Testing the climate intervention potential of ocean afforestation using the Great Atlantic Sargassum Belt

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Ensuring that global warming remains <2 °C requires rapid CO2 emissions reduction. Additionally, 100–900 gigatons CO2 must be removed from the atmosphere by 2100 using a portfolio of CO2 removal (CDR) methods. Ocean afforestation, CDR through basin-scale seaweed farming in the open ocean, is seen as a key component of the marine portfolio. Here, we analyse the CDR potential of recent re-occurring trans-basin belts of the floating seaweed Sargassum in the (sub)tropical North Atlantic as a natural analogue for ocean afforestation. We show that two biogeochemical feedbacks, nutrient reallocation and calcification by encrusting marine life, reduce the CDR efficacy of Sargassum by 20–100%. Atmospheric CO₂ flux into the surface seawater, after CO₂ fixation by Sargassum, takes 2.5–18 times longer than the CO₂-deficient seawater remains in contact with the atmosphere, potentially hindering CDR verification. Furthermore, we estimate that increased ocean albedo, due to floating Sargassum, could influence climate radiative forcing more than Sargassum-CDR. Our analysis shows that multifaceted Earth-system feedbacks determine the efficacy of ocean afforestation.
Capturing atmospheric CO₂ using ocean afforestation is receiving widespread attention in scientific and public discourse, due to the prospects of converting seemingly unproductive open ocean deserts into thriving seaweed ecosystems. Indeed, high-level political and economic stakeholders are beginning to advocate the rapid implementation of offshore seaweed farming, even though little is known about the CDR potential and side-effects of ocean afforestation when upscaled. Model simulations are the first step to explore these unknowns, but they inevitably miss many complexities associated with real-world systems. Therefore, performing in situ experiments must follow to inform model developments. However, most of what is experimentally and logistically feasible is orders of magnitude lower than the scale envisioned for climate-relevant ocean afforestation.

Studying natural analogues presents a low-risk, cost-effective way to evaluate the potential of ocean afforestation on the scale needed to contribute significantly to CDR. Indeed, there are prominent examples where studying natural analogues of climate intervention, such as the Mount Pinatubo volcanic eruption or glacial–interglacial ocean iron fertilisation, has led to breakthroughs in our understanding of the response of the Earth System to perturbation. The recent blooms of floating seaweed *Sargassum* in the (sub)tropical North Atlantic provide the first natural analogue for seaweed farming distributed across multiple open ocean regions. While the extent of these patchy *Sargassum* blooms (up to 6100 km²) are still below the maximum extent envisioned by some for ocean afforestation (e.g., 9% of the surface ocean), they are orders of magnitude larger than anything that is experimentally feasible. Thus, the recent emergence of this Great Atlantic Sargassum Belt (GASB) offers the unique, perhaps the only, opportunity to study large-scale ocean afforestation under real-world conditions before its potential application.

Floating *Sargassum* seaweed was historically found in the Sargasso Sea, but extended its bloom formations from 2011 onward across the Intra-Americas Sea and the (sub)tropical North Atlantic from West Africa to South America (Fig. 1a–c). This extension may have been initiated by an extreme negative phase of the North Atlantic Oscillation during 2009–2010 which shifted regional wind patterns. The subsequent development of the GASB is hypothesized to be sustained by increased nutrient runoff from the Amazon River. The *Sargassum* bloom, detected using satellites, follows a seasonal cycle with highest biomass in summer (Fig. 1d). The largest GASB bloom was observed in 2018 covering up to 6100 km² in numerous rafts over a ~9000 km belt just north of the Equator. The 2018 bloom commenced in November 2017 and peaked in June 2018 with a net build-up of 0.81 million tons (Mt) of particulate organic carbon (POC; Fig. 1d and Supplementary Table 1). This constitutes a lower bound as satellites neither detect small *Sargassum* rafts nor biomass submerged transiently by wind-driven ocean circulation, which could increase *Sargassum* stocks. POC formation in the GASB could be viewed as a first-order estimate for its capacity to absorb and sequester CO₂. However, this CDR potential will be modified through biogeochemical feedbacks. Here, we utilize insights from the GASB to elucidate the complex influence of calcification and nutrient reallocation on ocean afforestation.

**Results and Discussion**

Biogeochemical feedbacks reduce CDR by ocean afforestation. POC formation via photosynthesis consumes CO₂. In contrast, the formation of particulate inorganic carbon (PIC) through calcification (e.g., $\text{Ca}^{2+} + \text{CO}_3^{2-} \rightarrow \text{CaCO}_3$) releases CO₂ by reducing seawater alkalinity. Thus, to accurately assess the CDR potential of the GASB, we must subtract the CO₂ formed via calcification from the CO₂ consumed via photosynthesis. *Sargassum* and other fleshy seaweeds do not calcify themselves, but provide a habitat for colonising epibiont calcifiers such as bryoazas. Carbonate biominerals attached to *Sargassum* collected from the Sargasso Sea contribute 9.4% on an annual average to its wet weight biomass, equivalent to a PICO:POC ratio of ~0.25 (mol:mol). This reduces the CO₂ removal generated through photosynthetic POC formation by ~17% (Fig. 2a, c). It is currently unclear if 9.4% is applicable for the new *Sargassum* blooms occurring since 2011 in the GASB, or for other seaweeds potentially used for ocean afforestation. Slower/faster growing seaweeds may provide more/less time for epibiotic calcifiers to settle, which would affect the PIC:POC ratio accordingly. Over the range of wet weight CaCO₃ percentages reported for individual *Sargassum* samples from the Sargasso Sea (i.e., 4.3–21.4%), the PICO:POC is 0.11–0.9 and the offset to CO₂ removal 7–57% (Fig. 2c). This indicates that the calcification offset could range from being negligible to being a major factor reducing the CDR efficiency of ocean afforestation.

