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Supplement of

Reviews and Syntheses: Ocean acidification and its potential impacts on marine ecosystems

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1 **2.1 Increasing dissolution of atmospheric CO₂ to seawater**

2 The exchange of CO₂ between seawater and the atmosphere can take place following
3 different mechanisms and with different features. The outcome is a very varied degree of
4 CO₂ uptake from, and/or emission to, the atmosphere, with time scales for variability that
5 are in the daily to seasonal range. The main situations that may be encountered in
6 seawater are listed below.

7 **2.1.1 Sinking or balance of atmospheric CO₂ to seawater under sunlight, and**
8 **emission or balance of CO₂ to the atmosphere during the night.** This scenario can
9 take place with a time span ranging from minutes to several hours (Biswas et al., 2004;
10 DeGrandpre et al., 1998; Jiang et al., 2011; Nakamura and Nakamori, 2009; Zhang et al.,
11 2013), and two phenomena can be involved: (i) Elevated photosynthesis produces a
12 strong consumption of seawater CO₂, thereby decreasing the fugacity of CO₂ ($f\text{CO}_2$) in
13 water and making seawater itself an important sink of atmospheric CO₂. Elevated
14 photosynthesis would be associated to a high degree of degradation/respiration of organic
15 matter including primary producers (PP) and dissolved organic matter (DOM), which
16 causes a fast recycling of nutrients and provides optimum conditions for the primary
17 productivity to be high. Such a behavior is generally observed in waters with a high
18 degree of eutrophication. (ii) During the night, an elevated degree of
19 degradation/respiration of organic material would substantially increase the $f\text{CO}_2$ in
20 seawater, so that CO₂ will be released to the atmosphere (Ma and Green, 2004; Xie et al.,
21 2004). The high degree of respiration during the night is made possible by the availability
22 of organic material and, produced by the elevated photosynthetic activity at daytime.
23 Highly eutrophic waters with elevated nutrients that show such a behavior can be found
24 in the Jiulongjiang estuary (Fig. 2a), in Mangrove coastal locations (Biswas et al., 2004),
25 in Luhuitou fringing reef, Sanya Bay, South China Sea (Zhang et al., 2013), in the Gulf
26 of Biscay and the North Sea (Frankignoulle and Borges, 2001), in the coastal seawater,
27 Aodi and Nanwan Bay sites, Taiwan (Jiang et al., 2011), in the California coastal waters
28 during the day time (DeGrandpre et al., 1998), in reef crest algal community (Ishigaki
29 Island) Japan (Nakamura and Nakamori, 2009), in reef crest algal community (Ishigaki
30 Island) Japan (Nakamura and Nakamori, 2009), in branching Acropora (coral)

1 community, Ishigaki Island, Japan (Nakamura and Nakamori, 2009), in montiporastellata
2 (coral) community, Ishigaki Island, Japan (Nakamura and Nakamori, 2009), and in
3 Rukan-sho coral reef, Ishigaki Island, Japan (Ohde and van Woesik, 1999). The inner
4 shelf/coastal waters are more likely to act as sinks of atmospheric CO₂ during daytime
5 (Chen and Borges, 2009; Zhai et al., 2005).

6 The optimum conditions for such a situation to be found are associated with high
7 primary production and require a sufficient amount of nutrients and light (Behrenfeld et
8 al., 2002). For example, daytime sinking of atmospheric CO₂ can be observed in: in the
9 California coastal waters during the day time (DeGrandpre et al., 1998); in the coastal
10 seawater of the Bay of Bengal in sites (site 3) that are sufficiently rich in nutrients (the
11 same trend is not followed in nutrient-deficient sites) (Akhand et al., 2013), and during
12 the summer period (April-August) in subtropical eutrophic lakes such as Lake Hongfeng
13 and Lake Baihua (Wang et al., 2011). The latter two lakes have both a sufficient amount
14 of nutrients and have high water temperature (*e.g.* 9-26 °C during the whole year), as a
15 consequence of high sunlight intensity (Li et al., 2013; Long et al., 2013).

16 The balance between the daytime sinking of atmospheric CO₂ and the nighttime
17 emissions to the atmosphere may be variable, depending on the overall conditions and the
18 relative importance of photosynthesis and respiration. For instance, in the equatorial
19 Pacific Ocean it has been shown that the partial pressure of CO₂ (*p*CO₂) decreased from
20 430 µatm to 340 µatm while the sea surface temperature increased from 25.4°C to 28.3°C
21 (Goyet and Peltzer, 1997), a possible reason being elevated photosynthesis. In the highly
22 productive Seto Inland Sea, carbon dioxide is released to the atmosphere between June
23 and November, whilst CO₂ uptake by seawater occurs between December and May
24 (Ohtaki et al., 1993).

25 **2.1.2 Emission or balance of CO₂ to the atmosphere during daytime, and sinking**
26 **or balance of atmospheric CO₂ to surface water during the night.** This scenario is
27 typically observed in waters with low photosynthetic activity, where there is limited
28 consumption of CO₂ by photosynthesis during the day and limited degradation/respiration
29 of organic matter in the night time (Fransson et al., 2004; Kuffner et al., 2008; Yates et
30 al., 2007; Zhai et al., 2005). Such conditions are typically observed in waters that are

1 characterized by a low level of nutrients or sunlight irradiation, and/or in the presence of
2 temporary factors such as sunlight limitation because of cloudy weather or precipitation.
3 The fluxes of CO₂ between the atmosphere and seawater could depend for instance on the
4 water temperature, as more elevated daytime temperature would decrease the solubility of
5 CO₂ in water and the reverse would happen during the night. The daytime emission of
6 CO₂ from seawater would also be favored by the respiration/degradation of organic
7 matter. Such a scenario is often observed in waters where the primary production is
8 limited by the availability of nutrients, even in the presence of relatively elevated
9 amounts of organic carbon. These conditions can be found in the Jiulongjiang estuary
10 (Fig. 2a), in the Tampa and Florida Bay (Yates et al., 2007); in the northern South China
11 Sea (Dai et al., 2009), in the salt marshes and mangroves (Chen and Borges, 2009), in the
12 outer shelf/open ocean (Chen and Borges, 2009; Zhai et al., 2005) and in the Coral reef
13 (mesocosm), Moku O Loe, Kaneohe Bay, Hawaii (Kuffner et al., 2008). The availability
14 of nutrients, with typically seasonal time scales, influences photosynthesis under
15 conditions of elevated illumination in the surface layer (light saturation). In contrast, in
16 deeper waters photosynthesis responds on a shorter time scale to variations of incoming
17 irradiance, light attenuation and mixing layer depth (Behrenfeld et al., 2002).

18 **2.1.3 Emission or balance of seawater CO₂ to the atmosphere during both day and**
19 **night.** This scenario (Borges and Frankignoulle, 2003; Goyet and Peltzer, 1997; Kayanne
20 et al., 2005; Nakamura and Nakamori, 2009) is observed when CO₂ is produced by
21 degradation/respiration of organic matter (and also by abiotic photodegradation in
22 daytime), and when photosynthesis is unable to uptake the generated carbon dioxide. The
23 consequence is an increase of seawater *f*CO₂ and a subsequent release of the excess CO₂
24 to the atmosphere. Typical reasons for this scenario may be lack or scarcity of nutrients
25 and/or light, the latter because of *e.g.* cloudy weather. Such water characteristics may be
26 found in the Bay of Seine (Borges and Frankignoulle, 2003), in inner estuaries and near-
27 shore coastal areas, or in low-latitude shelves (Chen and Borges, 2009), in the coastal
28 seawater (site 3) of the Bay of Bengal (Akhand et al., 2013), in the equatorial Pacific
29 Ocean that linked to day-time increase in temperature (Goyet and Peltzer, 1997), mostly
30 source in the Southern Ocean surface water at sites WIE and APF2 (Fransson et al.,
31 2004), in coral reef in Ishigaki Island, Japan in summer (Kayanne et al., 2005), and in

1 *pocilloporaverrucosa-tabular* Acropora (coral) community, Ishigaki Island, Japan
2 (Nakamura and Nakamori, 2009).

