Supplementary Material to

Can coastal and marine carbon dioxide removal help to close the emissions gap? Scientific, legal, economic and governance considerations

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## **Text S1 Cumulative emissions and the emissions gap**

Cumulative emissions of CO2 are directly related to stabilisation temperature in future climate scenarios (Meinshausen et al., 2009; Smith et al., 2018) so it is important when considering the significance of carbon management actions the contribution of such actions to the total atmospheric CO2 budget relative to the required mitigation/negative emission strategies to prevent exceeding a certain level of warming.

In order to remain close to the internationally agreed target of limiting warming to below 2°C, humanity needs to limit CO2 to well within the second most stringent of the IPCC climate scenarios (RCP 2.6) and to limit warming to 1.5°C, must keep concentrations in line with the even more stringent RCP 1.9. Within these temperature/CO2 concentration targets and associated possible atmospheric CO2 trajectories (i.e., when and how quickly do we start cutting emissions) there are many possible scenarios employing different balances of conventional mitigation (i.e., emissions reductions) and negative emissions (i.e., enhanced uptake through either environmental carbon management or technological fixes).

Estimates of emissions to 2100 that are consistent with RCP2.6 range from 300 to 1200 Gt C and the total 21st Century emissions allowing stabilisation at or below 2°C are estimated to be as little as 200 Gt C (733 Gt CO2e) (e.g., Smith et al., 2018; Millar et al., 2017). Gasser et al (2015) estimate that between 200 and 600 Pg of additional carbon storage (700–2100 Gt CO2e) will be needed to mitigate climate change to below 2°C of warming in their integrated assessment of emissions mitigation and earth-system response. The UN Emissions gap report 2019 (UNEP 2019) further estimates that there is a gap between projected emissions and maximum emissions consistent with limiting warming to 1.5°C of up to 32 Gt CO2e equivalents by 2030.

There are a range of estimates of the magnitude of the ocean net sink for atmospheric CO2 in the present but are on the order of, e.g., 2.9 ± 0.4 Pg of carbon per year (Friedlingstein et al., 2022). The greater the total cumulative emissions the smaller the fraction of these emissions that will be taken up by the ocean due to the loss of buffering capacity, particularly in the more rapid emission scenarios (e.g., Le Quéré et al., 2009; Jones et al., 2013) and thus there will be a reduction of uptake rate relative to emission rate through time. In the most optimistic scenarios where atmospheric emissions are rapidly reduced, the year-on-year total sink strength of the ocean will actually decrease due to the reduction in the concentration difference driving the flux of carbon into the ocean (e.g., Jones et al., 2013; 2016), with CMIP5 models predicting an ocean uptake of between 100 and 150 Pg of carbon from 2005 to 2100 under RCP2.5 scenarios, with an ocean fractional uptake of 0.6 of total anthropogenic CO2 emissions.

This overall reduction in the ocean sink with time as mitigation activity is implemented has important implications for negative emissions which decrease atmospheric CO2 – any benefit they confer is mediated by a decrease in the effectiveness of the natural ocean carbon sink (Jones et al., 2016). Therefore, it is difficult to assess the net benefit of any carbon uptake precisely in the context of a given emissions scenario without a detailed modelling effort including the timing of a given action relative to other activities and the response of the whole system to it.

Given the complexity of assessing emissions magnitudes and the further complexity of predicting the response of the natural system to future mitigation activity, we take a simplified approach here of comparing the carbon storage capacity of different coastal and marine systems to each other, in the context of the need to sequester hundreds of extra petagrams of carbon by the end of the century. We consider absolute carbon storage rather than the impact of carbon storage on atmospheric carbon as the latter is time-varying and depends on the buffer capacity of the ocean.

## **Text S2 Coastal (‘Blue’) Carbon storage scenarios**

There has been considerable study of the carbon storage capacity and loss rates of salt marshes, mangroves, and seagrass meadows globally and while knowledge is still lacking on precise areas and loss rates, and in understanding the variability of observed carbon stocks and storage (e.g., Williamson et al., 2022), this paper cannot add to this knowledge and so we take a synthesis of the current literature values as presented in Table S1.