Another reduction in the CDR potential of the GASB is due to reallocation of nutrient resources from phytoplankton to *Sargassum*. Photosynthesis by *Sargassum* consumes nutrients which would otherwise fuel phytoplankton photosynthesis. Therefore, we must subtract the amount of natural phytoplankton CDR that would have been possible, with the amount of nutrients reallocated to *Sargassum* CDR. In the open (sub)tropical North Atlantic, Nitrogen (N) limits primary production while Phosphorus (P) and Iron (Fe) are occasionally co-limiting. The 2018 GASB consumed 2.7, 0.12, and 0.003 Gmol of N, P, and Fe, respectively. With these resources, phytoplankton could photosynthesise 0.26 Mt POCC (assuming N-limitation and a C:P ratio of 170 mol:170) and a C:N ratio of 8 mol:170, thereby reducing *Sargassum* CDR by 32% (31% reduction assuming P-limitation and a C:P ratio of 170 mol:170; Supplementary Fig. 1). A key question is whether phytoplankton POC formation and the associated natural CDR would be equally effective as CDR by *Sargassum*. Phytoplankton in the open (sub)tropical North Atlantic maintain surface nutrient concentrations close to zero, which is indicative of a ~100% efficiency of the phytoplankton carbon sequestration in this region. Thus, arguably, all phytoplankton POC that would have been fixed with the nutrients utilised by *Sargassum* must be subtracted from the *Sargassum* CDR, although this discount may be mitigated by the use of external nutrient sources (see also Supplementary discussion 1).

Nutrient reallocation from phytoplankton to *Sargassum* not only reduces phytoplankton POC, but also its PIC formation (i.e., calcification) and therefore the associated CO₂ formation. Hence, less CO₂ is formed by phytoplankton calcification because of the nutrient reallocation to *Sargassum*, and this saved amount of CO₂ must be added to the CDR potential of *Sargassum*. However, this addition is minor because the planktonic PIC:POC is small in the (sub)tropical North Atlantic (~0.01) relative to *Sargassum* assemblages (PIC:POC ~0.25; see above).

Considering the offsets through calcification and nutrient reallocation we define the theoretical CDR potential of ocean afforestation as:

$$\text{CDR}_{\text{theoretical}} = \text{POC}_{\text{seaweed}} - \text{PIC}_{\text{seaweed}} - \text{POC}_{\text{plankton}} + \text{PIC}_{\text{plankton}} \quad \text{(1)}$$

with the associated amounts of CO₂ bound/released given in Mt C. With respect to the 2018 GASB, POC$_{\text{seaweed}}$, PIC$_{\text{seaweed}}$, POC$_{\text{plankton}}$ and PIC$_{\text{plankton}}$ are 0.81, 0.13, 0.26, and 0.002 Mt C, respectively. Accordingly, the 2018 GASB generated a CDR$_{\text{theoretical}}$ of ~0.42 Mt C due to the higher organic C:N ratio of *Sargassum* (C:N = 25) than phytoplankton biomass (C:N = 8).
epibiont calcification. The average C:N ratios (mol:mol, range in parentheses) of other brown, red, and green seaweeds are 19 (8–55), 19 (6–78), and 14 (7–24), respectively, indicating that *Sargassum* is within the higher C:N range of seaweeds and closer to the upper bound of CDR<sub>theoretical</sub> achievable with ocean afforestation (see also Supplementary discussion 1).

Another addition to CDR<sub>theoretical</sub> by *Sargassum* is the photosynthetic formation of dissolved organic carbon (DOC) and its subsequent release into seawater. Indeed, *Sargassum* may have produced ~1 Mt DOC during the 2018 GASB which exceeds the build-up of POC of 0.81 Mt C (Supplementary Table 1). However, seaweed-DOC mixes with the background DOC pool, is displaced by ocean currents, and generally only ~2% of DOC escapes remineralization for ≥20 years. Therefore, accounting for CDR via DOC requires dedicated monitoring of vast ocean volumes with in situ sensors, that can differentiate DOC sources. So long as such monitoring is unfeasible, DOC may not be considered in CDR<sub>theoretical</sub> because accountability and independent verification of DOC storage are likely essential to finance ocean afforestation via carbon trading.
Hence, seawater containing a CO₂-deplete Sargassum range of particulate organic carbon (POC) formed by plankton (POCplankton, blue); Percent reduction due to PIC:POC observations, and the cross the C:N axis has no meaning. The solid symbols on the horizontal lines are averages (Sargassum average = 24.8). c Summary of discounts and additions to CDRtheoretical due to calcification and nutrient reallocation following Eq. 1 (upper and lower bounds in square brackets). Percent reductions due to PIC associated with seaweeds (PICseaweed, grey); Percent increase due to PIC associated with plankton (PICplankton, black).