3 **2.1.4 Sinking or balance of atmospheric CO₂ to surface water during both day and**
4 **night.** The main difference between this scenario and scenario #2 is that, in the present
5 case, daytime photosynthesis occurs to a higher extent and it is able to deplete seawater
6 CO₂, thereby causing a net uptake of carbon dioxide by seawater. Moreover, the organic
7 material generated during the day is not degraded to a sufficient extent in nighttime to
8 make the seawater *f*CO₂ increase above the atmospheric level. As in the case of scenario
9 #2, such conditions can be observed in water where the primary production is limited by
10 nutrients and/or light. These characteristics can be found in open shelves in temperate and
11 high-latitude regions, during all seasons (Chen and Borges, 2009), in the Belgian coastal
12 seawater (Borges and Frankignoulle, 1999), in surface waters of the Galician upwelling
13 system (Borges and Frankignoulle, 2001), mostly sinking in the Southern Ocean surface
14 water at site SIE1 (Fransson et al., 2004), in the surface seawater (east of Gotland) in the
15 Baltic Sea (Wesslander et al., 2011), and mostly sinking in coral reef in Ishigaki Island,
16 Japan in winter (Kayanne et al., 2005).

17 **2.1.5 Sinking or source or balance of atmospheric CO₂ to surface water during the**
18 **warm period.** This scenario generally takes place in highly productive and nutrient-rich
19 waters (Arrigo et al., 2008; Borges and Frankignoulle, 1999; Thomas et al., 2005; Zhai
20 and Dai, 2009; Zhai et al., 2005). During summer CO₂ is consumed by the elevated
21 photosynthesis, which takes place under optimum conditions as far as both nutrients and
22 light are concerned (Arrigo et al., 2008). Degradation/respiration of organic matter does
23 not keep pace with photosynthesis, and the decrease of seawater *f*CO₂ causes a net uptake
24 of carbon dioxide from the atmosphere. During the winter season, the reduction of solar
25 intensity and the shorter day-time ensure that respiration/degradation prevail over
26 photosynthesis, thereby increasing the *f*CO₂ in water at levels comparable to or above
27 those of the atmosphere. Under such circumstances, emission of carbon dioxide by
28 seawater can be observed. Wintertime emission of CO₂ to the atmosphere is generally
29 found in the South China Sea (Zhai et al., 2009; Zhai et al., 2005; Zhai et al., 2013), in
30 the East China Sea (Shim et al., 2007; Zhai and Dai, 2009), in the Bay of Bengal (Biswas
31 et al., 2004), in coral reef in Ishigaki Island, Japan (Kayanne et al., 2005), in a few

1 sampling sites in North Sea (Thomas et al., 2004), in the Southern Bight of the North Sea
2 (Schiettecatte et al., 2007), in the English Channel and adjacent seawater (Borges and
3 Frankignoulle, 2003), in the Scheldt plume off the Belgian coast (Borges and
4 Frankignoulle, 2002), in the seawater of Northwestern Mediterranean Sea (Copin-
5 Montégut et al., 2004), in the mid and outer shelf of US Middle Atlantic Bight
6 (DeGrandpre et al., 2002), in the coastal upwelling system (Gago et al., 2003), mostly
7 sinking in the Gulf of Biscay (Frankignoulle and Borges, 2001), mostly source in the
8 Kaneohe Bay, Oahu, Hawaii (Fagan and Mackenzie, 2007), mostly sinking in the
9 seawater of the Gotland Sea (Kuss et al., 2004; Schneider et al., 2003), mostly sinking in
10 the Chukchi Sea adjacent to the Arctic Ocean (Bates, 2006), mostly source in the
11 Antarctic shelf waters of Southern Ocean (Arrigo et al., 2008), mostly source in the
12 seawater of the Bermuda coral reef system (Bates, 2002), mostly sinking in the Chukchi
13 Sea adjacent to the Arctic Ocean along 65–75°N and 150–170°W (Bates, 2006), in the
14 Southern Ocean (Arrigo et al., 2008), in the Ross Sea (Sweeney, 2003), and strong
15 sinking during the summer period than in winter in the Gulf of Biscay (Frankignoulle and
16 Borges, 2001).

17 **2.1.6 Emission or sinking or balance of seawater CO₂ to the atmosphere during**
18 **cold period.** This scenario is similar to scenario #5 and takes place in the similar types of
19 seawaters (Bates, 2006; Borges and Frankignoulle, 2003; Thomas et al., 2004). Such a
20 scenario is observed in a few sampling sites of North Sea (Thomas et al., 2004), in the
21 Southern Bight of the North Sea (Schiettecatte et al., 2007), in the English Channel and
22 adjacent seawater (Borges and Frankignoulle, 2003), in the South China Sea (Zhai et al.,
23 2013), in the coastal seawater in the Bay of Bengal (Biswas et al., 2004) (Biswaste al.,
24 2004), in the seawater of Northwestern Mediterranean Sea (Copin-Montégut et al., 2004),
25 in the mid and outer shelf of US Middle Atlantic Bight (DeGrandpre et al., 2002), mostly
26 source in the seawater of the Gotland Sea (Kuss et al., 2004; Schiettecatte et al., 2007),
27 mostly source in the Scheldt plume off the Belgian coast (Borges and Frankignoulle,
28 2002), mostly source in the upper reach of the Pearl River Estuary near South China Sea
29 (Dai et al., 2006), in the Kaneohe Bay, Oahu, Hawaii (Fagan and Mackenzie, 2007),
30 mostly sinking in the Antarctic shelf waters of Southern Ocean (Arrigo et al., 2008),
31 mostly sinking in the seawater of the Bermuda coral reef system (Bates, 2002), mostly

1 sinking in the Chukchi Sea adjacent to the Arctic Ocean along 65–75°N and 150–170°W
2 (Bates, 2006), mostly sinking in East China Sea (Zhai and Dai, 2009), mostly sinking in
3 the Atlantic sector, the Southern Ocean (Chierici et al., 2004), and mostly sinking in coral
4 reef in Ishigaki Island, Japan (Kayanne et al., 2005).

5 Overall, the subtropical-tropical marginal seas and the continental shelves at low
6 latitudes and the equatorial Pacific (14°N–14°S) act as sources of CO₂ to the atmosphere
7 whilst the marginal seas and continental shelves at mid-high latitudes and the temperate
8 oceans between 141 and 501 in the both hemispheres acts as a strong sink of atmospheric
9 CO₂ (Borges, 2005; Borges et al., 2005; Cai et al., 2006; Chen and Borges, 2009;
10 Takahashi et al., 2009). Correspondingly, coral reefs act either as a source of CO₂ to the
11 atmosphere (Bates, 2002; Gattuso et al., 1999; Ware et al., 1992) or a oceanic sink (Bates,
12 2002; Gattuso et al., 1997; Kraines et al., 1996). But such conclusions might not be
13 precisely true in a variety of seawater that sampled once during the day time at the same
14 location. Six different scenarios of CO₂ in seawaters configured that seawater can act as a
15 source or sink or balance of CO₂ to the atmosphere which also significantly varied with
16 seasons affecting many factors (see the main text). This therefore indicates that the
17 diurnal variation of CO₂ is very important for better understanding of the ocean
18 acidification as well as CO₂ behavior in the world's oceans, and this issue should be the
19 focus for further study.

20 **2.3 Enhanced PP and respiration due to the effects of global warming and other** 21 **processes** (Some remaining contents of this section)

22 The enhancement of photosynthesis and primary production depends on several
23 factors, and those which prevail are linked to the nutrient status of water. Photosynthesis
24 is usually favored by enhanced input of terrestrial DOM into water (Bauer et al., 2013;
25 Porcal et al., 2009), but in DOM-rich waters the regeneration of autochthonous DOM and
26 nutrients from PP plays a more important role. In the presence of sufficient light or high
27 WT under GW conditions, such factors can also lead to harmful algal blooms (Fig. S1)
28 (Mostofa et al., 2013a). A link between DOM or PP and algal blooms or eutrophication
29 may be observed when the regeneration of nutrients (NH₄⁺, NO₃⁻ and NO₂⁻) is
30 operational (Cai et al., 2011; Mostofa et al., 2013a). In contrast, transformation of

1 nitrogen-containing nutrients to N₂ (Doney et al., 2007; Kuypers et al., 2003; Ward,
2 2013) largely eliminates the connection (Mostofa et al., 2013a).