For all habitats (saltmarsh, mangrove, seagrass, and macroalgae) we use the synthesised data in Table S1 to calculate scenarios as follows:

i) Business as usual: The current rate of loss of habitat (and the concurrent loss of carbon stock) is maintained to 2100, with the remaining habitat sequestering carbon at literature rates.

ii) Halt loss: All habitat is protected immediately so the loss is halted, and this constant area is maintained until 2100 (in reality this scenario would be achieved by counterbalancing unavoidable loss with new or restored habitat to maintain a constant area).

iii) Restoration: There is a year-on-year increase in habitat areas through aggressive habitat restoration to 2050 at which point pre-WWII areas have been restored. Each year the extant habitats fix carbon at the literature rates.

We note that neither i) nor ii) are realistically achievable but these are presented as optimistic ‘best cases’ (and serve to demonstrate the relatively small contribution coastal blue carbon can make even under these best-case scenarios).

**2.1 Macroalgal carbon storage**

There are fewer studies on macroalgae as carbon sinks so we synthesise these in more detail here. Whilst not leading to direct sedimentation of carbon in situ, it is thought that a small percentage of macroalgal primary production may be stored for long time periods as particulate organic carbon in shelf sediments or as refractory dissolved organic carbon in the marine carbon pool (e.g., Krause-Jensen and Duarte, 2016). This percentage is highly uncertain, however (from 0–6% percent of kelp primary production or 0–3% of other macroalgae; Legge et al., 2020; Krause-Jensen and Duarte, 2016). It is therefore potentially of no benefit to C sequestration, or of globally significant potential. In this study, we consider the range of carbon storage values presented by Krause-Jensen and Duarte (2016) of 61–268 TgC yr-1 and from these and their estimate of global area (3.54, million km2) infer a per-unit area net storage of 17–76 g C m-2 yr-1. We do not consider here the likely effect of increasing temperature on reduced sink strength of carbon from macroalgal detritus (Filbee-Dexter et al., 2022), but this is a potentially important issue for further consideration by future studies.

**2.2 Vulnerability of macroalgal storage**

Macroalgal standing stock and detrital export have been shown to be lower at warmer temperatures so a decrease in carbon sink might be expected as waters warm (Pessarrodona et al., 2018). Pessarrodona et al. (2018) find warm (southern UK) *Laminaria. hyperborea* had a standing C stock of ~140 g C m-2 (1.4 Mg C ha-1) versus 1200 g C m-2 (120 Mg C ha-1) in cold waters to the north of the UK (note these colder water stock values are the same order as for the sediments of a salt marsh or seagrass meadow, the difference being that the carbon is stored in the living biomass in a kelp forest, rather than in sediments of a salt marsh). We use these figures for the range of global per unit stock of carbon for macroalgae.

The current state of knowledge does not allow us to model future predictions of temperature response with any degree of confidence. However, Assis et al. (2017) observed a significant loss of macroalgae and a poleward shift of the southern edge of habitat extent around the coasts of SW Europe with a decrease between 1980 and 2014 of 30% by area with a further 38% decrease predicted to 2100. Similar predictions of regional loss are made elsewhere (e.g., NW Atlantic; Khan et al., 2018). We therefore estimate a 38% decrease in kelp carbon stock and thus carbon fixation by 2100) as species move North and the coldest water species (with the highest C fixation) are lost. This corresponds to a year-on year loss of 0.6% of stocks. More detailed work is needed to better estimate this potentially significant change to the ocean carbon cycle in the future. Macroalgal loss due to development, harvesting, and other human activity is difficult to estimate and in a large-scale synthesis of kelp forest change over the preceding 50 years, Krumhansl et al. (2016) found highly variable rates of change, including many examples of kelp forest increases. However, they find an overall rate of decrease of 1.8% per year globally, which we take as an upper estimate of the loss rate to 2100.

