1995, Kawahata et al. 1999, Kayanne et al. 2005) consistently show that calcification rates are correlated with changes in a variety of components of the carbonate system in seawater (alkalinity, pCO₂, saturation state). A recent study suggests that inorganic precipitation of calcium carbonate cements, an important binding component in coral reefs, is correlated with saturation state and that the abundance of such cements may increase reef resistance to erosion (Manzello et al. 2008).

Coral reef ecosystems are defined by their ability to produce a net surplus of CaCO₃ that produces the reef structure (Kleypas et al. 2001). Reef structures typically span only 10–30 m in depth, but are structurally complex and support high marine biodiversity. During the repeated glacial to interglacial sea level transgressions from 3 Mya to the present, reef ecosystems thrived because their rapid accretion rates migrated the coral community upward and maintained the community within the minimum light levels for continued growth. Under increasing ocean acidification, not only will coral community calcification decrease, but also dissolution rates will increase (Langdon et al. 2000, Yates & Halley 2006), particularly for those reefs that are already near the limit for reef growth (e.g., higher latitude reefs). Interestingly, even though global warming may allow corals to migrate to higher latitudes (Precht & Aronson 2004), the decrease in reef CaCO₃ production may restrict reef development to lower latitudes where aragonite saturation levels can support calcium carbonate accumulation (Guinotte et al. 2003, Kleypas et al. 2001).

Deep-Water Corals and Carbonate Mounds

Deep-water scleractinian corals thrive in the subphotic waters of continental slopes, usually in depths of 200–1000 m.

They are slow growing and often long lived, up to 1500 years old, and form habitat that supports high biodiversity and fisheries. The maximum depth of these communities, particularly of the aragonitic scleractinian corals, appears to coincide with the depth of the aragonite saturation horizon (Guinotte et al. 2006), which reaches an average depth of > 2000 m in the North Atlantic, but can be as shallow as 200 m in the North Pacific Ocean (Figure 2). As in the case of their tropical counterparts, deep-water corals can produce large mounds of calcium carbonate, albeit much more slowly (Roberts et al. 2006). In contrast to the equatorward contraction of tropical coral reefs, it is the depth distribution of deep-water coral communities that will contract; the deepest communities will be the first to experience a shift from saturated to undersaturated conditions (Figure 2).

Other Benthic Invertebrates

Calcareous skeletal hard parts are widespread among benthic invertebrate phyla, yet apart from corals, few studies have investigated the effects of climate-relevant CO₂ increases on calcification in these fauna. Gazeau and coworkers (2007) reported that calcification rates in the mussel *Mytilus edulis* and the Pacific oyster *Crassostrea gigas* decreased by 25% and 10%, respectively, when grown at ~740 ppmv CO₂. Similarly, reduced shell growth was observed in the gastropod *Strombus luhuanus* and two sea urchin species when grown at 560 ppmv CO₂ over a 6-month period (Shirayama & Thorton 2005). Moreover, mussels, sea urchins (Michaelidis et al. 2005, Miles et al. 2007), and to a much lesser extent the crab *Necora puber* (Spicer et al. 2007) are sensitive to internal acidification of body fluids and use shell dissolution to compensate. Calcification in the arms of a burrowing brittle star increased when organisms were grown in low pH water (Wood et al. 2008; Figure 4, case b). However, decreased muscle mass in arms was also observed, which would reduce arm movement and likely decrease respiration and feeding. Thus, the enhanced calcification observed in short-term experiments would probably not be sustainable in the long-term.

The response of early developmental stages of benthic invertebrates to CO₂-induced acidification of seawater has been investigated in bivalves and sea urchins. Kurihara & Shirayama (2004) reported reduced fertilization success, developmental rates, larval size, and spicule skeletogenesis

with increasing CO₂ in the sea urchins *Hemicentrotus pulcherrimum* and *Echinodetra mathei*. When gametes of the oyster *C. gigas* were exposed to 2268 μatm^{12} pCO₂, no differences in rates of fertilization or embryo development were observed, relative to the control group, until 24 h after fertilization, when more than 80% of the D-shaped larvae grown in high-CO₂ seawater displayed malformed shells or remained unmineralized (Kurihara et al. 2007).

Planktonic Calcification

The major calcareous plankton groups are calcite-forming coccolithophores and foraminifera- and aragonite-forming euthecosomatous pteropods. Most planktonic studies have focused on coccolithophores, although to date these studies have covered only four of approximately 250 to 500 living coccolithophore species (Young et al. 2005) and revealed nonuniform calcification response to high-CO₂/low-pH seawater (Figure 4). Many laboratory and mesocosm studies with the bloom-forming coccolithophore species of *Emiliania huxleyi* and *Gephyrocapsa oceanica* reported decreased calcification (Figure 4, case a) that ranged from -25% to -66% when pCO₂ was increased to 560–840 μatm (Riebesell et al. 2000; Zondervan et al. 2001, 2002; Sciandra et al. 2003; Delille et al. 2005; Engel et al. 2005).

Recently, Iglesias-Rodríguez and colleagues (2008) found a doubling of cell-specific calcification rates for laboratory *E. huxleyi* cultures grown at 750 μatm versus 300 μatm pCO₂ (Figure 4, case b). Overall size increased under high-CO₂ conditions, as did size and mass of individual calcitic coccoliths that surround the cell surface. However, at the same time, significantly reduced growth rates at elevated pCO₂ suggest that this *E. huxleyi* strain would be more at risk of outcompetition by other phytoplankton species under future high-CO₂ conditions.

Similarly, Langer and coworkers (2006) found differing responses to elevated pCO₂ for two other coccolithophore species, both important in calcite export to sediments. The coccolithophore *Coccolithus pelagicus* exhibited no significant change in calcification for CO₂ varying from 150 to 915 μatm

12. Microatmosphere (μatm): a measure of seawater pCO₂; 1 μatm = 10⁻⁶ atm

(Figure 4, case c). Calcification rates decreased and coccolith malformations increased for *Calcidiscus leptoporus* cultures in response to pCO₂ levels both above and below present-day values (Figure 4, case d). In contrast, examination of Atlantic Ocean sediments from the Last Glacial Maximum (LGM), when surface water pCO₂ was approximately 200 μ atm, revealed no malformed coccoliths in *C. leptoporus*, suggesting that this species was adapted to low CO₂ during the LGM and has subsequently adjusted to elevated modern CO₂. Such resiliency is consistent with the findings of Iglesias-Rodríguez and colleagues (2008), who examined a high-resolution sediment core and observed a distinct increase in the average coccolith mass, dominated by *C. leptoporus* and *C. pelagicus*, from 1960 to 2000 that follows the rise in atmospheric CO₂.

In laboratory experiments with two species of symbiont-bearing, planktonic foraminifera, shell mass decreased as carbonate ion concentration decreased (Spero et al. 1997; Bijma et al. 1999, 2002). When grown in seawater chemistry equivalent to 560 and 740 ppmv CO₂, the shell mass of foraminifers *Orbulina universa* and *Globigerinoides sacculifer* declined by 4–8% and 6–14%, respectively, compared with preindustrial CO₂ controls.

Data from a single species of euthesomatous pteropod (*Clio pyramidata*) indicate net shell dissolution occurs in live pteropods when the saturation state of seawater with respect to aragonite is forced to $\Omega < 1.0$ (Orr et al. 2005, Fabry et al. 2008). Even though the animals were actively swimming, shell dissolution occurs within 48 hours when live pteropods, collected in the subarctic Pacific, are exposed to aragonite undersaturation levels similar to those projected for Southern Ocean surface waters by year 2100.

Primary Production and Nitrogen Fixation

Carbon-concentrating mechanisms enable most marine phytoplankton species to accumulate intracellular inorganic carbon either as CO₂ or HCO₃⁻ or both (Giordano et al. 2005). Largely because of these mechanisms, most marine phytoplankton tested in single-species laboratory studies and field population experiments show little or no change in photosynthetic rates when grown under high pCO₂ conditions equivalent to ~760 μ atm (Tortell et al. 1997, Hein & Sand-Jensen 1997, Burkhardt et al. 2001, Tortell & Morell 2002, Rost et al. 2003, Beardall & Raven 2004, Giordano et al. 2005,

Martin & Tortell 2006). In contrast, the widely distributed coccolithophore *E. huxleyi* has low affinity for inorganic carbon and could be carbon limited in the modern ocean (Rost & Riebesell 2004). Whether *E. huxleyi* will show increased rates of photosynthesis with progressive oceanic uptake of atmospheric CO₂ may depend on nutrient and trace metal availability, light conditions, and temperature (Zondervan 2007).