3 The primary production and the occurrence of DOM in seawater are linked to
4 several other processes, including: (i) the increase in atmospheric CO₂ (Feng et al., 2009;
5 Gruber et al., 2002; Hein and Sand-Jensen, 1997; Le Quéré et al., 2009; Schippers et al.,
6 2004); (ii) the increased input of terrestrial OM and nutrients, because of human-induced
7 alteration of land use (Cai et al., 2006; Smith et al., 1999; Vitousek et al., 1997) and
8 because of enhanced soil erosion rates due to extreme climate weathering (Nearing et al.,
9 2004); (iii) the input of organic contaminants and nutrients from anthropogenic activities
10 (e.g. sewerage, industries, pharmaceuticals, or intensive fish cage farming (Fu et al.,
11 2010; Mostofa et al., 2013a; Nyenje et al., 2010; Pergent - Martini et al., 2006) ; (iv)
12 large ecological changes in forests (Graven et al., 2013) and transportation of OM by
13 weathering, combined with high terrestrial plant productivity induced by increasing
14 atmospheric CO₂ (Regnier et al., 2013); (v) atmospheric acid rain/precipitation (notably
15 HNO₃ and H₂SO₄) (Baker et al., 2007; Teira et al., 2013) and (vi) the effects of GW (Fig.
16 S1) (Engel et al., 2012; Feng et al., 2009; Hein and Sand-Jensen, 1997; Mostofa et al.,
17 2013b; Schippers et al., 2004). Many of the described phenomena are mostly susceptible
18 to affect coastal waters which, by the way, host between 15% and 30% of the oceanic
19 primary production and 80% of the OM burial (Gattuso et al., 1998). Coastal waters also
20 account for most of the benthic oceanic CaCO₃ production, for 20% of the surface
21 pelagic CaCO₃ stock (Balch et al., 2005), and for 50% of the oceanic CaCO₃ deposition
22 (Gattuso et al., 1998).

23 Some of the above-mentioned factors which affect primary production in seawater
24 are also closely interlinked. For instance, the increase of atmospheric temperature due to
25 GW could enhance DOM leaching from terrestrial soils because of high soil respiration,
26 and the DOM contents in surface waters could increase as a consequence (Porcal et al.,
27 2009). Increased atmospheric CO₂ would also increase the turnover rates of soil OM,
28 driven by the enhanced activity of soil microorganisms (Blagodatskaya et al., 2010) and
29 including a higher rate of decomposition of older and fresh plant residues (Blagodatskaya
30 et al., 2010; Dorodnikov M, 2011). Note that elevated CO₂ was shown to enhance soil

1 OM mineralization by 83–218% in a simulated wetland (Wolf et al., 2007). A net
2 increase of C emissions to the atmosphere would thus be induced by enhanced respiration
3 of soil OM under global warming condition (Davidson and Janssens, 2006; Knorr et al.,
4 2005).

5 The increase of net primary production in seawater is responsible for the gradual increase
6 of CO₂ in the surface seawater of some oceans, including the ice-free Arctic Ocean Basin
7 (Cai et al., 2010), the area near the Canary Islands (ESTOC), located in the northeast
8 Atlantic subtropical gyre (González-Dávila et al., 2003), and in the Northern Hemisphere
9 (Graven et al., 2013). Enrichment of surface seawater in CO₂ means that oceanic water
10 behaves less effectively as a sink and could even become a net CO₂ source to the
11 atmosphere, as is presently the case for the equatorial Pacific (Takahashi et al., 2009).
12 Average rates of increase in surface-water *f*CO₂ have been quantified, on the basis of
13 deseasonalized data, as 1.5 μatm y⁻¹. The basin- specific rates varied between 1.2±0.5
14 and 2.1±0.4 μatm y⁻¹ over the North Atlantic, North and South Pacific and Southern
15 Oceans, which cover about 27% of the global ocean areas (Takahashi et al., 2002;
16 Takahashi et al., 2009). Furthermore, the role of low-latitude continental shelves as net
17 sources of CO₂ to the atmosphere is thought to be a consequence of both GW that
18 increases WT and to higher inputs of terrestrial organic carbon (Cai et al., 2006; Corbière
19 et al., 2007). The effect of GW on the release of CO₂ to the atmosphere is related to a
20 series of processes, including lower solubility of CO₂ in warmer waters, increased
21 stratification and the effects on primary productivity (Keeling, 2005). Coastal waters are
22 also more susceptible to acidification than the open ocean, because of the effects of
23 eutrophication (such as emission of CO₂ by respiration and mineralization of organic N
24 and P) that decreases the buffering capacity of seawater. The combination of increasing
25 atmospheric CO₂ and of processes occurring in seawater produces different critical
26 scenarios, which are observed for CO₂ emissions from seabed and for sinking of
27 anthropic atmospheric CO₂ to seawater (see also supplementary section) (Borges and
28 Gypens, 2010; Cai et al., 2011; Dore et al., 2009). Moreover, changes in the biogenic
29 carbon flow in response to sea surface warming have the potential to reduce the transfer
30 of primary produced OM to higher trophic levels (Laws et al., 2000; Vázquez-
31 Domínguez et al., 2007; Wohlers et al., 2009). Such processes would weaken the ocean's

1 biological carbon pump and would provide a positive feedback to the rise of atmospheric
2 CO₂ (Laws et al., 2000; Vázquez-Domínguez et al., 2007; Wohlers et al., 2009).

3 **2.4 Impacts of nitrification-denitrification and sulfate reduction on acidification**

4 Nitrification ($\text{NH}_4^+ \rightarrow \text{NO}_2^- \rightarrow \text{NO}_3^-$ or $\text{NH}_4^+ + 2 \text{O}_2 \rightarrow \text{NO}_3^- + 2 \text{H}^+ + \text{H}_2\text{O}$)
5 (Doney et al., 2007) caused by ammonia-oxidizing bacteria, such as *Nitrosomonas* and
6 *Nitrobacter* (Martens-Habbena et al., 2009) can contribute to acidification, and it is for
7 instance partially responsible for the acidification of coastal seawater (Doney et al.,
8 2007). Denitrification ($\text{NO}_3^- \rightarrow \text{NO}_2^- \rightarrow \text{NO} \rightarrow \text{N}_2\text{O} \rightarrow \text{N}_2$), caused by denitrifying
9 bacteria (Pan et al., 2013) is an important process that occurs in oxygen minimum zones
10 (OMZs) and accounts for approximately 71% of the nitrogen loss, while anaerobic
11 ammonium oxidation (anammox) accounts for the remaining 29% (Ward, 2013). An
12 important issue is that increasing atmospheric CO₂ is expected to significantly modify the
13 occurrence of microorganisms in oceanic water and to favor global N₂ fixation and
14 possibly denitrification, probably at the expense of nitrification (Hutchins et al., 2013).
15 Denitrification stimulates the acidification by producing an effective total alkalinity flux
16 in seawater (Chen and Wang, 1999; Thomas et al., 2009). However, in addition to the
17 processes already shown, one should consider that ammonium and nitrate also take part
18 to additional cycles. It has been shown that NO₃⁻-supported PP generates alkalinity (106
19 $\text{CO}_2 + 16 \text{NO}_3^- + 138 \text{H}_2\text{O} \rightarrow \text{organics} + 16 \text{OH}^- + 138 \text{O}_2$), but the resulting basification
20 would only partially compensate for acidification if nitrate is generated by the oxidation
21 of ammonium according to the reaction $2\text{NH}_3 + 4\text{O}_2 \rightarrow 2\text{NO}_3^- + 2\text{H}_2\text{O} + 2\text{H}^+$; note that
22 ammonium oxidation to nitrate yields 2 H⁺ per processed NH₄⁺, while incorporation of
23 nitrate into organic nitrogen only consumes one H⁺ per nitrate ion processed. On the
24 contrary, NH₄⁺-supported PP eliminates alkalinity according to the overall reaction: 106
25 $\text{CO}_2 + 16 \text{NH}_4^+ + 106 \text{H}_2\text{O} \rightarrow \text{organics} + 16 \text{H}^+ + 106 \text{O}_2$ (Doney et al., 2007). On the
26 other hand, anaerobic ammonium oxidation ($\text{NH}_4^+ + \text{NO}_2^- \rightarrow \text{N}_2\text{H}_4 \rightarrow \text{N}_2$) carried out by
27 anammox bacteria (Ward, 2013) is H⁺-neutral ($\text{NH}_4^+ + \text{NO}_2^- \rightarrow \text{N}_2 + 2 \text{H}_2\text{O}$).

