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Augmenting the Biological Pump

THE SHORTCOMINGS OF GEOENGINEERED UPWELLING

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The ocean is the largest reservoir of mobile carbon over decadal to centennial time scales, absorbing approximately 41% of cumulative anthropogenic CO_2 emissions (Sabine and Tanhua, 2010). Various geoengineering solutions seek to exploit this uptake capacity (see Vaughan and Lenton, 2011, for a review), including CO_2 injection (Marchetti, 1977), iron fertilization (Martin et al., 1994), and artificial upwelling (Lovelock and Rapley, 2007). The ubiquity of social media—allowing anyone to "self-publish"—and funding from crowd-sources and private foundations have allowed some proposals to gain traction outside of the peer-reviewed scientific literature. A recent example is the proposal by theoretical neurobiologist W.H. Calvin (2013) to construct a massive array of push-pull pump systems to enhance the ocean's natural biological pump to sequester atmospheric CO_2.

Here, we evaluate Calvin's proposal in the context of other artificial upwelling proposals and studies, considering its feasibility and efficacy in terms of site selection, physical constraints and energy requirements, carbon cycle and nutrient dynamics, and potential ecosystem impacts. Despite novel aspects, we show that Calvin's proposal would be unlikely to sequester large amounts of carbon. Instead, its implementation could lead to an increase in CO_2 outgassing while causing significant biogeochemical and ecological changes. It is important that ocean experts publicly critique proposals like Calvin's, both to counter potentially harmful interventions and to encourage effective ones.

ARTIFICIAL UPWELLING PROPOSALS

Artificial upwelling aims to stimulate primary production by bringing nutrient-rich, sub-euphotic water up to the surface. The upwelled nutrients would fuel enhanced fixation of inorganic carbon into organic carbon, thus removing dissolved CO_2 from surface waters and potentially increasing the flux of atmospheric CO_2 into the ocean. Sinking of particulate organic carbon (POC) could then sequester that carbon in deep ocean waters for decades or centuries.

Proposed upwelling mechanisms include salt fountains (e.g., Stommel et al., 1956; Tsubaki et al., 2007), airlift pumps (e.g., Fan et al., 2013), and wind- or wave-powered systems (Kenyon, 2007). Models have also been used to explore the utility of artificial upwelling in sequestering carbon (Dutreuil et al., 2009; Yool et al., 2009; Oschlies et al., 2010) and its potentially undesirable global effects. As we describe below, these studies have clarified the constraints, limitations, and consequences of geoengineered global carbon sequestration.

Calvin's (2013) proposal includes two components: (1) creation of artificial upwelling by pumping deep, nutrient-rich waters to the euphotic zone to stimulate phytoplankton blooms, and (2) the novel component of pumping the resultant particulate and dissolved organic carbon (DOC) down into the deep ocean for long-term sequestration (Figure 1). The goal is a "big, quick, and secure" sequestration of 30 GtC yr^{-1} over 20 years, that is, the removal of all anthropogenic CO_2 emissions since 1750 (approximately 600 GtC). Calvin estimates that the remineralization and gradual return of these 600 GtC to the atmosphere would be spread over millennia and states that pumps using existing technology covering 1% of the ocean surface would be sufficient to implement his plan.

SITE SELECTION

Reaching the 1,000-Year Horizon

The efficacy of carbon sequestration increases with injection depth and varies geospatially (Stegen et al., 1993). Millennial-scale sequestration requires pumping carbon to an isopycnal that will remain out of contact with the atmosphere for at least 1,000 years. De Vries and Primeau (2011) showed that the shallowest 1,000-year depth horizon is in the North Pacific at ~ 2,000 m (Figure 2). Thus, 1,000-year sequestration would require pumping water to at least 2 km depth—more than 10 times deeper than Calvin proposed. Furthermore, in a Southern Ocean model, only 71% of particles injected to 2,000 m remain sequestered for more than 100 years, one-tenth of the time scale Calvin proposed (Robinson et al., 2014).

Mixed-Layer Depth

Water must be pumped up from below the deepest local mixed-layer depth (MLD) to acquire nutrient concentrations greater than those within the mixed layer. Because MLD varies considerably by geographic location and time of year, the proposed pipe length of 150 m might be sufficient for upward pumping in the subtropical North Pacific, which has typical MLDs of ~ 100 m. However, MLDs are greater in other parts of the ocean; 300 m is typical in the Southern Ocean (Talley et al., 2011), potentially requiring considerably longer pipes than Calvin suggested.

Nutrient Supply and Horizontal Advection

Maintaining upwelling nutrient plumes requires a deep nutrient resupply, which depends on regionally heterogeneous advection and remineralization. Simulations show that surface nutrient plumes disperse quickly; in one artificial upwelling model, nutrients were diluted to less than 2% of initial plume concentrations 10 m downstream from the injection site (Williamson et al., 2009).