In a recent mesocosm CO₂ manipulation study with a phytoplankton community dominated by diatoms and coccolithophores, Riebesell and coworkers (2007) reported 27% and 39% higher CO₂ uptake in experimental pCO₂ treatments of 700 and 1050 μatm, respectively, relative to the 350 μatm pCO₂ treatment. Direct extrapolation of such experimental results to large-scale ocean regions may be problematic, however. The global warming that accompanies ocean acidification increases thermal stratification of the upper ocean, thereby reducing the upwelling of nutrients, and has been linked to observed decreases in phytoplankton biomass and productivity on a global basis (Behrenfeld et al. 2006).

Fu and coworkers (2007) explored in culture the physiological responses of two key marine photosynthetic cyanobacteria to warmer, more CO₂-rich conditions. They observed only minimal changes for *Prochlorococcus*, whereas for *Synechococcus* they measured greatly elevated photosynthesis rates and a 20% increase in cellular C:P and N:P ratios. The CO₂ and temperature responses were synergistic; that is, the change in photosynthesis for warm, high-CO₂ conditions was much greater than the sum of the responses to either factor individually.

Although seagrasses are able to utilize HCO₃⁻, most species do so inefficiently. In all species tested to date, light-saturated photosynthetic rates increase dramatically with increased dissolved CO_{2(aq)} concentration (Zimmerman et al. 1997, Short & Neckles 1999, Invers et al. 2001). In experiments with the eelgrass *Zostera marina*, Palacios & Zimmerman (2007) reported an increase in biomass and reproductive output under high-CO₂ conditions, suggesting potentially higher productivity of seagrasses that form a critical habitat for many fish and invertebrate species. The benefits from CO₂-enriched seawater, however, may be offset by the negative effects of increased temperature on vegetative growth (Ehlers et al. 2008).

Nitrogen-fixing cyanobacteria in the genus *Trichodesmium*,

which support a large fraction of primary productivity in low nutrient areas of the world's oceans, show increased rates of carbon and nitrogen fixation and increased C/N ratios under elevated pCO₂ (Hutchins et al. 2007, Barcelos e Ramos et al. 2007). At CO₂ levels of 750 ppmv, *Trichodesmium* CO₂ fixation rates increased by 15–128% and N₂ fixation rates increased by 35–100% relative to the rates in present day CO₂ conditions (Hutchins et al. 2007).

ECOLOGICAL AND BIOGEOCHEMICAL IMPACTS

Food Webs and Ecosystems

The limited number of studies conducted at climate-relevant CO₂ levels hampers predictions of the impacts of ocean acidification on marine ecosystems (Fabry et al. 2008). Nevertheless, many CaCO₃-secreting organisms clearly exhibit reduced calcification with elevated CO₂ and decreasing pH and CO₃²⁻. Assuming calcified structures provide protection from predators and/or other benefits to the organism, these calcifiers will either need to adapt to the changing seawater chemistry, shift their distributions to more carbonate ion-rich regions, or be adversely impacted.

In an intriguing field study, Hall-Spencer et al. (2008) quantify dramatic shifts in the nearshore benthic community in the vicinity of natural subsurface volcanic CO₂ vents. Consistent with expectations from laboratory studies, the regions near the vents under high-CO₂, low-pH water were marked by the absence of scleractinian corals and reduced abundances in sea urchins, coralline algae, and gastropods. The vent areas were dominated instead by seagrasses with an increased frequency of non-native, invasive species.

One clear threshold that will affect marine organisms is when surface waters become undersaturated with respect to their shell mineralogy. Surface waters of high latitude regions, for example, are projected to become undersaturated with respect to aragonite as early as 2050 (Orr et al. 2005), and the progressive shoaling of the aragonite saturation horizon will likely limit aragonitic organisms and change food web dynamics. Eutecosomatous pteropods are important components of polar and subpolar regions, where their densities can reach upward of thousands of individuals per cubic meter (Bathmann et al. 1991, Pane et al. 2004), and they are prey for a variety of zooplankton and fish predators. In the

North Pacific, pteropods can be important prey of juvenile pink salmon, accounting in some years for > 60% by weight of their diet (Armstrong et al. 2005). If pteropods cannot adapt to living continuously in seawater that is undersaturated with respect to aragonite, their ranges will contract to shallower depths and lower latitudes. Detection of such acidification-driven population shifts will be difficult because of a lack of baseline data on their current distributions and abundances.

More subtle ecological effects are also likely to occur, but are less predictable and will be more difficult to detect. For one, the calcification thresholds of many organisms do not coincide with the chemical threshold where saturation state $\Omega = 1$, but instead can occur at both higher and lower values. Many reef-building corals, for example, appear to cease calcification at aragonite saturation as high as 2.0. A slowdown in calcification for any organism may (a) reduce its ability to compete with noncalcifying organisms, as observed when crustose coralline algae were exposed to high-CO₂ conditions (Kuffner et al. 2008); (b) reduce the age at sexual maturity; (c) change its buoyancy; or (d) change light behavior in the water column (Tyrrell et al. 1999). Given that many taxa exhibit species-specific effects (Fabry 2008, Ries et al. 2008), each of these changes is likely to affect community dynamics in complicated ways, similar to the projected effects of temperature increases on terrestrial communities. Even small variations in species' responses will become amplified over successive generations and could drive major reorganizations of planktonic and benthic ecosystems. In the North Sea, one study has provided an interesting link between decreasing ocean pH and increasing jellyfish frequency since 1970 (Attrill et al. 2007).

However, ocean acidification is not occurring in isolation; it is one of many stressors related to climate change and other factors. The already difficult challenge of predicting the ecological effects of ocean acidification is magnified by these simultaneous changes, especially because the pace of the changes is unprecedented except for the most abrupt catastrophic events in Earth's history (e.g., the bolide impact that caused the Cretaceous-Tertiary extinction).

Oceanic CaCO₃ Budget

Despite the increasing evidence that calcification in many organisms will decline in the face of increased CO₂ levels (Figure 4), the impact on the global CaCO₃ budget remains

poorly constrained. Estimates of carbonate production in the water column have large uncertainties that range from 0.6–1.6 ± 0.3 Pg C yr⁻¹ based on satellite- and sediment-trap-derived estimates of carbonate production (Balch et al. 2007, Milliman 1993) (Table 1) to 0.4–1.8 Pg C yr⁻¹ based on models of carbonate export (Moore et al. 2004, Murnane et al. 1999). Production rates estimated from seasonal surface-ocean alkalinity changes yield a minimum estimate of 1.4 ± 0.3 Pg C yr⁻¹ (Lee 2001), which is consistent with the higher estimates. On the dissolution side, the estimates range from 0.5–1.0 Pg C yr⁻¹ based on sediment trap studies (Honjo et al. 2008), alkalinity gradients and water mass ages (Berelson et al. 2007, Feely et al. 2004, Sarmiento et al. 2002), and models (Gehlen et al. 2007). Carbonate burial in deep marine sediments is estimated to be 0.1 Pg C yr⁻¹ (Table 1).

Table 1 Summary of CaCO₃ flux estimates for the ocean updated from Berelson et al. (2007)

Flux term	Flux estimate mmol CaCO ₃ m ⁻² d ⁻¹	Flux estimate Pg C yr ⁻¹
Production in the euphotic zone	0.4 – 1.0	0.5 – 1.6
Export from the surface (models)	0.3 – 1.1	0.4 – 1.8
Dissolution from 200 to 1500 m		
Atlantic	0.3	0.1
Indian	1.1	0.3
Pacific	0.8	0.6
Total		1.0
Export to sediment traps below 2000 m*	0.24 ± 0.03	0.4 ± 0.05
Dissolution on sea floor below 2000 m	0.24 ± 0.17	0.4 ± 0.3
Atlantic and Pacific only		
Burial in sediments	0.08	0.1

*Data from Honjo et al. (2008).