Multiple challenges to verify CDR by ocean afforestation. To this point, we have quantified how much CO₂ dissolved in seawater is fixed by Sargassum during the 2018 GASB (i.e., CDRtheoretical). Ultimately, however, we need to quantify the permanent influx of atmospheric CO₂ into the oceans after afforestation has generated a seawater CO₂ deficit. As a first step, this requires traceable air-sea CO₂ transfer following CO₂ fixation by seaweeds. We calculated air-sea equilibration timescales (τCO₂) for the locations of pronounced Sargassum occurrence in the GASB region following Jones et al. (see ref. 35) and found that these range between 2.5–18 months (mean = 5.5; Fig. 3). This is 2.5–18 (mean = 5.5) times longer than the modelled residence time of seawater in the surface mixed-layer (τres) over much of the GASB region, which is between 0.3–1.5 months (mean = 0.9; Fig. 3b, c; see ref. 35; but note that large datagaps occur in the Intra-Americas Sea). For the entire GASB region between 5°S–25°N and 89°W–15°E, τCO₂ is 2–46 (mean = 8) times longer than τres, which is between 0.2–2 months (mean = 0.9; Fig. 3b, c; see ref. 35). Hence, seawater containing a CO₂-deficit generated through afforestation has a high chance to lose contact with the atmosphere before the deficit is fully replenished with atmospheric CO₂ (i.e., before equilibration is complete). Complete equilibration will likely occur some time in the future when seawater is eventually reexposed long enough to the surface but this may take >100 years. The potential time-lag raises the question at what time-point can CDRtheoretical be considered to be realized, because CO₂-fixation and seawater CO₂-absorption can be spatially and temporarily uncoupled. In practice, CDR will likely only be rewarded in a carbon trading system when atmospheric CO₂ uptake is accountable and independently verifiable. Hence, a key challenge for the implementation of ocean afforestation (and other marine CDR methods) on a carbon trading market is to quantify the CO₂ influx into the ocean, as this requires sophisticated measurements of seawater CO₂ uptake over large spatial and temporal scales.

After quantifying air-sea CO₂ transfer, seaweed carbon must be stored in a reservoir isolated from the atmosphere. The two storage methods widely anticipated are (i) biomass combustion followed by underground CO₂ storage (BECCS) or (ii) biomass deposition on the deep seafloor. Underground storage in appropriate geological formations comes with additional offsets to CDRtheoretical for example through shipping Sargassum biomass (0.0014–0.0017 t CO₂·wt−1·km−1) or through CO₂ separation after combustion (e.g., ~10–20% in the case of BECCS). Seafloor deposition may be associated with less process-related offsets to CDRtheoretical when the deep sea is adjacent to the afforestation site. The value of geological and seafloor carbon deposition depends upon a yet to be established policy framework that needs to clarify how sequestration permanence is factored into a carbon price, and how stringently CO₂ leakage back into the atmosphere must be monitored. Geological sequestration in appropriate formations can be considered long-term with a 66–90% chance of less than 1% CO₂ leakage over 1000 years and possibly beyond. Seafloor deposition will be less permanent than that. Sargassum sinks at 2500 m/d when its flotation bladders are removed, so that there is little time for degradation until reaching the seafloor (see also Supplementary discussion 2). However, on the deep seafloor and within its sediments, >90% of deposited carbon is typically remineralized and released back into the water column. Thus, a significant fraction will eventually be transported back to the surface as respired CO₂ by ocean circulation and mixing. Data-constrained modelling suggests that the zonally-averaged mean time from deep seafloor remineralization to re-exposure to the atmosphere ranges between 700–900 years in the North Atlantic, where GASB biomass is deposited, and increases further along the global conveyor belt to >1400 years in the North Pacific. These timescales are generally longer than those considered for some land-based CDR methods like terrestrial afforestation or soil carbon sequestration, which are on the order of decades to a century. Nevertheless, the systematic differences reveal that some locations are better suited for seafloor deposition of seaweed biomass than others. The necessary monitoring of inevitable CO₂ leakage appears at least equally challenging as the monitoring of air-sea CO₂ uptake (see previous paragraph), because relatively small amounts of respired CO₂ must be traced in a vast volume and be distinguished from the background CO₂ pool. This monitoring challenge will likely influence the feasibility to implement seafloor seaweed deposition in carbon trading.