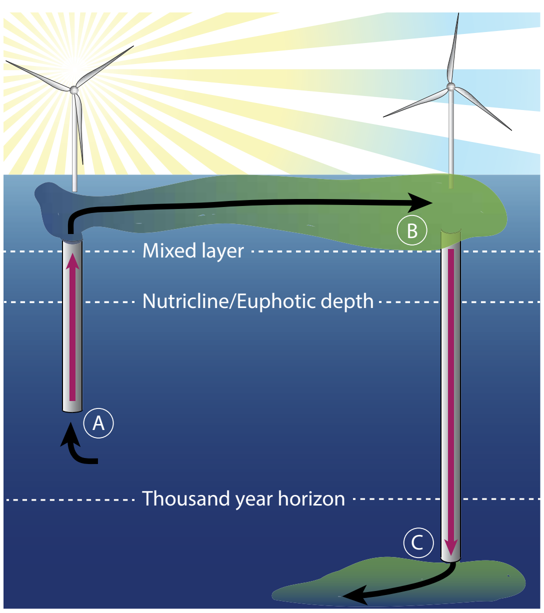

Figure 1. Schematic of the Calvin (2013) pumping plan. Nutrient-rich water from below the nutricline (A) is pumped up to the surface where it stimulates a phytoplankton bloom (B), which causes a drawdown in atmospheric CO_2. Particulate and dissolved organic carbon produced during the bloom are pumped back down below the 1,000-year horizon (C), transferring carbon from the atmosphere to the deep ocean.

Blooms are typically apparent five to seven days after nutrient injection to the euphotic zone, after which phytoplankton begin to sink or are grazed (Boyd et al., 2007). At a canonical horizontal speed of 10 cm s^{-1}, a bloom travels 50 km in five days. To capture the bloom, Calvin's down-pumps would need to be ~ 50 km to 100 km downstream of the up-pumps. However, defining "downstream" would be challenging in regions of even moderate mesoscale current variability.

PHYSICAL CONSIDERATIONS

Moving water against a density gradient requires an input of kinetic energy. We estimated lower-bound energy requirements using temperature and salinity data from a typical Southern Ocean Argo float profile; neglecting mixing, approximately 50 J m^{-3} would be required to raise water from 200 m (sub-nutricline) to 50 m depth (in the mixed layer). An additional 8,000 J m^{-3} would be required to return this water from the mixed layer to the shallowest 1,000-year horizon depth of 2,000 m. Given a very low pumping rate estimate of 202 Sv (see Carbon Cycle Dynamics) necessary to sustain a bloom, approximately 1.7 TW of energy would be needed solely for moving water (Figure 3). Estimates of the wind power transmitted to the ocean globally range from 0.6 to 1.2 TW (Munk and Wunsch, 1998; Watanabe and Hibiya, 2002) and

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those of tidal energy are ~ 3.5 TW. Thus, our estimate of energy required represents more than all global wind energy and over one-third of the entire internal energy budget of the world ocean.

Although upwelling rates of ~ 45 $m^3 h^{-1}$ have been achieved with conventional pumping methods, they are unsustainable for longer periods of time (White et al., 2010). Even novel techniques like airlifting operate with low (~ 15%) energy efficiency at high flow rates (White et al., 2010; Fan et al., 2013). Pumps driven by ocean salinity gradients require no permanent energy input but only produce flow rates of around 46 $m^3 day^{-1}$ (Tsubaki et al., 2007), substantially increasing the required 0.8 billion 1 m diameter pipes estimated by Yool et al. (2009) to enhance ocean CO_2 uptake by 1 GtC yr^{-1}. Additionally, surface entrainment of upwelled nutrients has not been conclusively demonstrated either in model or theory (Williamson et al., 2009; Fan et al. 2013). Finally, failed field tests using 0.4 m diameter pipes < 30 m long led Fan et al. (2013) to question the feasibility of deploying long, large-diameter pipes (see also Lovelock and Rapley, 2007) in a dynamic ocean.

CARBON CYCLE DYNAMICS

The biological and solubility pumps transport carbon into the ocean interior, creating a surface-to-deep dissolved inorganic carbon (DIC) gradient. The biological pump, driven by particulate carbon export, generates 90% of this gradient, and 10% is generated by the solubility pump, driven by temperature-dependent CO_2 solubility, water mass formation, and biologically generated air-sea pCO_2 gradients (Sarmiento and Gruber, 2006). Artificial upwelling might increase oceanic CO_2 uptake if phytoplankton blooms were to enhance export production. However, increased primary production does not always result in increased export production, and export production is not a straightforward predictor for air-sea gas exchange (Oschlies and Kahler, 2004). Because > 95% of organic carbon is remineralized within the upper 1,000 m (Yool et al., 2009), the increase in oceanic CO_2 uptake relative to primary production (i.e., efficiency) would be low.