Because ocean acidification is expected to decrease CaCO₃ saturation states (Figure 1b and c) and increase dissolution rates, ocean alkalinity and the ocean's capacity to take up more CO₂ from the atmosphere will presumably increase. If all carbonate production were shut down by ocean acidification, the atmospheric CO₂ would decline by approximately 10–20 ppmv (Gruber et al. 2004). In the near-term this may be

observed first in coastal regions where coral reef calcification rates could decrease by as much as 40% by the end of this century (Andersson et al. 2005, 2007). However, over the same timeframe, the uptake rate of anthropogenic carbon dioxide¹³ from the atmosphere could completely overwhelm these natural buffering mechanisms so the ocean's efficiency for taking up carbon will probably decline with time over the next two centuries.

Carbon and Nutrient Cycling

Ocean acidification also has the potential to alter ocean biogeochemical dynamics for organic carbon and nutrients via several more indirect pathways. Increased carbonate dissolution in the water column could decrease the contribution of CaCO₃ to the ballasting of organic carbon to the deep sea (Armstrong et al. 2002, Klaas & Archer 2002, Passow 2004), causing more organic carbon to remineralize in shallow waters and decreasing the ocean's CO₂ uptake efficiency. Laboratory and mesocosm experiments suggest a shift toward organic matter with higher C/N ratios for individual phytoplankton species and plankton communities grown at high CO₂. This could degrade the food quality for heterotrophic zooplankton and microbial consumers.

Changing elemental stoichiometries is one of the few mechanisms by which biology can alter ocean carbon storage (Boyd & Doney 2003), and higher C/N ratios in export material would make the ocean biological pump more efficient in exporting carbon to depth. Another such lever is nitrogen fixation because it decouples the otherwise tight connection between inorganic carbon and nitrogen in the water column. Higher nitrogen fixation rates at elevated CO₂ would provide additional new nitrogen in low-nutrient subtropical regions, zones where future primary production is expected otherwise to decline because of increased vertical stratification and reduced vertical nutrient inputs from below (Boyd & Doney 2002). The actual increase in nitrogen fixation, however, could be limited by phosphorus and iron supplies.

13. Anthropogenic carbon dioxide: excess carbon dioxide added to ocean and atmosphere from human fossil fuel combustion and deforestation

Chemical Speciation in Seawater

A major but underappreciated consequence of ocean acidification will be broad alterations of inorganic and organic seawater chemistry beyond the carbonate system. Analogous to the dramatic changes in the carbonate speciation, i.e., the measurable decrease in the concentration of carbonate ion and the increase in bicarbonate and aqueous CO₂, many other so-called weak acid species that undergo acid-base reactions in seawater will undergo significant speciation shifts with decreasing pH. Affected chemical species include those major elements such as boron, minor elements including phosphorus, silicon, and nitrogen, and trace elements such as iron, zinc, vanadium, arsenic, and chromium. For example, plots of chemical species concentration versus pH for phosphate, silicate, fluoride, and ammonia species as a function of pH show large changes with decreasing pH (figure 1.2.11 in Zeebe & Wolf-Gladrow 2001). These changes in speciation are important for understanding and modeling the responses of phytoplankton and other components of the marine ecosystem to changes in pH.

Similarly, many trace element species that are strongly hydrolyzed in seawater (e.g., aluminum, iron, chromium, bismuth, uranium) and form oxy-anion [MO_x-(OH)_n], hydroxyl [M(OH)_n], or carbonate complexes are also strongly influenced by variations in both temperature and pH (Byrne et al. 1988, Byrne 2002). These changes can directly affect their bioavailability to phytoplankton. For those species complexed with Cl⁻ ion, pH influences are much weaker. However, the details of the metal speciation as a function of pH, as well as the speciation impacts on bioavailability, are not as well known for the trace elements as they are for the major and minor elements. Even less well known is the influence of pH on metal organic complexes that are also abundant in the euphotic regions of the oceans (Bruland & Lohan 2004).

In much the same manner, dissolved organic matter that undergoes hydrolysis reactions in seawater (e.g., organic acids, amino acids, nucleic acids, proteins, humic materials) will also be strongly influenced by changing pH. The overall impact of decreasing pH on the structure and function of these biologically important organic compounds is largely unknown. Consequently, more research is needed on how ocean acidification will impact trace metal and organic matter speciation and biogeochemical

processes in the high-CO₂ oceans of the future.

GEOLOGICAL AND HISTORICAL EVIDENCE

Historical evidence for changes in ocean carbonate chemistry and calcification rates has been sought over three basic timescales: the recent past (decades; preindustrial through present); glacial-interglacial cycles (thousands of years); and the Phanerozoic (past 540 Ma), with particular emphasis on the Cenozoic (past 65 Ma). Boron isotope measurements, for example, are often used as a proxy for ocean pH (Palmer et al. 1998, Pearson & Palmer 2002, Sanyal et al. 1996), whereas changes in the CaCO₃ response are inferred from organism calcification rates or the depth of CaCO₃ deposition in the deep ocean.

Several studies have examined the coral calcification records from the previous decades to centuries for evidence of a recent decline in calcification. Many of these studies found little to no evidence that ocean acidification has caused a measurable decrease in calcification rates (Bessat & Buigues 2001; Lough & Barnes 1997, 2000). Analysis of coral calcification records is confounded by the difficulty of detecting an acidification signal within a naturally highly variable record, and ideally would include multiple cores across multiple locations (Lough 2004). As previously mentioned, one such analysis conducted on a suite of coral cores from widely spaced locations on the Great Barrier Reef showed that calcification rates declined by 21% between 1988 and 2003, although the cause of the decrease could not be specifically ascribed to ocean acidification (Cooper et al. 2008). Boron isotopes from coral skeletons have also been used to detect changes in seawater pH on a coral reef, but the recorded pH changes did not correlate well with the calcification signal (Pelejero et al. 2005); indeed, the use of boron isotopes in corals as a pH proxy is not uniformly accepted (Blamart et al. 2007, Honisch et al. 2004).

Over glacial-interglacial cycles (thousands of years), atmospheric CO₂ concentrations fluctuated between approximately 180 and 290 ppmv in concert with changes in orbital changes that affect solar forcing at the Earth's surface. Concurrent cycles in CaCO₃ deposition and dissolution occurred on the sea floor due to CaCO₃ compensation (Broecker & Takahashi 1978). In fact, the term ocean acidification was first used to describe a decrease in carbonate ion concentration in the western equatorial Atlantic at the onset of the last

glacial period (Broecker & Clark 2001). Various proxies (e.g., foraminiferal Zn/Ca and Mg/Ca ratios, boron isotopes, foraminiferal test characteristics) used to estimate changes in both surface and deep-ocean carbonate ion concentration tend to reinforce the CaCO₃ compensation hypothesis, but direct evidence for changes in the ocean carbonate system is lacking.

Over longer timescales (Ma), the paleontological record provides evidence for a strong correlation between atmospheric CO₂ and global temperature (Doney & Schimel 2007), but a less-than-convincing correlation between atmospheric CO₂ concentration and estimated ocean pH. Periods of high atmospheric CO₂ concentrations are common throughout the geologic record and some periods (e.g., Permian and Cretaceous) exhibit massive shallow-water CaCO₃ deposits, including reef structures. Initially this appears to be a conundrum: If high atmospheric CO₂ concentration produces acidic seas, why were CaCO₃ production and preservation so prevalent in these earlier high-CO₂ periods? The short answer to this question is that the carbonate saturation states may have been high during those periods despite the high pCO₂ levels. Fluid inclusions of seawater preserved in ancient halite deposits indicate that Ca²⁺, Mg²⁺ and SO₄²⁻ concentrations have varied by factor of two over the past 600 Ma (Mackenzie & Lerman 2006). The long answer is complicated and requires an understanding of the timescales over which various processes (e.g., climate and atmospheric CO₂, continental weathering, volcanism, methane clathrates, sea floor spreading) affect the carbon cycle and carbonate system in seawater (Dickens et al. 1995, Doney & Schimel 2007, MacKenzie & Morse 1992), as well as other long-term factors that affect calcification (e.g., biological evolution).