Albedo climate feedbacks of ocean afforestation. In addition to CO₂-related feedbacks on green-house radiative forcing, afforestation also influences radiative forcing by changing the Earth’s albedo, the ratio between reflected and incident solar flux at the Earth surface. Afforestation in terrestrial environments reduces the albedo because forests are usually darker and reflect less
short-wave radiation than other landscapes\textsuperscript{11}. This negative feedback can substantially reduce or even offset the cooling potential of terrestrial afforestation generated by CDR\textsuperscript{41–43}. In the case of ocean afforestation, albedo increases because seaweeds reflect more short-wave radiation than seawater, especially when they occur near the sea surface\textsuperscript{44}. We estimate that (sub)tropical ocean afforestation of the area of the 2018 GASB (\textasciitilde6100 km\textsuperscript{2}) would reduce radiative forcing in the range of 180 (seaweed at \textasciitilde5 m depth) to 1800 (seaweed floating at the surface) peta J/y. This exceeds an estimated reduction of radiative forcing of 42 peta J/y caused by 0.42 million tonnes of CDR\textsuperscript{theoretical}. Accordingly, the magnitude of these numbers suggest that ocean afforestation would also need to be assessed in the context of solar radiation management, i.e., the deliberate reflection of solar radiation into space. However, ocean afforestation is typically envisioned with benthic seaweeds cultured several meters below the surface to avoid storm damage\textsuperscript{14}. Such culturing practices would alleviate the albedo effect\textsuperscript{44}. Furthermore, our estimates do not consider indirect albedo feedbacks, such as the release of bioaerosols, which may counteract or enhance the albedo effect\textsuperscript{45}. Indeed, identifying and accounting for all albedo feedback mechanisms induced by ocean afforestation will likely be as challenging as the comprehensive assessment of CDR (see Supplementary discussion 3).

Synthesis and ramifications. The GASB analysis sets realistic bounds on ocean afforestation CDR and provides the first estimates of the relevance of the direct albedo effect. Two biogeochemical feedbacks, calcification and nutrient reallocation, reduce the theoretical CDR potential of ocean afforestation. Propagating both the upper and lower bounds for these feedbacks indicates that the CDR\textsuperscript{theoretical} of the 2018 GASB ranges from \textasciitilde0.03--0.8 Mt C. Accordingly, ocean afforestation at the scale of the GASB could constitute a net CO\textsubscript{2} source or, at best, contribute 0.0001--0.001\% to the amount of annual CDR required in 2100 under a low emission scenario. The large range of estimates—from positive to negative—underlines that CDR with ocean afforestation is heavily dependent on feedbacks with the Earth System, which need to be further constrained. In addition, our analysis confirms that verification of both air-sea CO\textsubscript{2} flux and the permanence of carbon storage remain major challenges. Thus, significant physical and chemical oceanographic questions must be answered, along with establishing governance rules on verification and permanence to fully evaluate ocean afforestation. Moreover, the influence of ocean afforestation on ocean albedo could be more relevant than the climatic effect of CDR. However, albedo enhancements are currently not considered with carbon trading and thus would need additional legislation to be incentivised and rewarded\textsuperscript{46}. Last, all the abovementioned feedbacks on the net climatic effect must also be evaluated for an environment that is constantly being modified due to ongoing climate change.

The assessment of terrestrial afforestation is further advanced than that of ocean afforestation, because ubiquitous natural analogues (forests) provide constraints on its efficacy. Thus, there is substantial knowledge of how Earth-system feedbacks modify the apparent CDR potential of terrestrial afforestation (see the wider discussion around ref. \textsuperscript{47}). The GASB provides a new opportunity to constrain ocean afforestation, and our analysis points towards a similar conclusion as for terrestrial afforestation; i.e., climate intervention through ocean afforestation must consider complicated Earth-system feedbacks (Fig. 4), which could influence the sign and magnitude of its cumulative climatic effect.

Acceleration in CDR research and development is urgently needed as global negative emissions must be upscaled to gigatonnes within this decade (Supplementary Fig. 2). This requires rapid identification of tractable CDR solutions, which enable accurate and ongoing accounting of their overall climatic influence. The GASB analysis reveals that the net climatic impact of ocean afforestation is associated with major uncertainties, largely due to the inherent complexity of biological systems (Fig. 4). Other marine biological CDR approaches such as ocean iron fertilization exhibit similar complexity, and two decades of investigation has revealed comparable uncertainties about their net climatic impact. These ongoing unknowns raise the question of whether CDR using marine biota has the potential to be sufficiently well understood within this decade to prioritize development. Instead, the complexity associated with such CDR approaches may provide a compelling argument to focus on bottom-up engineered and better understood abiotic methods.

Methods

Build-up of seaweed biomass in the Great Atlantic Sargassum Belt (GASB). Floating Sargassum biomass in the GASB has been quantified with satellites\textsuperscript{47,48}. We downloaded monthly mean cumulative Sargassum wet weight data for the GASB region (Caribbean Sea and Central Atlantic Ocean) from ref. \textsuperscript{16} under the following link (https://www.ncei.noaa.gov/access/metadata/landing-page/bin/iso?id=gov.noaa.nodc:0190272). The cumulative wet weight biomass was converted to total particulate carbon (TPC) by multiplication with the wet weight to TPC conversion factor of 0.0543 (g/g) given in the ref. \textsuperscript{49}. Supplementary Table 1 provides wet weights\textsuperscript{16} and corresponding TPC values for the 2018 growth cycle, which commenced in November 2017 and constitutes our natural analogue case study for ocean afforestation.
The next step was to calculate how many moles CO\(_2\) would be formed per mol PIC precipitated by the epibionts. This ratio, known as psi, was calculated following Frankignoulle et al.\(^{18}\) using the psi-function of the carbonate chemistry calculation software seacarb\(^{53}\) for R with recommended default settings for carbonate chemistry constants\(^{53}\) (e.g., K1 and K2 from ref. \(^{49}\)). For this calculation, we assumed subtropical conditions with a total alkalinity of 2350 \(\mu\)mol kg\(^{-1}\), a dissolved inorganic carbon concentration of 2047.5 \(\mu\)mol kg\(^{-1}\), a salinity of 35, and a temperature of 25°C (the following expression was used: psi = psiflg = 15, var = 2350c–6, S = 35, T = 25, P = 0, Pr = 0, St = 0, pHcalc = “T”, kf = “p”, k12 = “T”, ks = “d”). In this case psi is 0.63.