There is no scenario where we can stop this temperature change, although limiting GHG emissions may prevent a worse case. With habitat creation and artificial substrate for kelp forests in water with the correct temperature regime, it might be possible to create artificial habitats (noting that seaweed aquaculture is extant and growing across the globe), so we envisage a ‘stop loss’ scenario where stock losses are mitigated by artificial aquaculture schemes. This additional kelp production far surpasses the contribution of seaweed aquaculture considered feasible by Duarte et al. (2017) based on current technology and the trajectory of growth of seaweed aquaculture (6% of present-day wild seaweed carbon storage by 2050 is their prediction) and would require significant investment and technological development, even more so for the ‘restore’ scenario, where the 30% lost since 1980 is restored by 2050.

**Table S1**. Extent, C stocks and storage fluxes for coastal ecosystems globally. Data as synthesised by Duarte et al. (2013) except where citation given or recalculation demonstrated. All data presented here is converted to units used by Duarte et al. (2013).

|  |  |  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- | --- | --- |
| **Ecosystem** | **Global area (km2)** | **Per unit area C storage rate** **(g C m-2 yr-1)** | **Per unit area stock** **(Mg C ha-1)** | **Global C storage rate (Tg C yr-1)** | **Global C Stock** **(Pg C)** | **Gobal loss rate** **(% yr-1)** | **Loss since WWII (%)** |
| Salt marshes | 50,000 – 200,000 (after McOwen et al., 2017) | 218 +/-24 | 162 | 9.7–48 (calculated from McOwen extent and Duarte storage) | 0.81–3.24 (calculated from McOwen extent and Duarte storage) | 1–2 | 25–50 (McOwen et al., 2017) |
| Mangroves | 138,000–152,000 | 163 | 255 | 23–25 | 9.4–10.4 (note Duarte et al. get this using maximum observed not average unit area stock – we re-calculate as 3.5 and 3.9) | 0.7–3 | 30–50 |
| Seagrass meadows | 177,000–600,000 | 138 +/- 38 | 140 | 48–112 (calculated from extent and unit area storage gives 17–106) | 4.2–8.4 (calculated as above gives 2.5–8.4) | 0.9 | 30 |
| Macroalgae   | 1,400,000–5,700,000 (3,540,000; Kraus-Jensen and Duarte, 2016) | 17–76 (inferred from Krause-Jensen and Duarte, 2016) | 1.4–120 (Pessarrodona et al., 2017; *L. hyperborea* specifically) | 61–268 (173; Kraus-Jensen and Duarte, 2016) |  0.5–42 (calculated using average area from column 1 and unit area stock range from column 3) | 0.6–1.8 (lower: Assis et al., 2017; upper: Krumhansl et al., 2018) | 30 (since 1980) |

**Table S2**. Uptake capacity of coastal habitats 2024–2100 in Gt C. Abbreviations: BAU – business as usual.

|  |  |  |  |
| --- | --- | --- | --- |
| **Habitat** | **BAU** | **Stop loss** | **Restore** |
| Salt marsh | –0.25 to 0.9 | 0.79 to 3.9 | 1.0 to 7.1 |
| Mangrove | –0.13 to –2.5 | 1.8 to 2.0 | 2.5 to 3.6 |
| Seagrass | –0.25 to 1.8 | 1.4 to 8.6 | 1.9 to 12 |
| Macroalgae | 1.8 to 3.7 | 4.9 to 22 | 6.7 to 29 |
| TOTAL | 1.2 to 4.0 | 8.9 to 37 | 12.1 to 59.7 |

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**Figure S1.** Results of scenario modelling for CBC systems under business as usual scenario.



**Figure S2.** Results of the 3 scenarios for CBC systems



## **Text S3. Shelf sea carbon management potential**

### 3.1 Near-shore macroalgal aquaculture

Here we consider the potential benefits of targeted intensive seaweed aquaculture to capture anthropogenic nutrient fluxes from rivers. Gruber and Galloway (2008) estimate 50 Tg-N of anthropogenic origin flows down rivers. The N input to shelf seas could conceivably increase 5-fold by 2100 in a high-bioenergy scenario, or reduce to 50% of the current value in a high nutrient management scenario (Riahi et al., 2017). We assume 50% of current N input can be captured by coastal aquaculture by 2030 (the ‘low hanging fruit’ in the study of Lehan et al., 2016).