The novelty of Calvin's proposal lies in down-pumping organic carbon into the deep ocean after artificially stimulating blooms. Calvin estimates the net DIC flux to the surface at 0.48 g m^{-3} of water pumped (given a surface-to-deep DIC difference of 40 mmol kg^{-1}) would be eclipsed by resulting primary production (at least 30 GtC yr^{-1}). Performing similar calculations, we arrived at a very different answer. Taking (conservatively) the largest surface-to-deep phosphate difference of 3 µM (such as in the North Pacific; Paytan and McLaughlin, 2007) and a Redfield C:P ratio of 131:1 (including carbonates; Sarmiento and Toggweiler, 1984), a minimum pumping rate of 202 Sv—more than an order of magnitude greater than the Atlantic Meridional Overturning Circulation (18 Sv; Liu and Liu, in press)—is required to supply enough phosphate to fuel a steady-state production of 30 GtC yr^{-1}. Given the corresponding

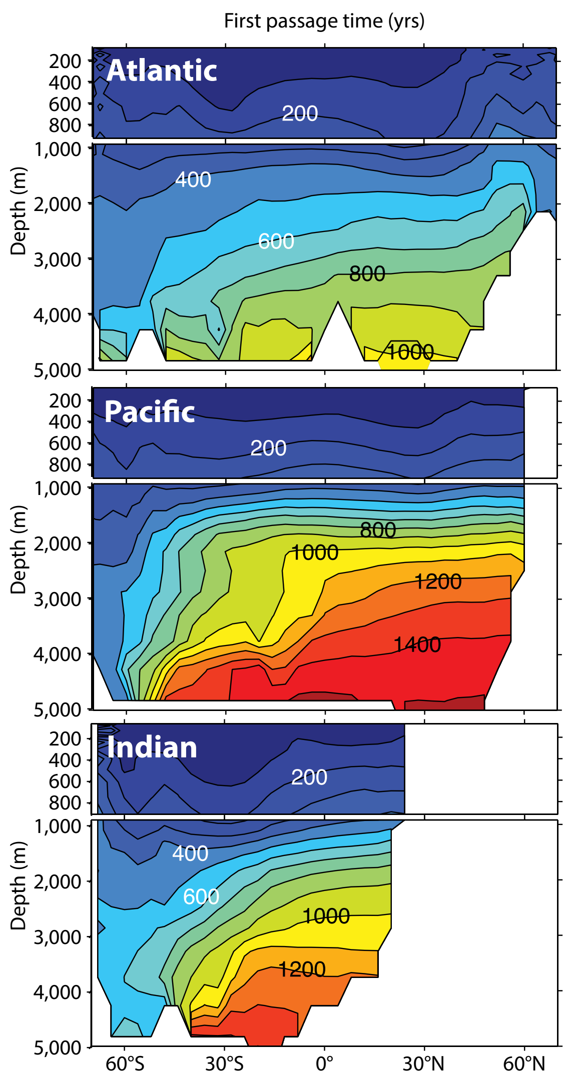

Figure 2. Mean first-passage times (years) calculated using a data-constrained ocean circulation model (modified from De Vries and Primeau, 2011). Mean first-passage time is the average time it would take a parcel of water to reach the surface for each ocean basin. The global minimum depth of the 1,000-year horizon is 2,000 m, located in the North Pacific.

surface-to-deep DIC difference (Millero, 2007), this pumping would bring 29 GtC yr^{-1} to the surface as DIC. Similar calculations with North Atlantic DIC and phosphate gradients yield pumping rates of ~ 1,000 Sv and surface DIC flux of ~ 50 GtC yr^{-1}.

Even if this pumping could be achieved, it would likely result in little to no net drawdown of atmospheric CO_2. Natural wind-driven coastal upwelling typically results in net outgassing of CO_2 at low latitudes (< 30°) at rates on the order of 12 gC m^{-2} yr^{-1} (Cai et al., 2006). Yool et al.'s (2009) model predicted that pipes placed in temperate and polar regions were likely to turn those areas into additional sources of CO_2. Oschlies et al. (2010) modeled the effects of artificial upwelling and found only an additional 0.9 GtC yr^{-1} to be sequestered, 90% of which would occur terrestrially due to lower global temperatures and decreased respiration.

Calvin argues that his plan would sequester carbon in excess of POC export because surface DOC pumped down represents a greater pool of carbon and has a refractory fraction that persists for millennia. Although refractory DOC has an average age of 4,000 to 6,000 years, photochemically reactive refractory DOC has a residence time of only 500 to 2,100 years (Mopper et al., 1991), releasing CO_2 upon degradation. Artificial upwelling may increase the photodegradation of refractory DOC brought to the surface and thus decrease the net export of refractory DOC.