Finally, we multiplied psi with the mols of PIC precipitated by Sargassum epibionts and compared the amounts of generated CO\(_2\) with the amounts of fixed CO\(_2\) by photosynthetic PIC production: 

\[
\text{Calcification offset} = \frac{\text{PIC} \times \psi}{\text{POC}}
\]

(3)

According to this, the calcification offset is 0.165 or 16.5% to the amount of POC fixed photosynthetically by Sargassum. The key assumption in this calculation is that 9.4% (ranging from 4.3 to 21.4%)\(^{20}\) of Sargassum wet weight is CaCO\(_3\), which is based on measurements with Sargassum collected in the Sargasso Sea\(^{20}\). These values provide a useful first estimate but they may differ for Sargassum occurring in the GASB since 2011, or for other species used for ocean afforestation as pointed out in the discussion.

**Nutrient reallocation from resident phytoplankton to Sargassum.** Ocean afforestation requires macro-nutrients (Nitrogen (N) and Phosphorous (P)) and micro-nutrients such as iron (Fe), which are the primary limiting resources for primary production in the subtropical Atlantic\(^{21}\). Nutrients fixed by afforested seaweeds in offshore environments can no longer support carbon fixation by phytoplankton. Thus, nutrient reallocation from phytoplankton to seaweeds reduces phytoplankton carbon sequestration, a natural marine CO\(_2\) sink. The phytoplankton carbon sequestration that would have been possible with the reallocated nutrients must be subtracted from the CDR potential of ocean afforestation. In the following paragraphs, we go through the calculations and considerations made to assess the degree of reduction in seaweed CDR through nutrient reallocation.

The 2018 Sargassum bloom in the GASB commenced in November 2017 and peaked in June 2018 with a net biomass build-up of 18.8 Mt wet weight (Supplementary Table 1). Wet weight was converted into Mt N by multiplication with the Sargassum N wet weight ratio (g/g) reported in Wang et al. (–0.002)\(^{19}\). Next, the mass of N was converted into gigamoles (Gmol) by division with the molecular weight of N (i.e., 14.007 g/mol). The same calculation was done for P using the P:wet weight ratio of 0.0002 (g/g) and the molecular P weight of 30.974 g/mol. Fe removal was calculated the same way (molecular weight of Fe 55.845 g/mol), but we had to derive a Fe:wet weight ratio from other references. We used Fe: wet weight ratios of 1.099 × 10\(^{-3}\) (g/g) and the molecular P weight of 30.974 g/mol for the conversion. While this may not be fully representative for holoplagic Sargassum from the GASB, it may provide a useful number within a plausible range.

Based on these calculations, Sargassum growth during the 2018 GASB consumed 2.7, 0.12, and 0.003 Gmol N, P, and Fe, respectively. In the sub(tropical), Atlantic phytoplankton is mostly N-limited\(^{21}\). Assuming a typical organic matter C:N ratio of 8 in the (sub)tropical Atlantic\(^{21}\), phytoplankton could fix 0.62 Mt C with the amount of N bound in Sargassum. The C:N ratio for Sargassum as determined by Wang et al. (see ref. \(^{49}\)) is 31.3 (mol/mol). However, this value includes PIC from epibionts which needs to be subtracted assuming a wet weight percentage of CaCO\(_3\), of 9.4% (see previous section). With this correction the Sargassum organic C:N is 24.8 (i.e., (TPC/PIC)/N, where TPC and N is from Wang et al. (see ref. \(^{49}\)) and PIC calculated as described in the previous section) or 28.2–16.5 for wet weight CaCO\(_3\) percentages from 4.3 to 21.4% (see ref. \(^{20}\)). Thus, phytoplankton are able to fix around one third (i.e., 8/24.8 × 100 = 32%) of the carbon fixed by Sargassum seaweeds (or 28–48% for the range of CaCO\(_3\) wet weight percentages). The 32% must be subtracted from the 30% N-potential of afforestation, because phytoplankton carbon fixation with these nutrients would likely have occurred under the absence of the GASB. To test whether this percentage offset is sensitive to the assumption of N-limitation in the sub(tropical) North Atlantic\(^{21}\), we repeated the calculation assuming P-limitation. The molar C:P ratio of Sargassum is 550 (i.e., (TPC/PIC)/P where TPC and P is from Wang et al. (see ref. \(^{49}\)) and PIC calculated as described in the previous section) compared to 170 typically found in oligotrophic plankton communities.\(^{21}\). Thus, as for N-limitation, phytoplankton are able to fix around one third (i.e., 170/550 × 100 = 31%) of the carbon fixed by Sargassum seaweeds when assuming P-limitation (Supplementary Fig. 1) (or 26–45% for the range of CaCO\(_3\) wet weight percentages). As the differences between N-limitation or P-limitation are marginal, our calculations on N ratios in our study.