28 In addition to potentially affecting the pH of seawater, the nitrogen cycle
29 processes are also substantially influenced by pH itself. For instance, nitrification rates

1 decrease to zero at a pH around 6.0–6.5, where the NH₃ substrate is practically all
2 protonated to ammonium (Huesemann et al., 2002). In fact, the proportions of NH₃ to
3 NH₄⁺ and of PO₄³⁻ to HPO₄²⁻ are both very sensitive to small pH variations around 8.0
4 according to the following equations

$$5 \quad \frac{[NH_3]}{[NH_4^+]} = \frac{k_a}{[H^+]} \quad \text{eq. 1}$$

$$6 \quad \frac{[PO_4^{3-}]}{[HPO_4^{2-}]} = \frac{k_{a1} k_{a2} k_{a3}}{[H^+] k_{a1} k_{a2}} \quad \text{eq. 2}$$

7 note that pK_a for NH₄⁺ is 9.3 and the three pK_a s for H₃PO₄ are 2.2, 7.1 and 12.4,
8 respectively (Zeebe and Wolf-Gladrow, 2001). Moreover, the nitrogen cycle may be
9 closely interlinked with that of carbon. For instance, the anaerobic oxidation of CH₄ to
10 CO₂ is connected either to the reduction of NO₃⁻ to NO₂⁻ and then N₂, or to anammox
11 that yields N₂ as well (Haroon et al., 2013). The conversion of SO₄²⁻ into H₂S, carried out
12 by sulfate-reducing bacteria such as desulfobacterales, desulfovibrionales and
13 syntrophobacterales (Muyzer and Stams, 2008) which can either be emitted to the
14 atmosphere or be buried as pyrite, also significantly increases the total alkalinity (Chen
15 and Wang, 1999). Therefore, denitrification and SO₄²⁻ reduction can raise the buffering
16 capacity of CO₂ that stimulates the acidification of seawater (Chen and Wang, 1999;
17 Thomas et al., 2009). Note that each mole of denitrified N and reduced SO₄²⁻ reduction
18 can release 0.99 and 1.98 mol of total alkalinity, respectively (Chen and Wang, 1999).

19 **2.5 Direct acidification and stimulation of PP by atmospheric acid rain**

20 Atmospheric acid rain can decrease the pH of seawater directly, by the input of HNO₃
21 and H₂SO₄, and indirectly through the addition of NH₄⁺ (Bates and Peters, 2007; Doney
22 et al., 2007). In fact, most of the anthropogenic NH₃ and NH₄⁺ (~ 98%) are nitrified to
23 nitrate in the upper ocean releasing H⁺ ions (NH₄⁺ + 2O₂ → NO₃⁻ + 2H⁺ + H₂O; see also
24 the previous section). This issue causes the effective net atmospheric input of ammonium
25 to be acidic (Doney et al., 2007). A rough estimate shows that direct atmospheric acidic

1 depositions could contribute 2% to the acidification of surface waters in the subtropical
2 North Atlantic Ocean since the early 1980's (Bates and Peters, 2007).

3 Another important impact of rainfall in general, including the acidic one is the
4 stimulating effect on primary production, as shown by field observations and
5 experimental simulations (Baker et al., 2007; Paerl and Fogel, 1994; Teira et al., 2013).
6 As discussed earlier, the enhanced primary production can contribute to seawater
7 acidification through increased CO₂. Note that atmospheric rainfall can bring nutrients to
8 seawater, including NO₃⁻, NH₄⁺ and PO₄³⁻ (Doney et al., 2007; Paerl and Fogel, 1994),
9 Fe²⁺ and Fe³⁺ (Baker et al., 2007; Kieber et al., 2001), dissolved Si (Baker et al., 2007)
10 and other major ions such as F⁻, Cl⁻, SO₄²⁻, K⁺, Na⁺, Ca²⁺, Mg²⁺, Co²⁺ and Zn²⁺ (Baker et
11 al., 2007; Zhang et al., 2007). Other compounds contained in rainwater that could
12 stimulate photosynthesis under definite conditions are H₂O₂, aldehydes and organic acids
13 (Sakugawa et al., 1993), as well as other dissolved organic substances (Kieber et al.,
14 2006). *In situ* incubation experiments demonstrate that chlorophyll *a* concentration is
15 increased by 2.6 times with the addition of 10% (v/v) rainwater (Zou et al., 2000).
16 Microcosm experiments on the effect of rainwater addition to natural marine plankton
17 populations also showed an increase in bacterial abundance and production (Teira et al.,
18 2013). Another issue is that rainfall on land can produce an important input of OM and
19 nutrients in seawater *via* terrestrial runoff. In addition to the possible stimulation of
20 acidification processes, the direct or indirect (through runoff) fertilization of seawater
21 carried out by rain could contribute to the occurrence of harmful algal blooms and to the
22 generation of pathogens (Doney et al., 2007; Flewelling et al., 2005; Sekar et al., 2006;
23 Sunda and Cai, 2012; Teira et al., 2013).

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1 **Figure captions**

2 Fig. S1

3 A conceptual schematic diagram about the response to global warming effects and high
4 organic matter (OM) including DOM and PP, as well as the possible effects on the
5 occurrence of harmful algal blooms in water. First, photoinduced transformation is the
6 primary step for the generation of the oxidizing species (*e.g.* H₂O₂ and HO[•]) from DOM.
7 Such processes, along with biological ones, can drive respiration/degradation of DOM
8 and dead organisms, yielding a variety of intermediates and byproducts among which
9 CO₂, DIC, and nutrients (NO₃⁻, PO₄³⁻). Second, such products can enhance
10 photosynthesis with a subsequent increase of the PP (phytoplankton). Third, PP further
11 induces photoinduced and microbial respiration that releases autochthonous DOM and
12 nutrients. Fourth, autochthonous DOM and nutrients undergo photoinduced and
13 microbial degradation that further yields CO₂, DIC and again nutrients. Increased
14 temperature following global warming and input of high OM in water can favor the
15 occurrence of high photosynthesis for a longer period of time and, as a consequence,
16 enhance primary production. This issue can lead to an increase of the worldwide
17 incidence of harmful algal blooms, in waters with high contents of DOM and PP.

18 Fig. S2

19 Monthly changes in pH, dissolved organic carbon (DOC) and chlorophyll *a* (Chl *a*) at
20 different water depths (2.5, 10, 20, 40 and 70-80 m) in Lake Biwa, the 11th largest lake in
21 the world with surface area of approximately 674.8 km². Samples were collected monthly
22 during the period of April 1999 to February 2001 at day-time from the northern basin,
23 where the maximum water depth is 104 m (Mostofa et al., 2005). The values of pH and of
24 the concentrations of DOC and Chl *a* gradually increased, reaching a maximum on July
25 (1999 and 2000) for pH, on September (1999) or July (2000) for DOC and on November
26 (1999) or October (2000) for Chl *a*.