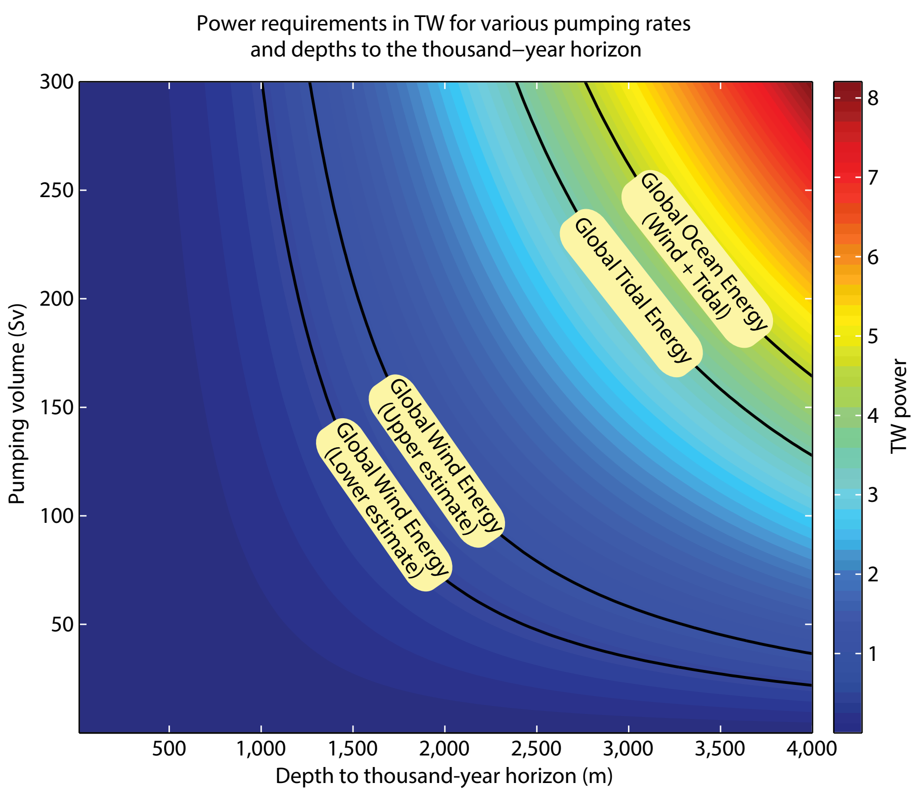

Figure 3. Contours of power required for up- and down-pumping of water as a function of volumetric pumping rate and depth of the 1,000-year horizon. Water is assumed to be pumped up from 200 m (typically below the mixed layer depth and nutricline) to a depth of 50 m within the euphotic zone. Temperature, salinity, and pressure data are from a high-resolution conductivity-temperature-depth (CTD) cast (WOD Unique Cast Number 15561184) during the 2011 Southern Ocean cruise operated by Scripps Institution of Oceanography. The lack of a significant pycnocline at this location ensures these energy estimates are conservative.

BIOGEOCHEMICAL CYCLING

Biogeochemical side effects induced by artificial upwelling may undermine the goal of carbon sequestration. Subsurface increases in respiration due to increased organic carbon flux would consume oxygen and expand pre-existing oxygen minimum zones (OMZs). Nitrous oxide (N_2O), CO_2, and methane (CH_4) are produced at the upper boundary of OMZs, as respiration demands alternative electron acceptors. N_2O and CH_4, which respectively possess 320 and 20 times the greenhouse potency of CO_2, could be released into the atmosphere, potentially either offsetting the benefits of atmospheric CO_2 reduction or causing a net increase in greenhouse equivalents of CO_2 (Fuhrman and Capone, 1991; Jin and Gruber, 2003). In addition to generating potent greenhouse gases, expansion of denitrification zones would increase the loss of nitrate, which fuels new production and is a limiting nutrient for global ocean primary productivity (Codispoti et al., 2001; Gruber, 2004). The remineralization of 600 GtC in the deep ocean over 20 years, as Calvin proposes, would consume 6.51 x 10^{16} mol O_2 (assuming a Redfield O_2:C ratio of 138:106) at an average rate of 3.25 x 10^{15} mol O_2 yr^{-1}—61 to 205 times the estimated average rates of global ocean O_2 inventory decrease from pre-industrial times to 2100 (Keeling et al., 2010).

ECOLOGICAL IMPACTS

Eutrophication

Planktonic biomass and community structure depend on nutrient fluxes; where surface nutrient fluxes are low, small phytoplankton and protistan grazers dominate. Upwelling regions with higher nutrient fluxes favor larger phytoplankton such as diatoms (Zarauz et al., 2009). Shifts from oligotrophic (nutrient-poor) to eutrophic (nutrient-rich) communities may be observed in as little as one week, the approximate time it takes for a phytoplankton bloom to develop (Aure et al., 2007).

