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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).

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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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.
those of tidal energy are \( \sim 3.5 \) 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 \( \sim 45 \text{ m}^3 \text{ h}^{-1} \) 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 \( 46 \text{ m}^3 \text{ day}^{-1} \) (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 \( p\text{CO}_2 \) 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 \( \mu \)M (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.](./image.png)
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 CO₂. Natural wind-driven coastal upwelling typically results in net outgassing of CO₂ 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 CO₂. 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 CO₂ upon degradation. Artificial upwelling may increase the photodegradation of refractory DOC brought to the surface and thus decrease the net export of refractory DOC.

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 (N₂O), CO₂, and methane (CH₄) are produced at the upper boundary of OMZs, as respiration demands alternative electron acceptors. N₂O and CH₄, which respectively possess 320 and 20 times the greenhouse potency of CO₂, could be released into the atmosphere, potentially either offsetting the benefits of atmospheric CO₂ reduction or causing a net increase in greenhouse equivalents of CO₂ (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 10¹⁶ mol O₂ (assuming a Redfield O₂:C ratio of 138:106) at an average rate of 3.25 x 10¹⁵ mol O₂ yr⁻¹ — 61 to 205 times the estimated average rates of global ocean O₂ 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 McClimens 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; Goodyay 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$_2$S (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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