27 Fig. S3

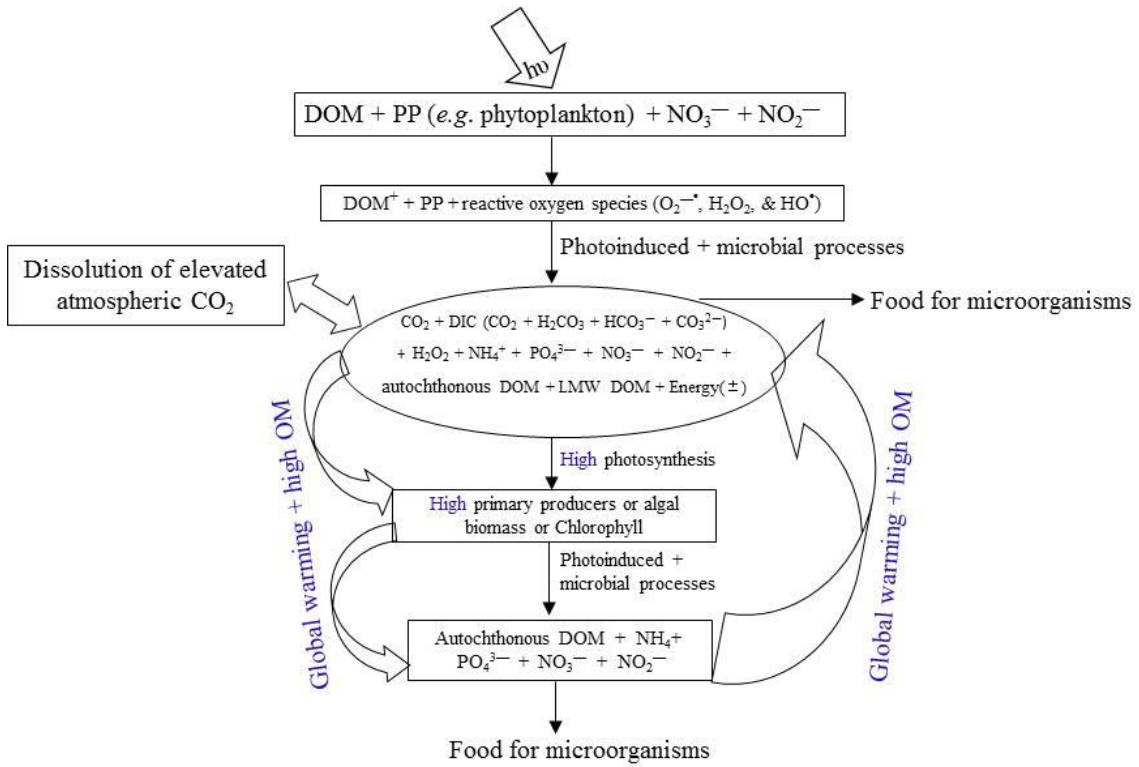
28 A substantial decline in dissolved O₂ contents in August (dashed lines) compared to June
29 (solid lines) in the seawater of the Yellow Sea. The same symbol indicates the same
30 sampling site for the samples collected during June and August.

31 Fig. S4

32 Relationship between pH and dissolved O₂ for a variety of the subsurface seawaters
33 collected from the Yellow Sea in 2012 and 2013. Experiments were carried out on July 2,
34 2013 using subsurface water collected at 37 m depth from the East China Sea (see the
35 caption of Fig. 4 for a detailed description).

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1 Fig. S1



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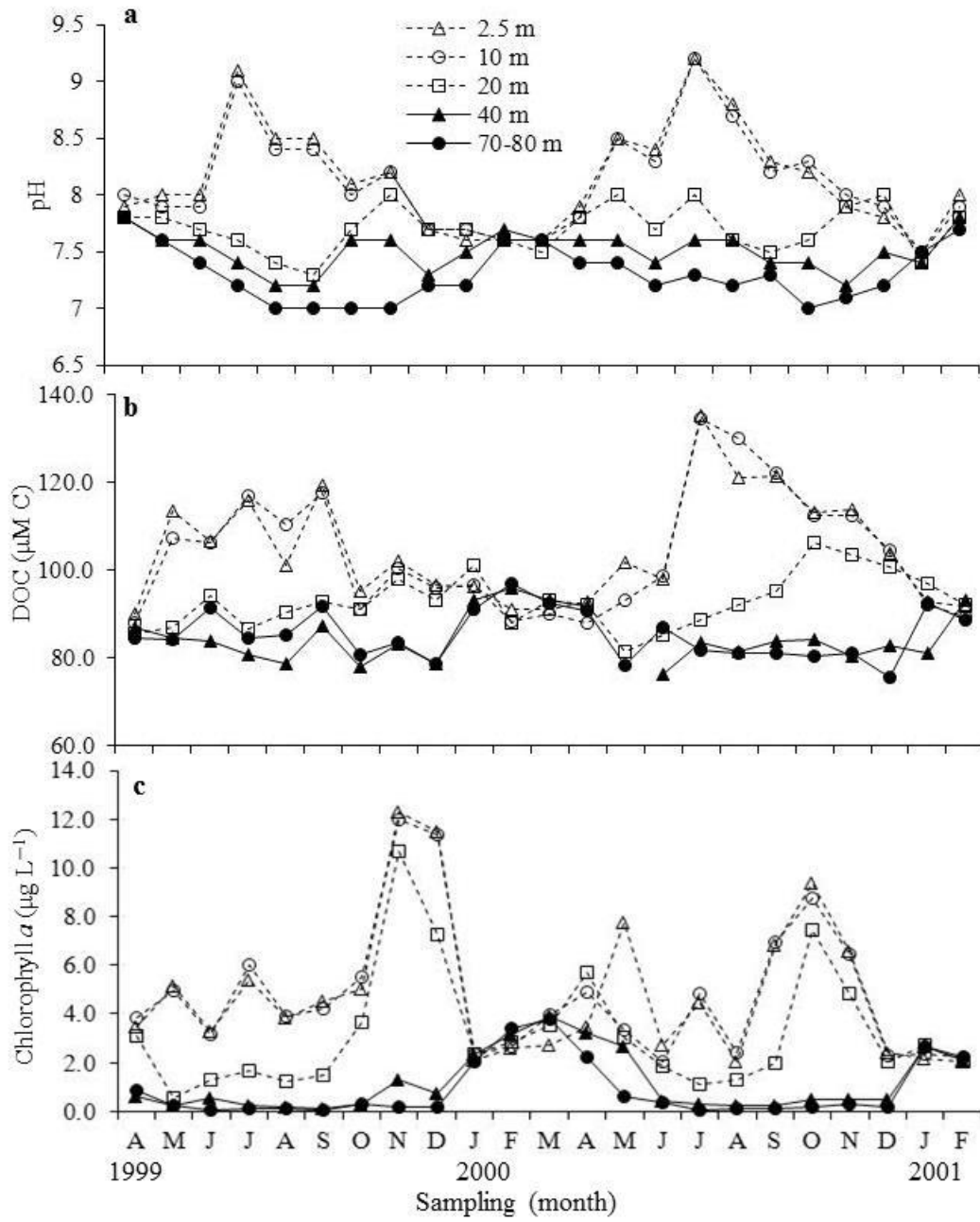
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1 Fig. S2



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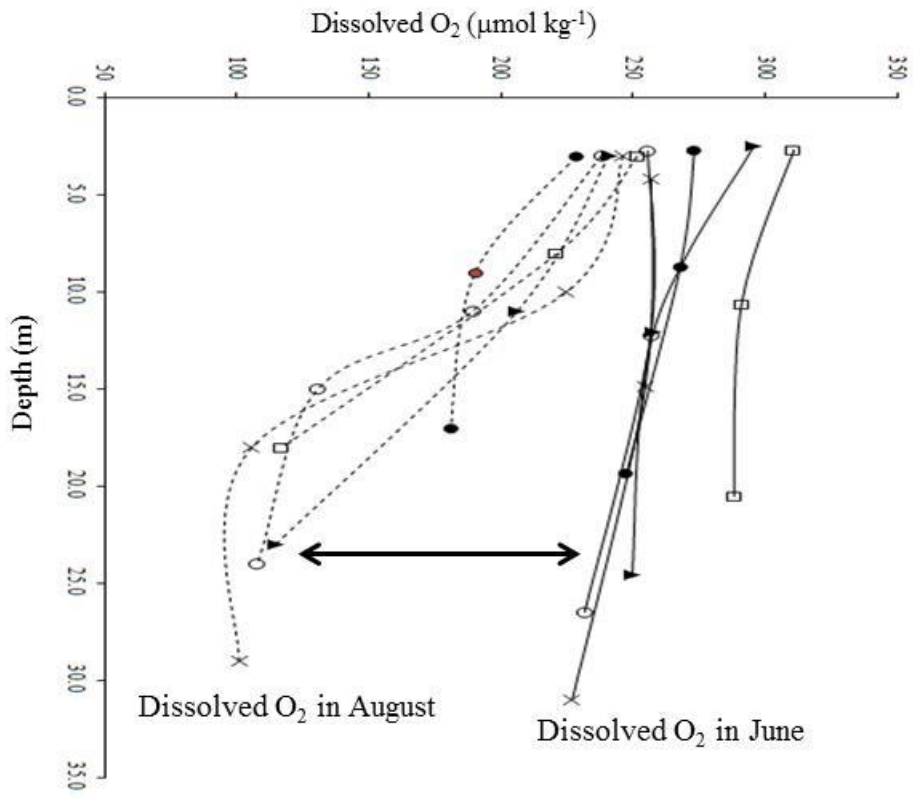
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1 Fig. S3