**Dissolved organic carbon (DOC) production.** Powers et al. (see ref. \(^{96}\)) conducted incubation experiments with Sargassum collected in the Sargasso Sea under in situ
light and temperature conditions, and reported DOC release rates of \(288 \pm 24 \mu \text{g C d}^{-1} \text{g wet wt}^{-1}\). Multiplying their daily rate with satellite derived biomasses yields a cumulative Sargassum DOC production for Nov. 2017–Dec. 2018 of 1.07 Mt C (Supplementary Table 1). Please note that this estimate is associated with several uncertainties that are difficult to quantify\(^{36}\). While the uncertainty of the reported rate is relatively minor (SD \(24 \pm 24 \mu \text{g C d}^{-1} \text{g wet wt}^{-1}\)), it has been determined only during a certain time of the year (September) when Sargassum was in a distinct growth stage. As mentioned by Pornat Assimilations\(^{22}\), DOC production rates may be different for early or late season Sargassum, and also change with the environmental conditions such as light and temperature.

**Air-sea \(\text{CO}_2\) exchange and equilibration timescales.** To generate atmospheric \(\text{CO}_2\) removal, the \(\text{CO}_2\) deficit generated through afforestation needs to be balanced by atmospheric \(\text{CO}_2\) influx into the ocean. Air-sea \(\text{CO}_2\) fluxes in the open ocean primarily depend on wind speeds (\(U\)) and the air-sea \(p\text{CO}_2\) difference. Temperature (\(T\)) and salinity (\(S\)) also play a role as they influence the Schmidt number (\(Sc\)) and the solubility (\(K\)) of \(\text{CO}_2\) in seawater\(^{37}\). We followed the approach by Jones et al. (see ref. 35) to calculate the timescale of air-sea equilibration (\(\tau_{\text{CO}_2}\)) as:

\[
\tau_{\text{CO}_2} = \frac{h \times R \times G}{N \times B}
\]

where \(h\) is the mixed layer depth (in \(m\)), \(R\) is the ratio of dissolved \(\text{CO}_2\) + \(\text{H}_2\text{CO}_3\) to total dissolved inorganic carbon (DIC), \(G\) is the gas transfer velocity (in \(m/s\)), and \(B\) is the Revelle Factor. In a first step, we calculated \(\tau_{\text{CO}_2}\) for a range of plausible scenarios for (sub)tropical environments, where ocean afforestation could be applied. In these scenarios, we assumed typical mixed layer depths ranging from 10 to 110 m\(^{38}\). \(R\) and \(B\) were calculated with the seaface package using the carb and buffergen functions with recommended default settings for equilibrium constants\(^{53}\).

Furthermore, Jones et al. and both analyses show the same regional features (compare Fig. 3 main text, Supplementary Fig. 5) are generally in good agreement with those calculated by Takahashi et al.\(^{59}\), with all data, noting that the time periods are different for these data and the Takahashi climatology\(^{59}\). The residence time of seawater in the surface mixed layer was provided by Jones et al. (see ref. 35) from their simulation of an idealized surface age tracer using circulation fields from the Estimating the Circulation and Climate of the Ocean– Global Climate Model\(^{35}\). The seasonality in coverage as shown in Supplementary Table 2; refs. 13, 16). The seasonality in coverage as shown in Supplementary Table 2; refs. 13, 16). The seasonality in coverage as shown in Supplementary Table 2; refs. 13, 16). The seasonality in coverage as shown in Supplementary Table 2; refs. 13, 16). The seasonality in coverage as shown in Supplementary Table 2; refs. 13, 16).

**Albedo modification by Sargassum.** Albedo is defined as the ratio between reflected and incoming solar flux at the Earth surface\(^{60}\). It is given as a dimensionless number between 0 (absorption of all incident solar flux) and 1 (reflection of all solar flux). Seawater has a relatively low albedo compared to other surfaces\(^{60}\). Therefore, marine vegetation on top or slightly below the sea surface increases the albedo\(^{61}\). In the following, we estimate how much the albedo enhancement caused by afforestation could reduce the radiative forcing. As mentioned by Pornat et al.\(^{22}\), the key to estimate how much radiative forcing is reduced by \(\text{CO}_2\) through afforestation. Last, we compare the reductions of radiative forcing. The calculations described below are based upon equations by Betts\(^{41}\) and Kirschbaum et al.\(^{43}\), who did a similar assessment in the context of terrestrial afforestation.