Assuming 50% of this N could be transferred into macroalgal biomass in intensive ‘farms’ at river mouths, and a seaweed C:N of 18, this would represent (12/15)\*50\*18 = 720 Tg C yr-1. At a high productivity of 2,500 g C m-2 yr-1 (Lüning et al., 2003) this would require 720 x1012 / 2.5 x103 = 288000 km2 of the coastal ocean or a coastal strip 10 km wide and 30,000 km long. The global coastline is about 1x106 km long so about 3% of the global coastline would be used in this case. Given that near-shore seaweed aquaculture is already established technology we assume that linear growth to maximum capacity in 2050 is achievable and model the resulting C sequestration accordingly. However, at current growth rates Duarte et al. (2017) estimate a maximum CO2 sequestration capacity of ~10 Tg C yr-1 by 2050 so clearly a massive, rapid mobilisation of resources would be necessary to enable such a scale of action.

The three scenarios projected to give a potential range of values are summarised in Table S3 and the results of the scenarios are presented in Figure S3.

**Table S3.** Scenario modelling details for near-shore macroalgal aquaculture

|  |  |  |  |  |  |
| --- | --- | --- | --- | --- | --- |
| **Scenario** | **N flux driving productivity now (Tg N yr-1)** | **N flux driving productivity in 2050** | **Year in which maximum productivity reached** | **Seaweed C:N** | **Fraction of total N flux utilised** |
| Min | 50 | 25 | 2050 | 18 | 0.5 |
| Medium | 50 | 50 | 2050 | 18 | 0.5 |
| Max | 50 | 250 | 2050 | 18 | 0.5 |

**Figure S3.** Range of potential carbon storage from near-shore macroalgal aquaculture utilising excess riverine nutrients.

### **3.2 Protection of shelf carbon stocks and storage**

Shelf sediment carbon storage and stock loss:

There are large stocks of POC and PIC in shelf sediments. Scaling up an estimate for the NW European shelf from Legge et al. (2020), total global stocks are estimated at 7 to 29 Pg C as POC and 170 to 550 Pg C as PIC.

Estimates of annual present-day global carbon storage in shelf sediments lie between 0.2 (Bauer et al., 2013) and 1.6 (Krumins et al., 2013) Pg C yr-1. This corresponds to 28 to 140 Pg Carbon stored up to 2100. This is implicitly accounted for in the ocean fraction of anthropogenic emissions as it is taken up across the air-sea interface, so it is not considered as an ‘additional’ carbon benefit for CDR. However, the vulnerability of stock and storage due to anthropogenic influence is within scope.

Luisetti et al. (2019) estimated a loss of sediment organic carbon stock due to trawl fishing in UK waters of between 2.9 and 6.4% per decade. This is an upper estimate for the region and furthermore, UK waters are some of the most heavily trawled globally, so we use the lower-end estimate to extrapolate a potential global loss of 2 to 8 Pg C by 2100. But note that the impact of trawling overall is very poorly constrained and even the direction of the impact is not known for sure (Legge et al., 2020, and references therein; Smeaton et al., 2022)

How might we enhance sediment carbon storage? If trawling were to be decreased then carbon might accumulate, although there is debate in the literature as to whether much carbon is lost during trawling (e.g., Smeaton et al., 2022). Taking the deposition estimate by Bauer (low end) or Krumins et al. (high end) and assuming that 10% of the global shelf area might be designated as a no-trawl zone, then the accumulation of an additional 100% of organic carbon per year over that area yields 1.7–13.3 Pg C, as well as preventing loss of 0.2 to 0.8 Pg C stock. This is highly speculative and should be considered an upper-limit potential value, with a high risk of little or no true net benefit to carbon storage (but still covered by the ‘low regret’ philosophy of Gatusso et al., 2021).