The most salient paleo-analog to the current atmospheric CO₂ increase is the strong ocean acidification event at the Paleocene-Eocene thermal maximum (PETM) 55 Mya. The PETM is marked by the sudden and massive carbon input to the ocean/atmosphere system, a shoaling of the deep ocean's calcite saturation horizon by at least 2 km in less than 2000 years that did not recover for tens of thousands of years, global warming of at least 5°C in less than 10,000 years, and major shifts in marine planktonic communities (Kennett & Stott 1991; Zachos et al. 1993, 2003, 2005). The only major extinctions occurred within the benthic foraminifera, though it is unclear whether ocean acidification was the main factor or

whether changes in ocean circulation led to anoxia in bottom waters (Zachos et al. 2008).

However, the similarity of the PETM and several comparable, but smaller, Eocene events to modern conditions is incomplete. First, whether the carbon excursion at the PETM was as rapid as the present-day excursion remains unclear. Second, the PETM and smaller events occurred within a background of already high CO₂ and global temperature. Third, the Mg:Ca ratio, an important factor that affects the carbonate mineralogy of many organisms, was also significantly different from that of today (Stanley & Hardie 2001). Finally, the marine biota during the PETM were also different. Corals and coral reefs had not yet re-established following the Cretaceous-Tertiary extinction (Wood 2001); modern coccolithophores are very different from those of the early Tertiary (Young 1994); and modern thecosomatous pteropod families appeared after the PETM, in the Eocene and Miocene (Lalli & Gilmer 1989).

SUMMARY POINTS

1. The surface ocean currently absorbs approximately one-third of the excess carbon dioxide (CO₂) injected into the atmosphere from human fossil fuel use and deforestation, which leads to a reduction in pH and wholesale shifts in seawater carbonate chemistry.
2. The resulting lowering of seawater carbonate ion concentrations and the saturation state for calcium carbonate are well documented in field data, and the rate of change is projected to increase over the 21st century unless predicted future CO₂ emissions are curbed dramatically.
3. Acidification will directly impact a wide range of marine organisms that build shells from calcium carbonate, from planktonic coccolithophores and pteropods and other molluscs, to echinoderms, corals, and coralline algae. Many calcifying species exhibit reduced calcification and growth rates in laboratory experiments under high-CO₂ conditions, whereas some photosynthetic organisms (both calcifying and noncalcifying) have higher carbon fixation rates under high CO₂.
4. Our present understanding of potential ocean acidification impacts on marine organisms stems

largely from short-term laboratory and mesocosm experiments; consequently, the response of individual organisms, populations, and communities to more realistic gradual changes is largely unknown (Boyd et al. 2008).

5. The potential for marine organisms to adapt to increasing CO₂ and the broader implications for ocean ecosystems are not well known; an emerging body of evidence suggests that the impact of rising CO₂ on marine biota will be more varied than previously thought, with both ecological winners and losers.
6. Ocean acidification likely will affect the biogeochemical dynamics of calcium carbonate, organic carbon, nitrogen, and phosphorus in the ocean as well as the seawater chemical speciation of trace metals, trace elements, and dissolved organic matter.
7. Acidification impacts processes so fundamental to the overall structure and function of marine ecosystems that any significant changes could have far-reaching consequences for the oceans of the future and the millions of people that depend on its food and other resources for their livelihoods.
8. Geo-engineering solutions that attempt to slow global warming without reducing atmospheric CO₂ concentration, such as injection of stratospheric aerosols (Crutzen 2006), will not reduce ocean acidification.

FUTURE ISSUES

A fully integrated program of laboratory, mesocosm, field monitoring, and modeling approaches is required to provide policymakers with informed management strategies that address how humans might best mitigate or adapt to these long-term changes. This program should emphasize how changes in the metabolic processes at the cellular level will be manifested within the ecosystem or community structure, and how they will influence future climate feedbacks. A program should include the following components:

1. A systematic monitoring system with high-resolution measurements in time and space of atmospheric and

surface water pCO₂, carbonate, alkalinity, and pH to validate model predictions and provide the foundations for interpreting the impacts of acidification on ecosystems;

2. In regions projected to undergo substantial changes in carbonate chemistry, tracking of abundances and depth distributions of key calcifying and noncalcifying species at appropriate temporal and spatial scales to enable the detection of possible shifts and discrimination between natural variability and anthropogenic-forced changes;
3. Standardized protocols and data reporting guidelines for carbonate system perturbation and calcification experiments;
4. Manipulative laboratory experiments to quantify physiological responses, including calcification and dissolution, photosynthesis, respiration, and other sensitive indices useful in predicting CO₂ tolerance of ecologically and economically important species;
5. New approaches to investigate/address long-term subtle changes that more realistically simulate natural conditions;
6. Mesocosm and field experiments to investigate community and ecosystem responses (i.e., shifts in species composition, food web structure, biogeochemical cycling, and feedback mechanisms) to elevated CO₂ and potential interactions with nutrients, light, and other environmental variables;
7. Integrated modeling approach to determine the likely implications of ocean acidification processes on marine ecosystems and fisheries, including nested models of biogeochemical processes and higher trophic-level responses to address ecosystem-wide dynamics such as competition, predation, reproduction, migration, and spatial population structure; and
8. Robust and cost-effective methods for measuring pH, pCO₂, and dissolved total alkalinity on moored buoys, ships of opportunity, and research vessels, floats, and gliders.

DISCLOSURE STATEMENT

The authors are not aware of any potential biases that might

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LITERATURE CITED

Agegian CR. 1985. *The biogeochemical ecology of Porolithon gardineri (Foslie)*. Ph.D. Thesis. Univ. Hawaii. 178 pp.

Andersson AJ, Bates NR, Mackenzie FT. 2007. Dissolution of carbonate sediments under rising pCO₂ and ocean acidification: observations from Devil's Hole, Bermuda. *Aquat. Geochem.* 13:237–64

Andersson AJ, Mackenzie FT, Lerman A. 2005. Coastal ocean and carbonate systems in the high CO₂ world of the anthropocene. *Am. J. Sci.* 305:875–918

Armstrong JL, Boldt JL, Cross AD, Moss JH, Davis ND, et al. 2005. Distribution, size, and interannual, seasonal and diel food habits of northern Gulf of Alaska juvenile pink salmon, *Oncorhynchus gorbuscha*. *Deep Sea Res. II* 52:247–65

Armstrong RA, Lee C, Hedges JI, Honjo S, Wakeham SG. 2002. A new, mechanistic model for organic carbon fluxes in the ocean based on the quantitative association of POC with ballast minerals. *Deep Sea Res. II* 49:219–36

Attrill MJ, Wright J, Edwards M. 2007. Climate-related increases in jellyfish frequency suggest a more gelatinous future for the North Sea. *Limnol. Oceanogr.* 52:480–85

Balch WM, Drapeau D, Bowler B, Booth E. 2007. Prediction of pelagic calcification rates using satellite measurements. *Deep Sea Res. II* 54:478–95

Barcelos e Ramos J, Biswas H, Schultz KG, LaRoche J, Riebesell U. 2007. Effect of rising atmospheric carbon dioxide on the marine nitrogen fixer *Trichodesmium*. *Glob. Biogeochem. Cycles* 21:GB2028

Bates NR. 2002. Seasonal variability of the effect of coral reefs on seawater CO₂ and air-sea CO₂ exchange. *Limnol. Oceanogr.* 47:43–52

Bathmann UV, Noji TT, von Bodungen B. 1991. Sedimentation of pteropods in the Norwegian Sea in autumn. *Deep Sea Res.* 38:1341–60

Beardall J, Raven JA. 2004. The potential effects of global climate change in microalgal photosynthesis, growth and ecology. *Phycologia* 43:31–45

Behrenfeld MJRT, O'Malley DA, Siegel CR, McClain JL, Sarmiento GC, et al. 2006. Climate-driven trends in contemporary ocean productivity. *Nature* 444:752–55