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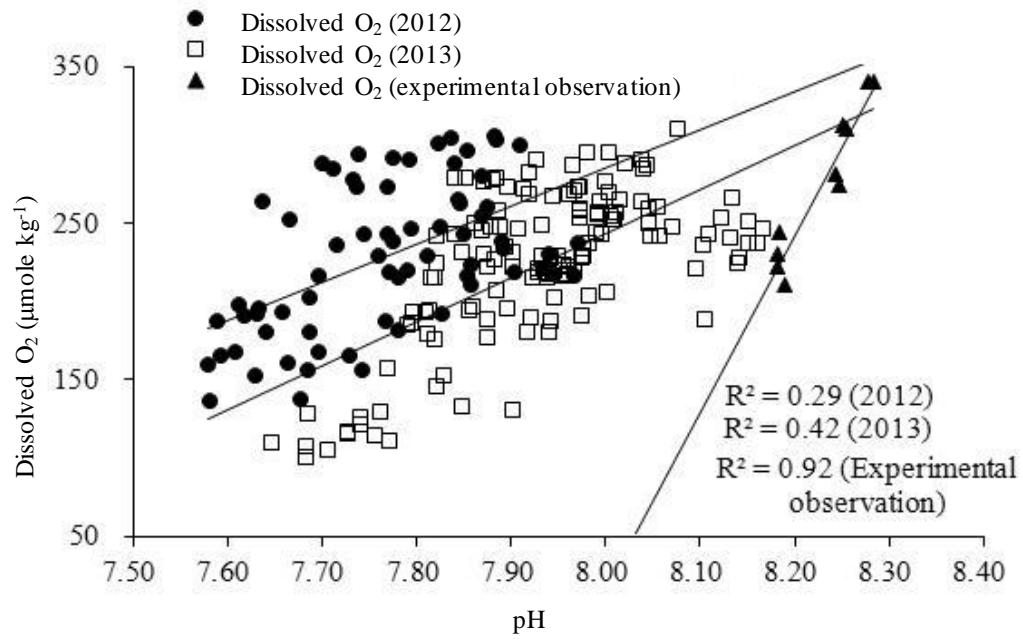
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1 Fig. S4

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Table S1. Ranges of pH, PCO_2 , DIC, chlorophyll a (Chl a), dissolved O_2 , hydrogen peroxide (H_2O_2) and water temperature (WT) in world's oceans.

Sampling	pH		$P CO_2$ (μ atm)	DIC (μ mol kg^{-1})	Chl a ($\mu g L^{-1}$)	Dissolved O_2		H_2O_2 (nM)	WT	References
	Surface water	Anoxic zone				Surface water	Anoxic zone			
	(0-20 m)					($mg L^{-1}$)				
Okinawa Island Seawater (Taira Bay, November)	8.02–8.79 (diurnal)	-	-	-	-	-	-	40-160	21.7-23.7	Arakaki et al., 2005
Okinawa Island Seawater (Taira Bay, January)	8.16–8.25 (diurnal)	-	-	-	-	-	-	40-100	18.8-20.9	Arakaki et al., 2005
Okinawa Island Seawater (Sesoko Island, January)	7.82–8.29 (diurnal)	-	-	-	-	-	-	30-110	17.7-20.2	Arakaki et al., 2005
Nanwan Bay, south coast of Taiwan (5-13 June)	8.04-8.24 (diurnal)	-	227-532	1830-2050	0.20-1.10	2.62-5.74	-	-	22.4-28.4	Jiang et al., 2011 ⁺
Coastal seawater, Aodi, northeast of Taiwan (1-5 August)	7.87-8.35 (diurnal)	-	178-582	1750-2090	0.30-1.30	2.70-6.56	-	-	25.0-28.0	Jiang et al., 2011 ⁺
Coastal seawater (Bay of Bengal, Stn 3)	8.12-8.37 (diurnal)	-	153-373	-	12.35±2.23	-	-	-	28.50-31.70	Akhand et al., 2013
Marine bathing waters (Southern California, USA)	~7.37-8.18 (diurnal)	-	-	-	-	-	-	5-370	18.6-22.5	Clark et al., 2010 ⁺
Inner shelf of southern Monterey Bay, California coast	7.7-8.1 (diurnal)	-	-	-	-	2.0-4.6	-	-	9.5-13.5	Booth et al., 2012 ⁺
Tampa Bay	7.95-8.24 (diurnal)	-	262-580	-	-	-	-	-	26.2-28.7	Yates et al., 2007
Florida Bay	8.17-8.39 (diurnal)	-	260-497	-	-	-	-	-	23.8-27.3	Yates et al., 2007
Belgian coastal seawater (near Zeebrugge)	8.314-8.358 (diurnal)	-	190-214	-	12.5-23.5	-	-	-	9.74-10.15	Borges and Frankignoulle, 1999 ⁺
Seawater, Jarvis Island	7.985-8.019 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Great Barrier Reef, Lizard Island (subsurface: within 5 cm of the surface)	7.93-8.23 (diurnal)	-	-	-	-	-	-	-	-	Gagliano et al., 2010
Great Barrier Reef, Lizard Island (<i>Pocillopora damicornis</i>)	7.93-8.30 (diurnal)	-	-	-	-	-	-	-	-	Gagliano et al., 2010
Great Barrier Reef, Lizard Island (<i>D. perspicillatus</i> algal garden)	7.95-8.25 (diurnal)	-	-	-	-	-	-	-	-	Gagliano et al., 2010
Great Barrier Reef, Lizard Island (open sand: within 3 cm of the sand)	7.80-8.24 (diurnal)	-	-	-	-	-	-	-	-	Gagliano et al., 2010
Great Barrier Reef, Heron Island	7.69-8.44 (diurnal)	-	-	-	-	0.46-7.59	-	-	22.2-32.0	Santos et al., 2011
Luhuitou fringing reef, Sanya Bay, South China Sea (winter)	~7.95-8.07 (diurnal)	-	~380-500	-	-	~5.0-8.0	-	-	24.5±0.5	Zhang et al., 2013 ⁺
Luhuitou fringing reef, Sanya Bay, South China Sea (spring)	~7.96-8.14 (diurnal)	-	~290-460	-	-	~5.5-8.0	-	-	33.8±0.1	Zhang et al., 2013 ⁺
Luhuitou fringing reef, Sanya Bay, South China Sea (summer)	~7.86-8.07 (diurnal)	-	~350-605	-	-	~3.5-7.2	-	-	29.5±0.6	Zhang et al., 2013 ⁺
Luhuitou fringing reef, Sanya Bay, South China Sea (autumn)	~7.94-8.18 (diurnal)	-	~245-495	-	-	~5.3-10.7	-	-	26.3±1.0	Zhang et al., 2013 ⁺
Seawater, Kingman Reef (site Palmyra Terrace North)	7.926-8.031 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Seawater, Kingman Reef (site Palmyra Terrace South)	7.901-8.017 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Seawater, Kingman Reef (site Palmyra Forereef North)	7.944-8.007 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Seawater, Kingman Reef (site Palmyra Forereef South)	7.979-8.007 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Seawater, Kingman Reef (site Kingman Reef)	8.021-8.028 (diurnal)	-	-	-	-	-	-	-	-	Price et al., 2012
Coral reef, Ishigaki Island, Japan	7.9-8.4 (diurnal)	-	-	-	-	-	-	-	-	Kayame et al., 2005
Coral reef, Rukan-sho, Ishigaki Island, Japan (site 1, October)	7.848-8.556 (diurnal)	-	151-900	1460-2103	-	1.81-13.37	-	-	26.4-30.2	Ohde and van Woessik, 1999
Coral reef, Rukan-sho, Ishigaki Island, Japan (site 1, September)	7.951-8.688 (diurnal)	-	68-770	1181-2095	-	-	-	-	27.3-34.2	Ohde and van Woessik, 1999
Reef crest algal community, Ishigaki Island, Japan	7.863-8.335 (diurnal)	-	210-762	-	-	-	-	-	29.84-33.19	Nakamura and Nakamori, 2009
Branching acropora (coral) community, Ishigaki Island, Japan (site 4)	8.030-8.289 (diurnal)	-	201-458	-	-	-	-	-	29.0	Nakamura and Nakamori, 2009
Branching acropora (coral) community, Ishigaki Island, Japan (site 5)	7.818-8.211 (diurnal)	-	271-770	-	-	-	-	-	27.56-30.73	Nakamura and Nakamori, 2009
Branching acropora (coral) community, Ishigaki Island, Japan (site 7)	7.801-8.197 (diurnal)	-	248-799	-	-	-	-	-	27.87-30.62	Nakamura and Nakamori, 2009
Montipora stellata (coral) community, Ishigaki Island, Japan (site 8)	7.773-8.387 (diurnal)	-	144-919	-	-	-	-	-	28.39-33.18	Nakamura and Nakamori, 2009
Pocillopora acropora (coral) community, Ishigaki Island, Japan (site 9)	7.974-8.099 (diurnal)	-	410-569	-	-	-	-	-	30.72-31.25	Nakamura and Nakamori, 2009
Pocillopora acropora (coral) community, Ishigaki Island, Japan (site 10)	7.939-8.113 (diurnal)	-	368-593	-	-	-	-	-	28.27-29.91	Nakamura and Nakamori, 2009
Seawater in the coral reef, Heron Island, Great Barrier Reef	7.96-8.36 (diel)	-	-	1850-2230	-	3.11-7.05	-	-	-	Cyronak et al., 2013 ⁺
Uva Reef, Gulf of Chiriqui, Pacific Panamá (dry season)	7.63–8.20 (day)	-	-	-	-	-	-	-	28.8±0.19	Manzello, 2010
Uva Reef, Gulf of Chiriqui, Pacific Panamá (dry season)	7.84-8.04 (night)	-	-	-	-	-	-	-	28.4±0.30	Manzello, 2010