### **Managing shelf carbon export to the deep ocean**

The continental shelf pump of carbon dioxide to deep waters is a fundamental natural carbon sink (e.g., Bauer et al., 2013). Here we consider speculative potential mechanisms to manage or control this flux through the management of inputs and cycling of nutrients within shelf seas. First, it is necessary to estimate the global export of carbon from the shelf to the deep ocean.

**Off-shelf DIC flux**

Bauer et al. (2013) estimate 0.5 to 0.7 Pg C/year globally which extrapolates to 41 to 58 Pg C to 2100. Model studies assessed by Legge et al. (2020) for NW European Shelf suggest 3.5 to 6 Tmol C yr-1. Scaling to the global shelf (and converting units) gives 1.3 to 2.1 Pg C year -> 100 to 173 Pg C to 2100

We do not believe that this flux is manageable in any way, or at least, the changes will be due to changes in the organic fraction/enhanced productivity, etc., and therefore counted elsewhere.

**Off-shelf DOC flux** is estimated by Barron and Duarte (2015) to be between 4 and 27 Pg C yr-1, which, assuming an arbitrary 30% of this DOC is not respired in the surface ocean within 100 years, gives 110–750 Pg C storage to 2100. This is larger than the estimated total ocean uptake of atmospheric CO2 so either i) a large proportion of this is terrestrial (likely) or ii) this flux of carbon is not captured in global climate models (also likely). It also appears that Barron and Duarte (2015) may have overestimated the cross-shelf exchange – the Huthnance (1995) values of 1 or more Sv per 1,000 km that they use are for transient features of shelf-edge exchange rather than a long-term average. Based on a recent analysis of NW European shelf (Spingys, 2017)) their shelf-edge exchange should arguably be revised down by a factor of 4. On this basis, we revise Barron and Duarte's potential storage down to 27 to 190 Pg C to 2100 (0.3 to 2.1 Pg C yr-1). Note these annual values equate to Barron and Duarte (2015) values (4 to 27 Pg C yr-1) divided by 4 and then multiplied by 0.3 to account for assumed 30% not respired.

Bauer et al. (2013) estimated the net export from the coastal ocean of DOC to be 0.15–0.35 Pg C yr-1 or 12–30 Pg C to 2100. This second estimate is based on budgeting carbon on the global shelf – this value is the residual. In the absence of better information, we take the lower estimate of Bauer et al., (2013) and (revised) upper estimate of Barron and Duarte (2015) and so assume 12 to 190 Pg C to 2100 (0.15 to 2.1 Pg C yr-1)

**Potentially manageable shelf carbon**

Chaichana et al. (2019) observe 30–50% differences in DOC inventory of the North Sea interannually. This may be due to circulation changes but is also potentially due to changing nutrient and runoff regimes changing the degradation rate and the bioavailability of DOM for respiration (Jaio et al., 2014). Any carbon not remineralised is not available for release back into the atmosphere. Therefore with a better understanding of the role of riverine nutrients in controlling the DOC flux and application of this knowledge we, for arguments’ sake, assume 10% of the variability in DOC concentration in marginal seas could exert a 4% change on the DOC flux. Assuming that the knowledge and capability to i) understand and modify and ii) monitor the additional carbon sequestration could be achieved by 2040, we model a scenario of 4% enhanced continental shelf organic carbon pump from 2040–2100, yielding a net additional carbon sequestration of 0.36 to 5.0 Pg C to 2100. Note that this management might well be achieved as an additional benefit of nutrient control with macroalgae in coastal seas and therefore be a co-benefit.

## **Text S4. Whole ocean solutions – mCDR using the global ocean**

Here we synthesise literature data to determine ranges of the maximum potential from each of the mCDR approaches considered below. These technological solutions require substantial spin-up time and so, unlike previous studies which have commonly applied maximum capacity from a given start date, e.g., 2030, we model a realistic growth rate up to maximum capacity in the latter half of the century (see Section 5).