Berelson WM, Balch WM, Najjar R, Feely RA, Sabine C, Lee K. 2007. Relating estimates of CaCO₃ production, export, and dissolution in the water column to measurements of CaCO₃ rain into sediment traps and dissolution on the sea floor: a revised global carbonate budget. *Glob. Biogeochem. Cycles* 21

Bessat F, Buigues D. 2001. Two centuries of variation in coral growth in a massive *Porites* colony from Moorea (French Polynesia): a response of ocean-atmosphere variability from south central Pacific. *Palaeogeogr. Palaeoclim. Palaeoecol.* 175:381–92

Bijma J. 1991. Lunar pulses of carbonate output by spinose planktonic Foraminifera. In *Protozoa and Their Role in Marine Processes*, ed. PC Reid, CM Turley, PH Burkill, pp. 353–54. Plymouth: Elsevier

Bijma J, Honisch B, Zeebe RE. 2002. Impact of the ocean carbonate chemistry on living foraminiferal shell weight: comment on “Carbonate ion concentration in glacial-age deepwaters of the Caribbean Sea” by W.S. Broecker and E. Clark. *Geochem. Geophys. Geosys.* 3(11):1064

Bijma J, Spero HJ, Lea DW. 1999. Reassessing foraminiferal stable isotope geochemistry: impact of the oceanic carbonate system (experimental results). In *Use of Proxies in Paleoceanography: Examples from the South Atlantic*, ed. G Fischer, G Wefer, pp. 489–512. Springer-Verlag

Blamart D, Rollion-Bard C, Meibom A, Cuif JP, Juillet-Leclerc A, Dauphin Y. 2007. Correlation of boron isotopic composition with ultrastructure in the deep-sea coral *Lophelia pertusa*: implications for biomineralization and paleo-pH. *Geochem. Geophys. Geosys.* 8:Q12001

Borowitzka MA. 1981. Photosynthesis and calcification in the articulated coralline red algae *Amphiroa anceps* and *Amphiroa foliacea*. *Mar. Biol.* 62:17–23

Boyd PW, Doney SC. 2002. Modelling regional responses by marine pelagic ecosystems to global climate change. *Geophys.*

Res. Lett. 29:1806

Boyd P, Doney SC. 2003. The impact of climate change and feedback process on the ocean carbon cycle. In *Ocean Biogeochemistry*, ed. M. Fasham, pp. 157–93. Springer

Boyd P, Doney SC, Strzepek R, Dusenberry J, Lindsay K, Fung I. 2008. Climate-mediated changes to mixed-layer properties in the Southern Ocean: assessing the phytoplankton response *Biogeosciences* 5:847–64

Brewer PG, Sarmiento JL, Smethie WM. 1985. The Transient Tracers in the Ocean (TTO) program—the North-Atlantic Study, 1981—the Tropical Atlantic Study, 1983. *J. Geophys. Res. Oceans* 90:6903–5

Broecker WS. 2003. The oceanic CaCO₃ cycle. In *The Oceans and Marine Geochemistry, Treatise on Geochemistry*, ed. H Elderfield, pp. 529–49. London: Elsevier

Broecker W, Clark E. 2001. A dramatic Atlantic dissolution event at the onset of the last glaciation. *Geochem. Geophys. Geosystems* 2:1065

Broecker WS, Takahashi T. 1978. Relationship between lysocline depth and in situ carbonate ion concentration. *Deep Sea Res.* 25:65–95

Broecker WS, Takahashi T. 1966. Calcium carbonate precipitation on the Bahama Banks. *J. Geophys. Res.* 71:1575–602

Bruland KW, Lohan MC. 2004. The control of trace metals in seawater. In *The Oceans and Marine Geochemistry, Treatise on Geochemistry*, Vol. 6. ed. H. Elderfield. Elsevier

Burkhardt S, Amoroso G, Riebesell U, Sültemeyer D. 2001. CO₂ and HCO₃ uptake in marine diatoms acclimated to different CO₂ concentrations. *Limnol. Oceanogr.* 46:1378–91

Byrne RH, Kump LR, Cantrell KJ. 1988. The influence of temperature and pH on trace metal speciation in seawater. *Mar. Chem.* 25:163–81

Byrne RH. 2002. Inorganic speciation of dissolved elements in seawater: the influence of pH on concentration ratios. *Geochem. Trans.* 3:11

Caldeira K, Wickett ME. 2003. Anthropogenic carbon and ocean pH. *Nature* 425:365

Caldeira K, Wickett ME. 2005. Ocean model predictions of chemistry changes from carbon dioxide emissions to the atmosphere and ocean. *J. Geophys. Res Oceans* 110:C09S4

Chung SN, Lee K, Feely RA, Sabine CL, Millero FJ, et al. 2003. Calcium carbonate budget in the Atlantic Ocean based

on water column inorganic carbon chemistry. *Glob. Biogeochem. Cycles* 17:1093

Chung SN, Park GH, Lee K, Key RM, Millero FJ, et al. 2004. Postindustrial enhancement of aragonite undersaturation in the upper tropical and subtropical Atlantic Ocean: the role of fossil fuel CO₂. *Limnol. Oceanogr.* 49:315–21

Cooper TF, De 'Ath G, Fabricius KE, Lough JM. 2008. Declining coral calcification in massive *Porites* in two nearshore regions of the northern Great Barrier Reef. *Glob. Chang. Biol.* 14:529–38

Craig H, Turekian KK. 1976. GEOSECS Program - 1973–1976. *Earth Planet. Sci. Lett.* 32:217–19

Craig H, Turekian KK. 1980. The GEOSECS Program – 1976–1979. *Earth Planet. Sci. Lett.* 49:263–65

Crutzen PJ. 2006. Albedo enhancement by stratospheric sulfur injections: a contribution to resolve a policy dilemma? *Clim. Chang.* 77:211–20

Delille B, Harlay J, Zondervan I, Jacquet S, Chou L, et al. 2005. Response of primary production and calcification to changes of pCO₂ during experimental blooms of the coccolithophorid *Emiliana huxleyi*. *Glob. Biogeochem. Cycles* 19:GB2023

Dickens GR, Oneil JR, Rea DK, Owen RM. 1995. Dissociation of oceanic methane hydrate as a cause of the carbon-isotope excursion at the end of the Paleocene. *Paleoceanography* 10:965–71

Doney SC, Schimel DS. 2007. Carbon and climate system coupling on timescales from the Precambrian to the Anthropocene. *Annu. Rev. Environ. Resour.* 32:31–66

Doney SC, Mahowald N, Lima I, Feely RA, Mackenzie FT, et al. 2007. Impact of anthropogenic atmospheric nitrogen and sulfur deposition on ocean acidification and the inorganic carbon system. *Proc. Natl. Acad. Sci. USA* 104:14580–85

Ehlers A, Worm B, Reusch TBH. 2008. Importance of genetic diversity in eelgrass *Zostera marina* for its resilience to global warming. *Mar. Ecol. Prog. Ser.* 355:1–7

Engel A, Zondervan I, Aerts K, Beaufort L, Benthien A, et al. 2005. Testing the direct effect of CO₂ concentration on a bloom of the coccolithophorid *Emiliana huxleyi* in mesocosm experiments. *Limnol. Oceanogr.* 50(2):493–507

Fabry VJ. 1990. Shell growth rates of pteropod and heteropod mollusks and aragonite production in the open ocean—implications for the marine carbonate system. *J. Mar.*

Res. 48:209–22

Fabry VJ. 2008. Marine calcifiers in a high-CO₂ ocean. *Science* 320:1020–22

Fabry VJ, Seibel BA, Feely RA, Orr JC. 2008. Impacts of ocean acidification on marine fauna and ecosystem processes. ICES. *J. Mar. Sci.* 65:414–32

Feely RA, Byrne RH, Acker JG, Betzer PR, Chen CTA, et al. 1988. Winter summer variations of calcite and aragonite saturation in the northeast Pacific. *Mar. Chem.* 25:227–41

Feely RA, Chen CTA. 1982. The effect of excess CO₂ on the calculated calcite and aragonite saturation horizons in the northeast Pacific. *Geophys. Res. Lett.* 9:1294–97