Table S1 (continued)

Sampling	pH		<i>P</i> CO ₂	DIC	Chl <i>a</i>	Dissolved O ₂		H ₂ O ₂	WT	References
	Surface water	Anoxic zone				Surface water	Anoxic zone			
	(0-20 m)					(µatm)	(µmol kg ⁻¹)			
Uva Reef, Gulf of Chiriqui, Pacific Panamá (wet season)	7.94-8.26 (day)	-	-	-	-	-	-	28.9±0.06	Manzello, 2010	
Uva Reef, Gulf of Chiriqui, Pacific Panamá (wet season)	7.86-8.01 (night)	-	-	-	-	-	-	28.7±0.03	Manzello, 2010	
Saboga Reef, Gulf of Panamá (dry season)	7.95-8.08 (day)	-	-	-	-	-	-	21.0	Manzello, 2010	
Saboga Reef, Gulf of Panamá (wet season)	7.95-8.23 (day)	-	-	-	-	-	-	28.6±0.11	Manzello, 2010	
Saboga Reef, Gulf of Panamá (wet season)	7.96-8.00 (night)	-	-	-	-	-	-	27.5	Manzello, 2010	
Reef, Galápagos, Eastern Pacific (cool)	7.65-7.99 (day)	-	-	-	-	-	-	25.0	Manzello, 2010	
Reef, Galápagos, Eastern Pacific (warm)	7.70-8.07 (day)	-	-	-	-	-	-	25.0	Manzello, 2010	
Fore reef (benthic), Palmyra (5.86614°N, 162.1172°W)	7.915-8.035 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Reef terrace (benthic), Palmyra (5.884°N, 162.1218°W)	7.851-8.104 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Fringing reef (benthic), Moorea (17.4803°S, 149.7989°W)	8.017-8.118 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Kingman Reef (benthic), Tropical Central Pacific, Open Ocean	8.009-8.034 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
SBC Mohawk Reef (benthic), Kelp (34.3943°N, 119.73°W)	7.700-8.244 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Coral Reef Lagoon seawater (The Indo-Pacific Oceans)	8.264-8.350	-	292-370	-	-	-	-	28.6-30.0	Suzuki and Kawahata, 1999	
Coral reef (mesocosm), Moku O Loe, Kaneohe Bay, Hawaii	7.55-8.08	-	500-1950	-	-	-	-	-	Kuffner et al., 2008	
Seagrass community, Chwaka Bay (Zanzibar, Tanzania)	7.9->8.9 (diurnal)	-	-	-	-	-	-	~27.0-33.0	Semese et al., 2009	
Seagrass community, Ishigaki Island, Japan (site 1)	7.763-8.547 (diurnal)	-	111-1085	-	-	-	-	30.87-34.32	Nakamura and Nakamori, 2009	
Seagrass community, Ishigaki Island, Japan (site 2)	7.894-8.353 (diurnal)	-	204-671	-	-	-	-	27.99-33.22	Nakamura and Nakamori, 2009	
Seagrass community (<i>C. nodosa</i>), Alfacs Bay, Mediterranean (1 m)	8.11-8.59 (diurnal)	-	-	-	-	-	-	-	Invers et al., 1997	
Seagrass community (<i>P. oceanica</i>), Alfacs Bay, Mediterranean (1 m)	8.15-8.45 (diurnal)	-	-	-	-	-	-	-	Invers et al., 1997	
Seagrass community (<i>P. oceanica</i>), Alfacs Bay, Mediterranean (4 m)	8.17-8.34 (diurnal)	-	-	-	-	-	-	-	Invers et al., 1997	
Benthos community, Ishigaki Island, Japan (site 2)	7.963-8.154 (diurnal)	-	318-540	-	-	-	-	28.66-31.61	Nakamura and Nakamori, 2009	
Macroalgal Kelp bed (inside), Morbihan Bay, Southern Ocean	8.07-9.11 (diurnal)	-	-	-	-	-	-	0.7-15.3	Delille et al., 2000	
Macroalgal Kelp bed (outside), Morbihan Bay, Southern Ocean	7.94-8.46 (diurnal)	-	-	-	-	-	-	0.7-8.5	Delille et al., 2000	
Macroalgae (outside), macroalgal mats, Mediterranean coastal lagoon	8.3-9.0 (diurnal)	-	2500-3000	-	-	-	-	-	Menéndez et al., 2001	
Macroalgae (inside), macroalgal mats, Mediterranean coastal lagoon	8.4-9.6 (diurnal)	-	2500-3000	-	-	-	-	-	Menéndez et al., 2001	
Macroalgal habitats, inshore Danish waters (0.2 m)	7.5-9.1 (diurnal)	-	-	-	-	-	-	-	Middelboe and Hansen, 2007*	
Macroalgal habitats, inshore Danish waters (0.4 m)	7.8-9.0 (diurnal)	-	-	-	-	-	-	-	Middelboe and Hansen, 2007*	
Macroalgal habitats, inshore Danish waters (0.6 m)	8.0-8.9 (diurnal)	-	-	-	-	-	-	-	Middelboe and Hansen, 2007*	
Macroalgal Kelp bed, La Jolla (pelagic)	7.970-8.229 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Southern Ocean surface water (Sta. SIE1)	~8.13-8.17 (diurnal)	-	~280-305	-	-	-	-	~(-0.05)-(-0.25)	Fransson et al., 2004*	
Southern Ross Sea under Antarctic sea ice (site: Cape Evans)	8.039-8.134 (diel, 15 m)	-	-	-	-	-	-	-	Matson et al., 2011	
Temperate Eastern Pacific (CCE1), Open Ocean	8.059-8.082 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Polar seawater (benthic), Cindercones, Antarctic Ocean	8.039-8.134 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Polar seawater (benthic), Cape Evans, Antarctic Ocean	8.002-8.050 (30 days)	-	-	-	-	-	-	-	Hofmann et al., 2011	
Upper Reach of the Pearl River Estuary (South China Sea)	6.98-7.75	-	≤7460	1900-2740	-	0.25-3.61	-	~14.5-16.8	Dai et al., 2006	
Outer estuary of Yangtze River (East China Sea)	7.76-8.03 (spring)	-	560-800	1575-1840	-	-	-	14.6-15.7	Zhai et al., 2007	
Outer estuary of Yangtze River (East China Sea)	7.92-8.04 (winter)	-	316-477	1935-1995	-	-	-	7.1-15.5	Zhai et al., 2007	
Bohai and Liaodong Bay (0 to ~15-20 m)	7.90-8.12 (June)	~7.82-8.04	-	-	-	~7.05-10.16	~7.21-9.51	~13.0-18.5*	Zhai et al., 2012	
Bohai and Liaodong Bay (0 to ~15-20 m)	7.86-8.17 (August)	~7.64-7.96	-	-	-	~7.21-9.18	~3.28-7.38	~19.0-27.0*	Zhai et al., 2012	
Yellow Sea: Northern side	7.67-7.92 (winter)	~7.67-7.75	-	-	-	~9.90- (6 m)	-	(-1.45)-6.74	Zhai et al., 2014	
Yellow Sea: Northern side	7.85-8.11 (spring)	~7.69-7.78	-	-	-	-	-	~4.50-13.22	Zhai et al., 2014	
Yellow Sea: Northern side	7.86-8.14 (summer)	~7.62-7.92	-	-	-	7.44-9.54 (13-19 m)	-	~5.00-24.83	Zhai et al., 2014	
Yellow Sea: Northern side	7.76-8.11 (autumn)	7.53-7.60	-	-	-	6.82 (19 m)	-	~7.50-18.37	Zhai et al., 2014	
East China Sea	8.04-8.22	-	-	1980-2220	-	-	-	~12-25	Wang et al., 2000	