Artificial upwelling would inevitably shift the phytoplankton community toward larger cells, which contribute disproportionately to exported POC (Boyd and Newton, 1999; Brzezinski et al., 2011). Though Calvin suggests an additional set of tubes to pump diatoms to the deep sea, capturing a bloom is problematic (see Nutrient Supply and Horizontal Advection). Thus, significant biological carbon sequestration would be dependent on diatoms sinking rather than on downward pumping. However, when diatoms have access to sufficient nutrients, they often make themselves positively buoyant (Waite et al., 1997; Acuña et al., 2010).

Larger phytoplankton at the base of a marine food web result in greater biomass at all trophic levels, including commercially valuable species. Unfortunately, upwelling enhancement of fisheries will reduce the efficacy of carbon sequestration because of the respiration of the fixed organic carbon. In some systems, upwelling leads to harmful algal blooms (HABs) (Ryan et al., 2009; though see McClimans et al., 2010), causing fish and marine mammal mortality (Flewelling et al., 2005) as well as economic and health concerns (Jin et al., 2008). Additional ad hoc solutions to address HABs only complicate Calvin's scheme; such risks must be thoroughly assessed prior to engaging in geoengineering because the complexity of ocean dynamics precludes deterministic predictions (Cullen and Boyd, 2008).

Deoxygenation

Keller et al. (2014) modeled global-scale effects of artificial upwelling and predicted a 265% increase in the size of suboxic zones. Mobile, predatory pelagic fish with high oxygen demands like tuna avoid these regions (Prince and Goodyear, 2006), but sessile organisms can be decimated (Wu, 2002; Gooday et al., 2009; Levin et al., 2009). Hypoxia-tolerant species, like gelatinous organisms, benefit from OMZs through, for example, reduced benthic predation on their larvae (Purcell et al., 2007); these species dominate previously stressed communities, further decreasing biodiversity. Oxygen depletion can also significantly alter the ecological activity of microbes, impacting nutrient cycling and increasing the production of toxic H_2S (Bartoli et al., 2009).

Acidification

Models project that artificial upwelling would decrease pH up to 0.15 units beyond the present trajectory (e.g., Keller et al., 2014) through respiration of POC in deep waters. Acidification of oceanic environments is already an ongoing, serious threat to ocean life (see Doney, 2006, and references therein), and this would be exacerbated by artificial upwelling and downwelling.

The benefits of any geoengineering plan that would change the trophic state, oxygen content, or pH of the ocean must be weighed against potential negative impacts on marine organisms and ecosystems. These consequences will be serious and long-lasting; while artificial upwelling can be shut off within a year, ecosystem and biogeochemical effects would persist much longer (Cullen and Boyd, 2008; Lampitt et al., 2008; Law, 2008).

CONCLUSIONS

Artificially enhanced upwelling of deep, nutrient-rich waters and the subsequent downward pumping of fixed organic carbon is an appealing idea for the sequestration of anthropogenic atmospheric CO_2. However, we show that studies of artificial upwelling-enhanced carbon sequestration require unworkable magnitudes of pumping and energy input, raise practical issues associated with system implementation, and bring to light significant negative impacts on marine and atmospheric carbon dynamics and marine ecosystems. Given the absence of positive supporting scientific evidence, we do not recommend pursuing geoengineering through artificial upwelling; our calculations indicate it is unfeasible and may amplify the warming trend it seeks to reduce.

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REFERENCES

Acuña, J.L., M. López-Alvarez, E. Nogueira, and F. González-Taboada. 2010. Diatom flotation at the onset of the spring phytoplankton bloom: An in situ experiment. *Marine Ecology Progress Series* 400:115–125, http://dx.doi.org/10.3354/meps08405.

Aure, J., S. Oivind, S.R. Erga, and T. Strohmeier. 2007. Primary production enhancement by artificial upwelling in a western Norwegian fjord. *Marine Ecology Progress Series* 352:39–52, http://dx.doi.org/10.3354/meps07139.

Bartoli, M., L. Vezzulli, D. Nizzoli, R. Azzoni, S. Porrello, M. Moreno, M. Fabiano, and P. Viaroli. 2009. Short-term effect of oxic to anoxic transition on benthic microbial activity and solute fluxes in organic-rich phytotreatment ponds. *Hydrobiologia* 629:123–136, http://dx.doi.org/10.1007/978-90-481-3385-7_11.

Boyd, P.W., and P.P. Newton. 1999. Does planktonic community structure determine downward particulate organic carbon flux in different oceanic provinces? *Deep Sea Research Part I* 46(1):63–91, http://dx.doi.org/10.1016/S0967-0637(98)00066-1.