The 2018 GASP had initially a surface Sargassum coverage of \(\sim 870\, \text{km}^2\) in December 2017 (Supplementary Table 2). It then extended to a maximum of 6093 km\(^2\) in June 2018 after which coverage declined to \(\sim 807\, \text{km}^2\) (Supplementary Table 2; refs. 13, 16). The seasonality in coverage as shown in Supplementary Table 2 prescribe the growth cycle for our ocean afforestation scenario, where we assume that (i) seaweed farms of 6093 km\(^2\) are maintained north of South America and the GBAS region (lat \(\sim 10°\)N; 60°W), (ii) growth occurs during approximately the first half of the year, and (iii) harvest/processing during the second half matching biomass build-ups and declines as in Supplementary Table 2. The (sub)surface circulation of marine vegetation increases the albedo of seawater\(^{62}\), thereby decreasing the daily average radiative forcing (\(\text{RF}_{\text{daily}}\)) calculated as:

\[
\text{RF}_{\text{daily}} = Q_{\text{i}} \times 1.2 \times (1 - \text{albedo})
\]

Here, \(Q_{\text{i}}\) is the daily total downward solar flux (in \(J \text{m}^{-2} \text{d}^{-1}\)). At the change in albedo over the shortwave spectrum (i.e., \(\sim 0.28-2500\, \text{nm}\) due to seaweed cover, and \(\text{albedo}\) is the proportion of shortwave-radiation absorbed by the atmosphere \(\sim 20\%\)\(^{43,59}\). \(Q_{\text{i}}\) data was obtained from the Giovanni online data system, developed and maintained by the NASA GES DISC. More specifically, we downloaded maps of monthly averages (December 2017-November 2018) of Surface incoming shortwave flux (0.5–6.25°) from the MERRA-2 Model M2TMXRAD v5.12.4. (This data is provided in W m\(^{-2}\) and was multiplied with 1000 to convert to J m\(^{-2}\) d\(^{-1}\)). Afterswards we extracted representative monthly incoming solar fluxes for the anticipated study period (lat \(\sim 10°\)N; 60°W). \(Q_{\text{i}}\) changes over the course of a seasonal cycle but is relatively...
stable in the low latitude considered in our scenario, ranging from 180 to 280 W m
−2 (Supplementary Table 2, equivalent to 15.5–24 MJ m−2 d−1). The increase of albedo, as the seagrass spread, has, to the best of our knowledge, not been determined so far. However, there are assessments of Δa due to seagrass in shallow water published by Fogarty et al. (see ref. 44). These authors found that albedo of seagrass meadows is by about 0.01–0.07 higher than open water albedo with the highest increase observed, when the coverage of the canopy was dense and close to the surface. This highest they considered was 0.25 m2. Therefore, it is not fully representative for our Sargassum analogue, Fogarty et al.’s data (see ref. 44) provides a useful first indication how much seagrass could increase albedo relative to open ocean seawater. Arguably, Sargassum rafts would be on the upper end or above of the 0.01–0.07 range, because they float right at the surface and have a very dense coverage45.

Applying Eq. (6), we first calculate cumulative monthly ΔRF (i.e., ΔRF_month days of the month) and then cumulative yearly ΔRF (i.e., sum of monthly ΔRF). We do this calculation for a range of plausible Δa (0.01–0.1)44, to account for different types of ocean afforestation where seaweeds could grow at different depths in the water column, even at the surface like Sargassum (Supplementary Tables 2 and 3). In all scenarios, ΔRF_year decreases ranging from 181–1811 PJ y−1.

The reduction in radiative forcing per year−1 due to the removal of 1 tonne of carbon from the atmosphere (ΔRF_year in y−1) can be calculated as:

\[
\Delta RF = \frac{86,400 \times 365 \times 5.35 \times 10^2}{C_{CFLS} \times 510^2},
\]

where \(C_{CFLS}\) is the CO2 concentration in the atmosphere (we use 410 p.p.m.v. for our calculations), 86,400 is the number of seconds per day, 365 the number of days per year, 510 the surface area of the Earth, and 2.10 \(\times 10^{12}\) is the tonnes of carbon that lead to a 1 p.p.m.v. increase of atmospheric CO2 (see ref. 33). The factor 5.35 and 510 account for the increase of radiative forcings due to increasing atmospheric CO2 (see ref. 71). According to Eq. 5, ΔRF equals −99 GJ IC−1 y−1. This number is slightly lower than −104 GJ IC−1 y−1 calculated by Kirschbaum et al.45 because they used a \(C_{CFLS}\) of 390 p.p.m.v. and the climate sensitivity is a logarithmic function of \(C_{CFLS}\) (see ref. 71).

The assessment shows that ocean afforestation increases albedo thereby constituting a positive feedback in addition to the reduction of radiative forcing caused by CDR. The CDR effect on radiative forcing, i.e., −99 GJ IC−1 y−1 × 10^{20} t C = −41.6 PJ y−1, is considerably smaller than the albedo effect during the first years (420,000 t C = CDRtheoretical of the 2018 GSB; see main text). However, the reduction of radiative forcing through CDR is cumulative from season to season, so that it eventually becomes more important than the albedo effect. In our seaweed farms totaling 6093 km2 in the tropics, CDR would outweigh the albedo effect after ~181–1811 to ~1811–1811.6 (i.e., 4.4–44) annual seaweed growth cycles assuming instantaneous (and therefore unrealistic) atmospheric CO2 invasion into seawater. Please note that this calculation of timescales neglects the change in \(C_{CFLS}\) due to ongoing emissions and changes in atmospheric CO2 as this has only a small influence on the relatively short timescales considered here.