### **Macroalgal geoengineering**

It is worth considering what might be achieved on a ‘war footing’ with an active sequestration scenario where additional C storage from global seaweed aquaculture is directed to carbon sequestration: The model of Lehahn et al. (2016) provided an estimate of the theoretical, potential global farmed seaweed biomass production as 1011 dry weight (DW) tonnes year-1, over a surface area of ~108 km2 (a little less than 30% of the total ocean surface!). As they demonstrate, a significant amount of ocean area would be dedicated solely to seaweed production at the expense of, e.g., shipping, wildlife, etc and this is clearly neither practicable nor desirable. A more realistic maximum theoretical yield is probably at least an order of magnitude less than this (based on arguments in Buschmann et al., 2017), although noting some areas considered unsuitable for energy and bioproducts production might be suitable for carbon sequestration. Assuming a carbon content of 25% dry weight (Duarte et al., 2017) this 10% of the Lehahn et al. (2016) maximum value gives a possibly more realistic kelp production of 2.5 x 109 tonnes C (2.5 Gt C yr-1). This is difficult or impossible to achieve without, e.g., open ocean kelp farms fed by deep water nutrients and thus represents large-scale geoengineering. Seaweed aquaculture at present levels produces approximately 3 million tonnes dry weight seaweed production (Duarte et al., 2017), so to reach this assumed upper limit represents a 1,000-fold increase in macroalgal production. The seaweeds produced would need to be securely stored away from the atmosphere by, e.g., burial in the deep ocean or charring for addition to soils as biochar.

de Ramon N’Yeurt et al. (2012) claim that utilisation of 9% of the ocean surface could sequester 53 Gt C per year whilst also producing biomethane for energy and other side-products. They achieve this by in-situ processing and nutrient recovery. Using this approach, they assume maximum yields (average of 1800 g C m-2 yr-1) could be achievable anywhere in the ocean. This would require massive engineering – floating cities farming kelp and nutrient pipes bringing nutrients up from depth, etc. We use this figure, downscaled by a factor of 10 as with Lehahn et al. (2016), to give the potential upper limit of productivity.

If seaweed is grown for other purposes (food, bioproducts, etc) then only by-product C sequestration can be counted (about 10% of net primary productivity according to Krause-Jensen and Duarte 2016). Therefore, a minimum scenario of utilising 50% of produced seaweed for other purposes yields a C storage of 55% of the total potential.

Table S4: Macroalgal geoengineering maximum storage rates:

|  |  |  |  |  |
| --- | --- | --- | --- | --- |
| **Scenario** | **Max production figure (source)**  | **Downscaling factor** | **Seaweed harvest used directly for C storage (%)** | **Max storage rate (Gt C yr-1)** |
| Low | 100 Gt DW year-1 (Lehahn et al., 2016) \* 25% Carbon | 0.1 | 50 | 1.375 |
| Medium | 100 Gt DW year-1 (Lehahn et al., 2016) \* 25% Carbon | 0.1 | 100 | 2.5 |
| High | 53 Gt C year-1 (de Ramon N’Yeurt et al., 2012) | 0.1 | 100 | 5.3 |

It is important to note that we have applied here a somewhat subjective limit of 10% of the stated potential values from the de Ramon N’Yeurt et al. (2012) and Lehahn et al. (2016) papers. If their quoted uptakes were achievable, large-scale macroalgal aquaculture coupled with some kind of carbon storage could on its own meet the negative emissions requirement under most future climate scenarios, notwithstanding any side effects of such large scale farming, or the huge logistical issues associated with such an undertaking. As seen in the discussion of growth rates below, the time to achieve such enormous levels of macroalgal aquaculture would also likely be far too long to reach such levels by 2100.

### **Other whole ocean solutions**

Also considered are ocean alkalinity enhancement, and fertilisation by iron or macronutrient addition or through artificial upwelling. The potential maximum values for a low, medium, or high scenario are given in Table S4.