Boyd, P.W., T. Jickells, C.S. Law, S. Blain, E.A. Boyle, K.O. Buesseler, K.H. Coale, J.J. Cullen, H.J.W. de Baar, M. Follows, and others. 2007. Mesoscale iron enrichment experiments 1993–2005: Synthesis and future directions. *Science* 315:612–617, http://dx.doi.org/10.1126/science.1131669.

Brzezinski, M.A., J.W. Krause, M.J. Church, D.M. Karl, B. Li, J.L. Jones, and B. Updyke. 2011. The annual silica cycle of the North Pacific subtropical gyre. *Deep Sea Research Part I* 158:988–1,001, http://dx.doi.org/10.1016/j.dsr.2011.08.001.

Cai, W.J., M. Dai, Y. Wang. 2006. Air-sea exchange of carbon dioxide in ocean margins: A province-based synthesis. *Geophysical Research Letters* 33, L12603, http://dx.doi.org/10.1029/2006GL026219.

Calvin, W.H. 2013. Using the oceans to remove CO_2 from the atmosphere. http://geo-engineering.blogspot.com/2013/03/using-the-oceans-to-remove-co2-from-the-atmosphere.html (accessed March 9, 2014).

Codispoti, L.A., J.A. Brandes, J.P. Christensen, A.H. Devol, S.W.A. Naqvi, H.W. Paerl, and T. Yoshinari. 2001. The oceanic fixed nitrogen and nitrous oxide budgets: Moving targets as we enter the Anthropocene? *Scientia Marina* 65:85–105, http://dx.doi.org/10.3989/scimar.2001.65s285.

Cullen, J.J., and P.W. Boyd. 2008. Predicting and verifying the intended and unintended consequences of large-scale ocean iron fertilization. *Marine Ecology Progress Series* 364:295–301, http://dx.doi.org/10.3354/meps07551.

De Vries, T., and F. Primeau. 2011. Dynamically and observationally constrained estimates of water-mass distributions and ages in the global ocean. *Journal of Physical Oceanography* 41:2,381–2,401, http://dx.doi.org/10.1175/JPO-D-10-05011.1.

Doney, S.C. 2006. The dangers of ocean acidification. *Scientific American Series* 294:58–65, http://dx.doi.org/10.1038/scientificamerican0306-58.

Dutreuil, S., L. Bopp, and A. Tagliabue. 2009. Impact of enhanced vertical mixing on marine biochemistry: Lessons for geo-engineering and natural variability. *Biogeosciences* 6:901–912, http://dx.doi.org/10.5194/bg-6-901-2009.

Fan, W., J. Chen, Y. Pan, H. Huang, C.-T.A. Chen, and Y. Chen. 2013. Experimental study on the performance of an air-lift pump for artificial upwelling. *Ocean Engineering* 59:47–57, http://dx.doi.org/10.1016/j.oceaneng.2012.11.014.

Flewelling, L.J., J.P. Naar, J.P. Abbott, D.G. Baden, N.B. Barros, G.D. Bossart, M.-Y.D. Bottein, D.G. Hammond, E.M. Haubold, C.A. Heil, and others. 2005. Brevetoxicosis: Red tides and marine mammal mortalities. *Nature* 435:755–756, http://dx.doi.org/10.1038/nature435755a.

Fuhrman, J.A., and D.G. Capone. 1991. Possible biogeochemical consequences of ocean fertilization. *Limnology and Oceanography* 36:1,951–1,959.

Gooday, A.J., L.A. Levin, A. Aranda da Silva, B.J. Bett, G.L. Cowie, D. Dissard, J.D. Gage, D.J. Hughes, R. Jeffeys, P.A. Lamont, and others. 2009. Faunal responses to oxygen gradients on the Pakistan margin: A comparison of foraminiferans, macrofauna and megafauna. *Deep Sea Research Part II* 56:488–502, http://dx.doi.org/10.1016/j.dsr2.2008.10.003.

Gruber, N. 2004. The dynamics of the marine nitrogen cycle and its influence on atmospheric CO_2. Pp. 97–148 in *The Ocean Carbon Cycle and Climate*. M. Follows and T. Oguz, eds, Kluwer Academic, Dordrecht.

Jin, D., E. Thunberg, and P. Hoagland. 2008. Economic impact of the 2005 red tide event on commercial shellfish fisheries in New England. *Ocean & Coastal Management* 51:420–429, http://dx.doi.org/10.1016/j.ocecoaman.2008.01.004.

Jin, X., and N. Gruber. 2003. Offsetting the radiative benefit of ocean iron fertilization by N_2O emissions. *Geophysical Research Letters* 30, 2249, http://dx.doi.org/10.1029/2003GL018458.