Propagating the upper and lower bounds for the reduction of CDRtheoretical through calcification and nutrient reallocation. In this section we propagate the wide range of assumptions for the calcification and nutrient reallocation effects to estimate the upper and lower bounds of CDRtheoretical of afforestation with Sargassum. The range of assumptions are constrained by published data as described in methods on calcification and nutrient reallocation. The goal of such propagation is to ensure that the best-case and worst-case scenarios for ocean afforestation are represented to provide a balanced viewpoint. Furthermore, we aim to illustrate the degree of uncertainty of the theoretical CDR potential of ocean afforestation due to these biogeochemical feedbacks. Our estimation is based on Eq. 1.

To estimate the lower bound of CDRtheoretical, we first assume an arbitrary amount of 100 mol TPC associated with Sargassum. Assuming the maximum of CaCO3 contribution Sargassum wet weight by epibenthic calcifiers of 21.4%49 yields an upper value PIC:POC ratio of ~0.9 (mol:mol). Thus, 100 mol TPC would be split in ~47 mol PIC and ~53 mol POC. The 53 mol of POC would bind the same amount of CO2 as to what the formation of 47 mols PIC would release ~29.6 mols of CO2 (i.e., PICseaweed calculated as 47 psii). Accordingly, CDRtheoretical would decrease from 53 to 23.4 mols CO2 in this high calcification scenario. The 53 mols of POC would require ~3.3 mols of nitrogen assuming a lower bound C:N ratio of 16 (mol:mol) reported for holopelagic Sargassum fluitans22. These amounts of N would support 26.5 mols of POC fixation by phytoplankton assuming a phytoplankton C:N ratio of 8 (see ref. 23). Thus, CDRtheoretical would decrease by another 26.5 mols (POCplankton) and become slightly negative (i.e., −3.3−29.6−26.5 = −3.1). However, nutrient reallocation from phytoplankton to Sargassum would also reduce calcification by phytoplankton. The planktonic PIC:POC ratio in the (sub)tropical Atlantic is ~0.01 (see ref. 46, 25−27), so that 26.5 mols of POC would be associated with 0.265 mols PIC. This phytoplankton PIC is not formed due to the nutrient per se reallocation to Sargassum which saves ~0.17 mols of CO2 (PICplankton) from being released (0.265 psii). Overall, the CDRtheoretical would be close to zero under the abovementioned lower bound assumptions (i.e., 53 − 29.6 − 26.5 + 0.17 = −3 mols for 100 mol TPC initially fixed by the Sargassum community). For the 2018 GSB with a 1.02 Mt TPC (Supplementary Table 1), this would be ~0.03 Mt C (formation of ~0.1 Mt CO2).

To estimate the upper bound of CDRtheoretical we do exactly the same calculations but with upper bound assumptions: CaCO3 contribution Sargassum wet weight by epibenthic calcifiers of 4.3%20 leading to a Sargassum PIC:POC of ~0.11 (mol:mol). CN Sargassum = 108 mol:mol22 CN phytoplankton = 8 (mol:mol, same as above); planktonic PIC:POC = 0.01 (mol:mol, same as above). Under these conditions, 10 of the initially 100 mol TPC formed in Sargassum habitats would be present as PIC and release 6.3 mol CO2 (i.e., PICrelease). The 90 mols of POC would require 0.83 mols of N. This would support 6.6 mol of phytoplankton PIC (i.e., POCplankton), which would form 0.066 mols of PIC thereby releasing 0.04 mol CO2. Thus, the upper bound CDRtheoretical is 90 − 6.3 − 6.6 + 0.04 = 77 mols for 100 mol TPC initially fixed. For the 2018 GSB with a production of ~1.02 Mt TPC (Supplementary Table 1), this would be 0.79 Mt C (consumption of ~2.9 Mt CO2).

The amount of CO2 removal to limit global warming below 2 °C, and the hypothetical contribution of the GASS to these CO2 removal targets. Results from simulations of shared socioeconomic pathways (SSPs) 1–5 with Integrated Assessment Models (IAMs) were downloaded in November 2018 from the database of the International Institute for Applied Systems Analysis (IIASA) under the link https://secure.iiasa.ac.at/web-apps/ene/SspDbAction.html?page=about. The database was filtered for SSP scenarios with an additional temperature forcing of 42 W m−2 where global SSP2 (SSP2.6) where global SSP1 (SSP1.2) was below 2 °C (transient temperature overshoots above 2 °C were not considered as an exclusion factor). This yielded 32 SSP2.6 simulations. Afterwards, we examined each of the 32 simulations for the variable: Emissions [CO2]Carbon Capture and Storage | integration period | calculation and nutrient reallocation.

Data availability

All calculations, equations, and applied software are detailed and referenced in the methods and supplementary material.

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Author contributions

L.T.B., V.T., and P.W.B. designed the study. L.T.B. and V.T. analysed the data. L.T.B. drafted and revised the manuscript. P.W.B. and V.T. worked on the different versions of the manuscript. I.G., C.L.H., and J.A.R. commented on and contributed to the improvement of the manuscript.

Competing interests

The authors declare no competing interests

Additional information

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