Feely RA, Orr J, Fabry VJ, Kleypas JA, Sabine CL, Landgon C. 2008. Present and future changes in seawater chemistry due to ocean acidification. In *AGU Monograph on the Science and Technology of Carbon Sequestration*. ed. B.J. McPherson, E.T. Sundquist. Am. Geophys. Union. In press

Feely RA, Sabine CL, Hernandez-Ayon JM, Ianson D, Hales B. 2008. Evidence for upwelling of corrosive “acidified” water onto the continental shelf. *Science* 320:1490–92

Feely RA, Sabine CL, Lee K, Berelson W, Kleypas J, et al. 2004. Impact of anthropogenic CO₂ on the CaCO₃ system in the oceans. *Science* 305:362–66

Feely RA, Sabine CL, Lee K, Millero FJ, Lamb MF, et al. 2002. In situ calcium carbonate dissolution in the Pacific Ocean. *Glob. Biogeochem. Cycles* 16:1144

Fine M, Tchernov D. 2007. Scleractinian coral species survive and recover from decalcification. *Science* 315:1811

Foster MS. 2001. Rhodoliths: between rocks and soft places. *J. Phycol.* 37:1–9

Fu F-X, Warner ME, Zhang Y, Feng Y, Hutchins DA. 2007. Effects of increased temperature and CO₂ on photosynthesis, growth and elemental ratios of marine *Synechococcus* and *Prochlorococcus* (cyanobacteria). *J. Phycol.* 43:485–96

Gattuso JP, Frankignoulle M, Bourge I, Romaine S, Buddemeier RW. 1998. Effect of calcium carbonate saturation of seawater on coral calcification. *Glob. Planet. Chang.* 18:37–46

Gattuso JP, Pichon M, Frankignoulle M. 1995. Biological control of air-sea CO₂ fluxes: effect of photosynthetic and calcifying marine organisms and ecosystems. *Mar. Ecol. Prog. Ser.* 129:307–12

Gazeau F, Quiblier C, Jansen JM, Gattuso J-P, Middelburg

JJ, Heip CHR. 2007. Impact of elevated CO₂ on shellfish calcification. *Geophys. Res. Lett* 34:L07603

Gehlen M, Gangsto R, Schneider B, Bopp L, Aumont O, Ethe C. 2007. The fate of pelagic CaCO₃ production in a high CO₂ ocean: a model study. *Biogeosciences* 4:505–19

Giordano M, Beardall J, Raven JA. 2005. CO₂ concentrating mechanisms in algae: mechanisms, environmental modulation, and evolution. *Ann. Rev. Plant Biol.* 56:99–131

Gruber N, Friedlingstein P, Field CB, Valentini R, Heimann M, et al. 2004. The vulnerability of the carbon cycle in the 21st century: an assessment of carbon-climate-human interactions. In *The Global Carbon Cycle: Integrating Humans, Climate and the Natural World*, ed. CB Field, MR Raupach, pp. 45–76. Washington, DC: Island Press

Guinotte JM, Buddemeier RW, Kleypas JA. 2003. Future coral reef habitat marginality: temporal and spatial effects of climate change in the Pacific basin. *Coral Reefs* 22:551–8

Guinotte JM, Fabry VJ. 2008. Ocean acidification and its potential effects on marine ecosystems. *Ann. New York Acad. Sci.* 1134:320–42

Guinotte JM, Orr J, Cairns S, Freiwald A, Morgan L, George R. 2006. Will human-induced changes in seawater chemistry alter the distribution of deep-sea scleractinian corals? *Front. Ecol. Environ.* 4:141–46

Hall-Spencer JM, Rodolfo-Metalpa R, Martin S, Ransome E, Fine M, et al. 2008. Volcanic carbon dioxide vents show ecosystem effects of ocean acidification. *Nature* 454:96–99

Hein M, Sand-Jensen K. 1997. CO₂ increases oceanic primary production. *Nature* 388:526

Henderson C. 2006. Ocean acidification: the other CO₂ problem *New Scientist*. <http://environment.newscientist.com/article/mg19125631.200>

Honisch B, Hemming NG, Grottoli AG, Amat A, Hanson GN, Buma J. 2004. Assessing scleractinian corals as recorders for paleo-pH: Empirical calibration and vital effects. *Geochim. Cosmochim. Acta* 68:3675–85

Honjo S, Manganini SJ, Krishfield RA, Francois R. 2008. Particulate organic carbon fluxes to the ocean interior and factors controlling the biological pump: a synthesis of global sediment trap programs since 1983. *Prog. Oceanogr.* 76:217–85

Hutchins DA, Fe F-X, Zhang Y, Warner ME, Feng Y, et al. 2007. CO₂ control of *Trichodesmium* N₂ fixation, photosynthesis, growth rates, and elemental ratios:

implications for past, present, and future ocean biogeochemistry. *Limnol. Oceanogr.* 52(4):1293–1304

Iglesias-Rodríguez MD, Halloran PR, Rickaby REM, Hall IR, Colmenero-Hidalgo E, et al. 2008. Phytoplankton calcification in a high CO₂ world. *Science*.320:336–39

Invers O, Zimmeran RC, Alberte RS, Perez M, Romero J. 2001. Inorganic carbon sources for seagrass photosynthesis: an experimental evaluation of bicarbonate use in species inhabiting temperate waters. *J. Exp. Mar. Biol. Ecol.* 265:203–217

Kawahata H, Suzuki A, Goto K. 1999. Coral reefs as sources of atmospheric CO₂—spatial distribution of PCO₂ in Majuro Atoll. *Geochem. J.* 33:295–303

Kayanne H, Hata H, Kudo S, Yamano H, Watanabe A, et al. 2005. Seasonal and bleaching-induced changes in coral reef metabolism and CO₂ flux. *Glob. Biogeochem. Cycles* 19:GB3015

Keeling CD, Rakestraw NW, Waterman LS. 1965. Carbon dioxide in surface waters of the Pacific Ocean. 1. Measurements of the distribution. *J. Geophys. Res.* 70:6087–97

Kennett JP, Stott LD. 1991. Abrupt deep-sea warming, palaeoceanographic changes and benthic extinctions at the end of the Paleocene. *Nature* 353:225–29

Klaas C, Archer DE. 2002. Association of sinking organic matter with various types of mineral ballast in the deep sea: implications for the rain ratio. *Glob. Biogeochem. Cycles* 16:1116

Kleypas JA, Buddemeier RW, Archer D, Gattuso JP, Langdon C, Opdyke BN. 1999. Geochemical consequences of increased atmospheric carbon dioxide on coral reefs. *Science* 284:118–20

Kleypas JA, Buddemeier RW, Gattuso JP. 2001. The future of coral reefs in an age of global change. *Int. J. Earth Sci.* 90:426–37

Kleypas JA, Feely RA, Fabry VJ, Langdon C, Sabine CL, Robbins LL. 2006. Impacts of ocean acidification on coral reefs and other marine calcifiers: a guide for future research. 88 pp. *Report of a workshop sponsored by NSF, NOAA, and the U.S. Geological Survey*. St. Petersburg, Florida

Krogh A. 1904. On the tension of carbonic acid in natural waters and especially in the sea. *Medd. Groenl.* 26:231–342

Kuffner IB, Andersson AJ, Jokiel PL, Rodgers KS, Mackenzie FT. 2008. Decreased abundance of crustose coralline algae due to ocean acidification. *Nat. Geosci.* 1:77–140

Kurihara H, Kato S, Ishimatsu A. 2007. Effects of increased seawater pCO₂ on early development of the oyster *Crassostrea gigas*. *Aquat. Biol.* 1:91–98

Kurihara H, Shirayama Y. 2004. Effects of increased atmospheric CO₂ on sea urchin early development. *Mar. Ecol. Prog. Ser.* 274:161–69

Lalli CM, Gilmer RW. 1989. *Pelagic Snails: The Biology of Holoplanktonic Gastropod Mollusks*. Stanford: Stanford Univ. Press. 275 pp.