Table S1 (continued)

Sampling	pH		<i>P</i> CO ₂ (μ atm)	DIC (μ mol kg ⁻¹)	Chl <i>a</i> (μ g L ⁻¹)	Dissolved O ₂		H ₂ O ₂ (nM)	WT	References
	Surface water (0-20 m)	Anoxic zone				Surface water (mg L ⁻¹)	Anoxic zone			
Coastal seawater, Northern South China Sea	< 8.2-8.6	~8.18 (65 m)	-	~1500-1900	-	-	-	-	-	Dai et al., 2008
Seto Inland Sea	8.15-8.26 (0-10 m)	> 7.96 (40 m)	-	~1750-1950	-	~ 6.0-9.5 (0-10 m)	> 3.0 (40 m)	-	~9-26	Taguchi and Fujiwara, 2010 [*]
Coastal seawater (Bay of Bengal)	7.96-8.38	-	416-1062	-	-	-	-	-	26.0-28.0	Biswas et al., 2004
Puget Sound Estuary, main basin (Summer)	8.05	7.83 (> 100 m)	-	1884	-	~ 5.74-8.20	-	-	-	Feeley et al., 2010
Puget Sound Estuary, main basin (Winter)	7.71	7.73 (> 100 m)	-	1985	-	8.2	-	-	-	Feeley et al., 2010
North Sea (English Coastal waters)	~8.10-8.16	-	-	-	-	~4.07-4.66	-	-	-	Frankignoulle et al., 1996 [†]
North Sea (French coastal waters)	~8.10-8.50; ~7.95-8.28	-	-	-	-	~3.69-6.39; 3.77-4.18	-	-	-	Frankignoulle et al., 1996 [†]
Northern Gulf of Mexico (summer)	> 8.2	< 7.6-8.0	-	-	-	-	-	-	-	Cai et al. 2011 [†]
Northern Gulf of Mexico (spring)	> 8.2	7.8-8.2	-	-	-	-	-	-	-	Cai et al. 2011 [†]
Northern Gulf of Mexico	8.10-8.18 (1-2 m)	7.71-7.93 (10-37 m)	-	-	-	5.48-6.10	0.10-2.79	-	23.0-31.2	Sunda and Cai, 2012
Monterey Bay L20 (pelagic), near Shore	7.857-8.356 (30 days)	-	-	-	-	-	-	-	-	Hofmann et al., 2011
Upwelling seawater (pelagic), Pt. Concepcion, Monterey Bay	7.869-8.266 (30 days)	-	-	-	-	-	-	-	-	Hofmann et al., 2011
Upwelling seawater (pelagic), Pt. Ano Nuevo, Monterey Bay	7.685-8.152 (30 days)	-	-	-	-	-	-	-	-	Hofmann et al., 2011
North Pacific Ocean (25-55°N, 152°W)	8.2-7.6	~7.6-7.3 (> 500)	-	-	-	-	-	-	-	Byrne et al. 2010
Southern Ocean (Atlantic sector)	8.060-8.123	-	313-377	-	-	5.31-5.89	-	-	~(-2)—13.8	Chierici et al., 2004 [†]
Southern Ocean (Atlantic sector)	~8.06-8.12	-	~313-377	-	~0.2-1.0	-	-	-	-	Chierici et al., 2004 [†]
Cold-water, northeast Atlantic (5 sites; depths 0-1930 m)	< 8.10-8.19	> 7.92-8.06	-	-	0.02-1.2	-	-	-	-	Findlay et al., 2014
Submarine spring, Puerto Morelos, Mexico	7.143-8.048 (30 days)	-	-	-	-	-	-	-	-	Hofmann et al., 2011
CO ₂ vent, Ischia (South Zone), Italy	6.699-8.129 (30 days)	-	-	-	-	-	-	-	-	Hofmann et al., 2011
CO ₂ vent, off Ischia Island, Italy (surface water at site S1)	8.13-8.17 (21 days)	-	304-346	2120-2160	-	-	-	-	18.0-19.7	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (surface water at site S2)	7.35-8.16 (21 days)	-	314-2,626	2120-2510	-	-	-	-	17.8-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (surface water at site S3)	6.07-7.37 (21 days)	-	2,448-51,997	2510-4220	-	-	-	-	18.0-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (surface water at site N1)	7.96-8.19 (21 days)	-	286-552	2100-2240	-	-	-	-	18.0-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (surface water at site N2)	7.54-8.17 (21 days)	-	309-1,644	2130-2440	-	-	-	-	18.0-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (surface water at site N3)	6.31-7.90 (21 days)	-	654-29,885	2280-3500	-	-	-	-	18.0-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (at site P1 on 3 m depth)	8.15-8.18 (21 days)	-	296-322	2110-2130	-	-	-	-	18.1-20.0	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (at site P2 on 3 m depth)	8.13-8.20 (21 days)	-	278-347	2100-2160	-	-	-	-	18.5	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (at site P3 on 3 m depth)	7.67-8.16 (21 days)	-	315-1,173	2130-2390	-	-	-	-	19.0-19.5	Hall-Spencer et al., 2008
CO ₂ vent, off Ischia Island, Italy (at site P4 on 3 m depth)	6.98-8.14 (21 days)	-	334-6,253	2140-2670	-	-	-	-	19.0-19.5	Hall-Spencer et al., 2008
Hiroshima Bay, Japan	-	-	-	-	-	-	-	143-448	19.5-22.1	Akane et al. 2004
Hiroshima Bay, Japan	-	-	-	-	-	-	-	0-195	-	Sakugawa et al., 2000
Iyo-Nada Bay, Japan	-	-	-	-	-	-	-	7-146	-	Sakugawa et al., 2000
Seto Inland Sea, Japan	-	-	-	-	-	-	-	40-191	-	Sakugawa et al., 1995
Tokyo Bay	-	-	-	-	-	-	-	20-207	-	Sakugawa et al., 1995
Sagami Bay	-	-	-	-	-	-	-	40-80	-	Sakugawa et al., 1995
Mediterranean, Israeli coastal waters	-	-	-	-	-	-	-	10.0-80.0	-	Herut et al., 1998
Red Sea, Gulf of Aqaba	-	-	-	-	-	-	-	8-100	-	Herut et al., 1998
Lagrangian, Atlantic Ocean	-	-	-	-	-	-	-	23-55	-	Yuan and Shiller, 2001
Underway, Atlantic Ocean	-	-	-	-	-	-	-	27-47	-	Yuan and Shiller, 2001
Bermuda Atlantic Time Series Station	-	-	-	-	-	-	-	25-84	-	Avery Jr. et al., 2005

The dissolved O₂ values measured at different units adjusted using 1 ml L⁻¹ ≈ 1.4 mg L⁻¹ at 20 psu and 2 mgO₂ L⁻¹ ≈ 61 μmolO₂ kg⁻¹ (Gray et al., 2002).

^{*}Data extracted from the graph.