Keeling, R.F., A. Kortzinger, and N. Gruber. 2010. Ocean deoxygenation in a warming world. *Annual Review of Marine Science* 2:199–229, http://dx.doi.org/10.1146/annurev.marine.010908.163855.

Keller, D.P., E.Y. Feng, and A. Oschlies. 2014. Potential climate engineering effectiveness and side effects during a high carbon dioxide-emission scenario. *Nature Communications* 5, 3304, http://dx.doi.org/10.1038/ncomms4304.

Kenyon, K.E. 2007. Upwelling by a wave pump. *Journal of Oceanography* 63:327–331, http://www.terrapub.co.jp/journals/JO/pdf/6302/63020327.pdf.

Lampitt, R.S., E.P. Achterberg, T.R. Anderson, J.A. Hughes, M.D. Iglesias-Rodriguez, B.A. Kelly-Gerreyn, M. Lucas, E.E. Popova, R. Sanders, J.G. Shepherd, and others. 2008. Ocean fertilization: A potential means of geoengineering? *Philosophical Transactions of the Royal Society A* 366:3,919–3,945, http://dx.doi.org/10.1098/rsta.2008.0139.

Law, C.S. 2008. Predicting and monitoring the effects of large-scale ocean iron fertilization on marine trace gas emissions. *Marine Ecology Progress Series* 364:283–288, http://dx.doi.org/10.3354/meps07549.

Levin, L.A., W. Ekau, A.J. Gooday, F. Jorissen, J.J. Middelburg, W. Naqvi, C. Neira, N.N. Rabalais, and J. Zhang. 2009. Effects of natural and human-induced hypoxia on coastal benthos. *Biogeosciences Discussions* 6:2,063–2,098, http://dx.doi.org/10.5194/bg-6-2063-2009.

Liu, W., and Z. Liu. In press. Assessing the stability of the Atlantic meridional overturning circulation of the past, present, and future. *Journal of Meteorological Research* 28, http://dx.doi.org/10.1007/s13351-014-4006-6.

Lovelock, J.E., and C.G. Rapley. 2007. Ocean pipes could help the Earth to cure itself. *Nature* 449:403, http://dx.doi.org/10.1038/449403a.

Marchetti, C. 1977. On geoengineering and the CO_2 problem. *Climatic Change* 1:59–68, http://dx.doi.org/10.1007/BF00162777.

Martin, J.H., K.H. Coale, K.S. Johnson, S.E. Fitzwater, R.M. Gordon, S.J. Tanner, C.N. Hunter, V.A. Elrod, J.L. Nowicki, T.L. Coley, and others. 1994. Testing the iron hypothesis in ecosystems of the equatorial Pacific Ocean. *Nature* 371:123–129, http://dx.doi.org/10.1038/371123a0.

McClimans, T.A., A. Handå, A. Fredheim, E. Lien, and K.I. Reitan. 2010. Controlled artificial upwelling in a fjord to stimulate non-toxic algae. *Aquacultural Engineering* 42:140–147, http://dx.doi.org/10.1016/j.aquaeng.2010.02.002.

Millero, F.J., 2007. The marine inorganic carbon cycle. *Chemical Reviews* 107:308–341, http://dx.doi.org/10.1021/cr0503557.

Mopper, K., X. Zhou, R.J. Kieber, D.J. Kieber, R.J. Sikorski, and R.D. Jones. 1991. Photochemical degradation of dissolved organic carbon and its impacts on the oceanic carbon cycle. *Nature* 353:60–62, http://dx.doi.org/10.1038/353060a0.

Munk, W., and C. Wunsch. 1998. Abyssal recipes II: Energetics of tidal and wind mixing. *Deep Sea Research Part I* 45:1,977–2,010, http://dx.doi.org/10.1016/S0967-0637(98)00070-3.

Oschlies, A., and P. Kahler. 2004. Biotic contribution to air-sea fluxes of CO_2 and O_2 and its relation to new production, export production, and net community production. *Global Biogeochemical Cycles* 18, GB1015, http://dx.doi.org/10.1029/2003GB002094.

Oschlies, A., M. Pahlow, A. Yool, and R.J. Matear. 2010. Climate engineering by artificial ocean upwelling: Channeling the sorcerer's apprentice. *Geophysical Research Letters* 37, L04701, http://dx.doi.org/10.1029/2009GL041961.

Paytan, A., and K. McLaughlin. 2007. The oceanic phosphorus cycle. *Chemical Review* 107:563–576, http://dx.doi.org/10.1021/cr0503613.

Prince, E.D., and C.P. Goodyear. 2006. Hypoxia-based habitat compression of tropical pelagic fishes. *Fisheries Oceanography* 15:451–464, http://dx.doi.org/10.1111/j.1365-2419.2005.00393.x.