Langdon C, Atkinson MJ. 2005. Effect of elevated pCO₂ on photosynthesis and calcification of corals and interactions with seasonal change in temperature/irradiance and nutrient enrichment. *J. Geophys. Res. Oceans* 110:C09S7

Langdon C, Broecker WS, Hammond DE, Glenn E, Fitzsimmons K, et al. 2003. Effect of elevated CO₂ on the community metabolism of an experimental coral reef. *Glob. Biogeochem. Cycles* 17:1011

Langdon C, Takahashi T, Sweeney C, Chipman D, Goddard J, et al. 2000. Effect of calcium carbonate saturation state on the calcification rate of an experimental coral reef. *Glob. Biogeochem. Cycles* 14:639–54

Langer MR, Geisen M, Baumann K-H, Kläs J, Riebesell U, et al. 2006. Species-specific responses of calcifying algae to changing seawater carbonate chemistry. *Geochem. Geophys. Geosys.* 7:Q09006

Leclercq N, Gattuso JP, Jaubert J. 2000. CO₂ partial pressure controls the calcification rate of a coral community. *Glob. Chang. Biol.* 6:329–34

Lee K. 2001. Global net community production estimated from the annual cycle of surface water total dissolved inorganic carbon. *Limnol. Oceanogr.* 46:1287–97

Lough JM. 2004. A strategy to improve the contribution of coral data to high-resolution paleoclimatology. *Palaeogeogr. Palaeoclim. Palaeoecol.* 204:115–43

Lough JM, Barnes DJ. 1997. Several centuries of variation in skeletal extension, density and calcification in massive *Porites* colonies from the Great Barrier Reef: a proxy for seawater temperature and a background of variability against which to identify unnatural change. *J. Exp. Mar. Biol. Ecol.* 211:29–67

Lough JM, Barnes DJ. 2000. Environmental controls on growth of the massive coral *Porites*. *J. Exp. Mar. Biol. Ecol.* 245:225-43

Lüthi D, Le Floch M, Bereiter B, Blunier T, Barnola J-M, et al. 2008. High-resolution carbon dioxide concentration record 650,000-800,000 years before present. *Nature* 453:379–82

Mackenzie FT, Lerman A. 2006. *Carbon in the Geobiosphere—Earth's Outer Shell*. Dordrecht, The Netherlands: Springer. 402 pp.

Mackenzie FT, Morse JW. 1992. Sedimentary carbonates through Phanerozoic time. *Geochim. Cosmochim. Acta* 56:3281–95

Manzello DP, Kleypas JA, Budd DA, Eakin CM, Glynn PW, Langdon C. 2008. Poorly cemented coral reefs of the eastern tropical Pacific: possible insights into reef development in a high CO₂ world. *Proc. Nat. Acad. Sci. USA*. In press

Martin, CL, Tortell PD. 2006. Bicarbonate transport and extracellular carbonic anhydrase activity in Bering Sea phytoplankton assemblages: results from isotope disequilibrium experiments. *Limnol Oceanogr* 51:2111–21

Marubini F, Atkinson MJ. 1999. Effects of lowered pH and elevated nitrate on coral calcification. *Mar. Ecol. Prog. Ser.* 188:117–21

Marubini F, Davies PS. 1996. Nitrate increases zooxanthellae population density and reduces skeletogenesis in corals. *Mar. Biol.* 127:319–28

Michaelidis B, Ouzounis C, Paleras A, Pörtner HO. 2005. Effects of long-term moderate hypercapnia on acid-base balance and growth rate in marine mussels *Mytilus galloprovincialis*. *Mar. Ecol. Prog. Ser.* 293:109–118

Miles H, Widdicombe S, Spicer JI, Hall-Spencer J. 2007. Effects of anthropogenic seawater acidification on acid-base balance in the sea urchin *Psammechinus miliaris*. *Mar. Pollut. Bull.* 54:89–96

Millero FJ, Pierrot D, Lee K, Wanninkhof R, Feely R, et al. 2002. Dissociation constants for carbonic acid determined from field measurements. *Deep Sea Res. I Oceanogr. Res. Papers* 49:1705–23

Milliman JD. 1993. Production and accumulation of calcium carbonate in the ocean—budget of a nonsteady state. *Glob. Biogeochem. Cycles* 7:927–57

Milliman JD, Droxler AW. 1996. Neritic and pelagic carbonate sedimentation in the marine environment: ignorance is not bliss. *Geol. Rundsch.* 85:496–504

Moore JK, Doney SC, Lindsay K. 2004. Upper ocean ecosystem dynamics and iron cycling in a global three-

dimensional model. *Glob. Biogeochem. Cycles* 18:GB4028

Mucci A. 1983. The solubility of calcite and aragonite in seawater at various salinities, temperatures, and one atmosphere total pressure. *Am. J. Sci.* 283:780–99

Murnane RJ, Sarmiento JL, Le Quere C. 1999. Spatial distribution of air-sea CO₂ fluxes and the interhemispheric transport of carbon by the oceans. *Glob. Biogeochem. Cycles* 13:287–305

Orr JC, Fabry VJ, Aumont O, Bopp L, Doney SC, et al. 2005. Anthropogenic ocean acidification over the twenty-first century and its impact on calcifying organisms. *Nature* 437:681–86

Palacios S, Zimmerman RC. 2007. Response of eelgrass *Zostera marina* to CO₂ enrichment: possible impacts of climate change and potential for remediation of coastal habitats. *Mar. Ecol. Prog. Ser.* 344:1–13

Palmer MR, Pearson PN, Cobb SJ. 1998. Reconstructing past ocean pH-depth profiles. *Science* 282:1468–71

Pane L, Feletti M, Francomacaro B, Mariottini GL. 2004. Summer coastal zooplankton biomass and copepod community structure near the Italian Terra Nova Base (Terra Nova Bay, Ross Sea, Antarctica). *J. Plankton Res.* 26(12):1479–88

Passow U. 2004. Switching perspectives: Do mineral fluxes determine particulate organic carbon fluxes or vice versa? *Geochem. Geophys. Geosyst.* 5:Q04002

Pearson PN, Palmer MR. 2002. The boron isotope approach to paleo-pCO₂ estimation. *Geochim. Cosmochim. Acta* 66:A586

Pelejero C, Calvo E, McCulloch MT, Marshall JF, Gagan MK, et al. 2005. Preindustrial to modern interdecadal variability in coral reef pH. *Science* 309:2204–7

Precht WF, Aronson RB. 2004. Climate flickers and range shifts of reef corals. *Front. Ecol. Environ.* 2:307–14

Rees SA, Opdyke BN, Wilson PA, Henstock TJ. 2007. Significance of *Halimeda* bioherms to the global carbonate budget based on a geological sediment budget for the Northern Great Barrier Reef, Australia. *Coral Reefs* 26:177–88

Reynaud S, Leclercq N, Romaine-Lioud S, Ferrier-Pages C, Jaubert J, Gattuso JP. 2003. Interacting effects of CO₂ partial pressure and temperature on photosynthesis and calcification in a scleractinian coral. *Glob. Chang. Biol.* 9:1660–68

Riebesell U, Schulz KG, Bellerby RGJ, Botros M, Fritsche P, et al. 2007. Enhanced biological carbon consumption in a high CO₂ ocean. *Nature* 450:545–48

Riebesell U, Zondervan I, Rost B, Tortell PD, Zeebe RE,

Morel FMM. 2000. Reduced calcification of marine plankton in response to increased atmospheric CO₂. *Nature* 407:364–67

Ries JB. 2005. Aragonite production in calcite seas: effect of seawater Mg/Ca ratio on the calcification and growth of the calcareous alga *Penicillus capitatus*. *Paleobiology* 31:445–58

Ries JB, Cohen AL, McCorkle DC. 2008. *Marine biocalcifiers exhibit mixed responses to CO₂-induced ocean acidification*. 11th Int. Coral Reef Symp., Ft. Lauderdale

Ries JB, Stanley SM, Hardie LA. 2006. Scleractinian corals produce calcite, and grow more slowly, in artificial Cretaceous seawater. *Geology* 34:525–28

Roberts JM, Wheeler AJ, Freiwald A. 2006. Reefs of the deep: the biology and geology of cold-water coral ecosystems. *Science* 312:543–47

Rost B, Riebesell U, Burkhardt S. 2003. Carbon acquisition of bloom-forming marine phytoplankton. *Limnol. Oceanogr.* 48:55–67

Rost B, Riebesell U. 2004. Coccolithophores and the biological pump: responses to environmental changes. In *Coccolithophores—From Molecular Processes to Global Impact*, ed. Thierstein HR, Young JR, p. 76–99. Springer

Royal Society. 2005. *Ocean acidification due to increasing atmospheric carbon dioxide*, London: The Royal Society, 57 pp.