Purcell, J.E., S.I. Uye, and W.T. Lo. 2007. Anthropogenic causes of jellyfish blooms and their direct consequences for humans: A review. *Marine Ecology Progress Series* 350:153–174, http://dx.doi.org/10.3354/meps07093.

Robinson, J., E.E. Popova, A. Yool, M. Srokosz, R.S. Lampitt, and J.R. Blundell. 2014. How deep is deep enough? Ocean iron fertilization and carbon sequestration in the Southern Ocean. *Geophysical Research Letters* 41:2,489–2,495, http://dx.doi.org/10.1002/2013GL058799.

Ryan, J.P., A.M. Fischer, R.M. Kudela, J.F.R. Gower, S.A. King, R. Marin III, and F.P. Chavez. 2009. Influences of upwelling and downwelling winds on red tide bloom dynamics in Monterey Bay, California. *Continental Shelf Research* 29:785–795, http://dx.doi.org/10.1016/j.csr.2008.11.006.

Sabine, C.L., and T. Tanhua. 2010. Estimation of anthropogenic CO_2 inventories in the ocean. *Annual Reviews of Marine Science* 2:175–198, http://dx.doi.org/10.1146/annurev-marine-120308-080947.

Sarmiento, J.L., and J.R. Toggweiler. 1984. A new model for the role of the oceans in determining atmospheric $p$$CO_2$. *Nature* 308(12):621–624, http://dx.doi.org/10.1038/308621a0.

Sarmiento, J.L., and N. Gruber. 2006. *Ocean Biogeochemical Dynamics*. Princeton University Press 528 pp.

Stegen, G.R., K.H. Cole, and R. Bacastow. 1993. The influence of discharge depth and location on the sequestration of carbon dioxide. *Energy Conversion and Management* 34:857–864, http://dx.doi.org/10.1016/0196-8904(93)90029-A.

Stommel, H., A.B. Arons, and D. Blanchard. 1956. An oceanographical curiosity: The perpetual salt fountain. *Deep Sea Research* 3:152–153, http://dx.doi.org/10.1016/0146-6313(56)90095-8.

Talley, L.D., G.L. Pickard, W.J. Emery, and J.H. Swift. 2011. *Descriptive Physical Oceanography: An Introduction*, 6th ed. Elsevier, Boston, 560 pp.

Tsubaki, K., S. Maruyama, A. Komiya, and H. Mitsugashira. 2007. Continuous measurement of an artificial upwelling of deep sea water induced by the perpetual salt fountain. *Deep Sea Research Part I* 54:75–84, http://dx.doi.org/10.1016/j.dsr.2006.10.002.

Vaughan, N.E., and T.M. Lenton. 2011. A review of climate geoengineering proposals. *Climatic Change* 109:745–790, http://dx.doi.org/10.1007/s10584-011-0027-7.

Waite, A., A. Fisher, P.A. Thompson, and P.J. Harrison. 1997. Sinking rate versus cell volume relationships illuminate sinking rate control mechanisms in marine diatoms. *Marine Ecology Progress Series* 157:97–108, http://dx.doi.org/10.3354/meps157097.

Watanabe, M., and T. Hibiya. 2002. Global estimates of wind-induced energy flux to inertial motion in the surface mixed layer. *Geophysical Research Letters* 29(8), http://dx.doi.org/10.1029/2001GL014422.

White, A., K. Björkman, E. Grabowski, R. Letelier, S. Poulos, B. Watkins, and D. Karl. 2010. An open ocean trial of controlled upwelling using wave pump technology. *Journal of Atmospheric and Oceanic Technology* 27:385–396, http://dx.doi.org/10.1175/2009JTECHO679.1.

Williamson, N., A. Komiya, S. Maruyama, M. Behnia, and S.W. Armfield. 2009. Nutrient transport from an artificial upwelling of deep sea water. *Journal of Oceanography* 65:349–359, http://dx.doi.org/10.1007/s10872-009-0032-x.

Wu, R.S. 2002. Hypoxia: From molecular responses to ecosystem responses. *Marine Pollution Bulletin* 45:35–45, http://dx.doi.org/10.1016/S0025-326X(02)00061-9.

Yool, A., J.G. Shepherd, H.L. Bryden, and A. Oschlies. 2009. Low efficiency of nutrient translocation for enhancing oceanic uptake of carbon dioxide. *Journal of Geophysical Research* 114, C08009, http://dx.doi.org/10.1029/2008JC004792.

Zarauz, L., X. Irigoien, and J.A. Fernandes. 2009. Changes in plankton size structure and composition, during the generation of a phytoplankton bloom, in the central Cantabrian Sea. *Journal of Plankton Research* 31:193–207, http://dx.doi.org/10.1093/plankt/fbn107.