Sabine CL, Feely RA. 2007. The oceanic sink for carbon dioxide. In *Greenhouse Gas Sinks*, ed. D Reay, N Hewitt, J Grace, K Smith, pp. 31–49. Oxfordshire: CABI Publishing

Sabine CL, Feely RA, Gruber N, Key RM, Lee K, et al. 2004. The oceanic sink for anthropogenic CO₂. *Science* 305:367–71

Sabine CL, Key RM, Feely RA, Greeley D. 2002. Inorganic carbon in the Indian Ocean: distribution and dissolution processes. *Glob. Biogeochem. Cycles* 16:1067

Sanyal A, Hemming NG, Broecker WS, Lea DW, Spero HJ, Hanson GN. 1996. Oceanic pH control on the boron isotopic composition of foraminifera: evidence from culture experiments. *Paleoceanography* 11:513–17

Sarmiento JL, Dunne J, Gnanadesikan A, Key RM, Matsumoto K, Slater R. 2002. A new estimate of the CaCO₃ to organic carbon export ratio. *Glob. Biogeochem. Cycles* 16:1107

Sciandra A, Harlay J, Lefevre D, Lemee R, Rimmelin P, et al. 2003. Response of coccolithophorid *Emiliania huxleyi* to elevated partial pressure of CO₂ under nitrogen limitation. *Mar. Ecol. Prog. Ser.* 261:111–22

Shirayama Y, Thorton H. 2005. Effect of increased

atmospheric CO₂ on shallow water marine benthos. *J. Geophys. Res.* 110:C09S08

Short FT, Neckles HA. 1999. The effects of global climate change on seagrasses. *Aquat. Bot.* 63:169–96

Smith AD, Roth AA. 1979. Effect of carbon dioxide concentration on calcification in the red coralline alga *Bossiella orbigniana*. *Mar. Biol.* 52:217–25

Smith SV, Buddemeier RW. 1992. Global change and coral reef ecosystems. *Annu. Rev. Ecol. Syst.* 23:89–118

Solomon S, Qin D, Manning M, Chen Z, Marquis M, et al. 2007. *Climate Change 2007: The Physical Science Basis: Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. New York: Cambridge Univ. Press

Spero HJ, Bijma J, Lea DW, Bemis BE. 1997. Effect of seawater carbonate concentration on foraminiferal carbon and oxygen isotopes. *Nature* 390:497–500

Spicer JI, Raffo A, Widdicombe S. 2007. Influence of CO₂-related seawater acidification on extracellular acid-base balance in the velvet swimming crab, *Necora puber*. *Mar. Biol.* 151:1117–25

Stanley GD, Fautin DG. 2001. The origins of modern corals. *Science* 291:1913–14

Stanley SM, Hardie LA. 1998. Secular oscillations in the carbonate mineralogy of reef-building and sediment-producing organisms driven by tectonically forced shifts in seawater chemistry. *Palaeogeogr. Palaeoclim. Palaeoecol.* 144:3–19

Stanley SM, Hardie LA. 2001. Secular variation in Phanerozoic marine biocalcification and the original mineralogy and mode of calcification of receptaculitids: a reply. *Palaeogeogr. Palaeoclim. Palaeoecol.* 168:363–64

Stanley SM, Ries JB, Hardie LA. 2005. Seawater chemistry, coccolithophore population growth, and the origin of Cretaceous chalk. *Geology* 33:593–96

Takahashi T. 1961. Carbon dioxide in the atmosphere and in Atlantic Ocean water. *J. Geophys. Res.* 66:477–94

Takahashi T, Broecker WS, Bainbridge AE, Weiss RF. 1980. Carbonate chemistry of the Atlantic, Pacific, and Indian Oceans: The results of the GEOSECS expeditions, 1973–1978 National Science Foundation, Washington D.C.

Takahashi T, Sutherland SC, Feely RA, Wanninkhof R. 2006. Decadal change of surface water pCO₂ in the North Pacific: a synthesis of 35 years of observations. *J. Geophys.*

Res. 111: C07S05

Tortell PD, Morel FMM. 2002. Sources of inorganic carbon for phytoplankton in the eastern Subtropical and Equatorial Pacific Ocean. *Limnol. Oceanogr.* 47:1012–22

Tortell PD, Reinfelder JR, Morel FMM. 1997. Active uptake of bicarbonate by diatoms. *Nature* 390(6657):243–44

Turley C. 2005. The other CO₂ problem. openDemocracy. http://www.acamedia.info/sciences/sciliterature/globalw/reference/carol_turley.html

Tyrrell T, Holligan PM, Mobley CD. 1999. Optical impacts of oceanic coccolithophore blooms, *J. Geophys. Res.* 104(C2):3223–41

Weiner S, Dove PM. 2003. An overview of biomineralization processes and the problem of the vital effect. In *Biomineralization*, Reviews in Mineralogy and Geochemistry. Volume 54. eds. PM Dove, JJ De Yoreo, S Weiner, pp. 1–29. Mineralog. Soc. Am., Geochem. Soc.

Wood R. 2001. *Reef Evolution*. Oxford: Oxford Univ. Press. 414 pp.

Wood HL, Spicer JI, Widdicombe S. 2008. Ocean acidification may increase calcification rates, but at a cost. *Proc. R. Soc. London Ser. B* 275:1767–73

Yates KK, Halley RB. 2006. CO₃²⁻ concentration and pCO₂ thresholds for calcification and dissolution on the Molokai reef flat, Hawaii. *Biogeosciences* 3:357–69

Young JR. 1994. Functions of coccoliths. In *Coccolithophores*, ed. A Winter, WG Seisser, pp. 63–82. New York: Cambridge Univ. Press

Young JR, Geisen M, Probert I. 2005. A review of selected aspects of coccolithophore biology with implications for paleobiodiversity estimation. *Micropaleontology* 51:267–88

Zachos JC, Dickens GR, Zeebe RE. 2008. An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics. *Nature* 451:279–83

Zachos JC, Lohmann KC, Walker JCG, Wise SW. 1993. Abrupt climate change and transient climates during the Paleogene—a marine perspective. *J. Geol.* 101:191–213

Zachos JC, Rohl U, Schellenberg SA, Sluijs A, Hodell DA, et al. 2005. Rapid acidification of the ocean during the Paleocene-Eocene thermal maximum. *Science* 308:1611–15

Zachos JC, Wara MW, Bohaty S, Delaney ML, Petrizzo MR, et al. 2003. A transient rise in tropical sea surface temperature during the Paleocene-Eocene thermal maximum. *Science*

302:1551–54

Zeebe RE, Wolf-Gladrow D. 2001. *CO₂ in Seawater: Equilibrium, Kinetics, Isotopes*. Amsterdam: Elsevier Science, B.V. 346 pp.

Zimmerman RC, Kohrs DG, Steller DL, Alberte RS. 1997. Impacts of CO₂-enrichment on productivity and light requirements of eelgrass. *Plant Physiol.* 115:599–607

Zondervan I. 2007. The effects of light, macronutrients, trace metals and CO₂ on the production of calcium carbonate and organic carbon in coccolithophores—a review. *Deep Sea Res. II* 54:521–37

Zondervan I, Rost B, Riebesell U. 2002. Effect of CO₂ concentration on the PIC/POC ratio in the coccolithophore *Emiliania huxleyi* grown under light-limiting conditions and different daylengths. *J. Exp. Mar. Biol. Ecol.* 272:55–70

Zondervan I, Zeebe RE, Rost B, Riebesell U. 2001. Decreasing marine biogenic calcification: A negative feedback on rising atmospheric pCO₂. *Glob. Biogeochem. Cycles* 15:507–16