**Table S5 Open ocean mCDR scenario calculation details**

|  |  |  |  |  |
| --- | --- | --- | --- | --- |
| **mCDR technique** | **Low scenario (Gt C yr-1)** | **Med scenario (Gt C yr-1)** | **High scenario (Gt C yr-1)** | **Notes** |
| Ocean alkalinity enhancement | 2 | 3.8 | 10 | Low: Ilyina et al. (2013) 2.4 Pmol yr-1 total alkalinity addition \* 0.83 moles CO2 uptake per unit alkalinity (Renforth and Henderson, 2017). Medium: Keller et al. (2014) total atmospheric CO2 uptake in simulation to 2100 averaged over their integration time (2020–2100). High: Max capacity from Renforth and Henderson (2017). Note in all studies the limit to this technique is related to the capacity of shipping to deliver alkaline mineral materials to the open ocean.  |
| Ocean iron fertilisation | 1.04 | 2.07 | 3.1 | 2.07 Gt C yr-1 from Keller et al. (2014) +/- 50% for low and high scenarios.  |
| Ocean macronutrient fertilisation | 0.75 | 1.5 | 2.25 | Harrison et al. (2017) 1.5 Pg C yr-1 +/- 50% |
| Artificial ocean upwelling | 1.6 | 3.2 | 4.8 | 3.17 Gt C yr-1 as per Keller et al. (2014) +/- 50% |

## **Text S5. Growth rate of open ocean CDR technologies.**

Given the high tech, high infrastructure needs and high-risk nature of open ocean CDRs we consider here the likely maximum growth rates of such technologies and therefore the achievable impacts of such technologies through time.

Examples of the highest known rates of industrial output growth are associated with military production in WWII and the mobilization of resources for that war has been used as an analogy for the action needed to mitigate climate change (e.g., Delina and Diesendorf, 2013). Mobilization of US industry enabled roughly a year on year doubling of aeroplane and tank production between 1939 and 1944, but towards the end of this period, capacity limits were being reached even before production was scaled down to avoid overproduction as the conflict started to reach a conclusion, suggesting that such a rate of production would not be sustainable over longer periods (Morgan, 1994).

A more recent example of rapid industrial output capacity increase is the growth of motor vehicle manufacture in China, which roughly doubled every three years between 2000 and 2009 (Lane et al., 2019)). Doubling every three years is an approximately year-on-year growth of 40%.

Given the risks associated with open ocean mCDRs and therefore the need for greater scrutiny and regulation (as well as the fact that the above examples are largely scaling up the production of a proven technology rather than developing new ones), we suggest that doubling carbon uptake capacity by NETs every 5 years (19% per year) is the fastest possible achievable growth rate and doubling every decade (8%) is probably more realistic. These are comparable to the rollout of nuclear power (11%) or Flue Gas Desulphurisation (15%) and the growth of coal and gas energy globally (5 to 10%) (Iyer et al., 2015).

Nuclear grew on average at about 11% per year after huge investments and some strong regulatory controls (Iyer et al., 2015). We, therefore, suggest mCDRs, with massive infrastructural challenges, the need for stringent regulation and accounting for false starts is unlikely to surpass 11% per year.

## **Text S6. Natural ocean carbon system responses to anthropogenic forcings**

In this section we consider a selection of natural Earth-system responses to increased CO2 and other anthropogenic forcings which could potentially enhance or confound attempts to mitigate atmospheric CO2 with CDR approaches. See main text for context.

### **Shelf alkalinity pump response**

Denitrification releases molecular nitrogen (N2) and alkalinity as products. As such, this constitutes a direct coupling between anthropogenic perturbations of the nitrogen cycle, i.e., fertilizer use, and carbon storage of the oceans. In order to shed light on the potential scale of that coupling, we here provide a “back-of-the-envelope”, likely upper bound estimate: Under the assumptions that *1:* 1 T biologically fixed CO2 (as wheat, for example) requires 0.02T nitrogen supplied as fertilizer (Pallière, 2004), that *2:* SSP5 assumes 50 GT CO2 cycling per year for bioenergy (after Peters, 2017, and Riahi et al., 2017), which corresponds to 1 Pg N, that *3:* approximately 40% of that fertilizer will leach into the coastal ocean (Nevison et al., 2016), and that *4:* all of that N (0.4 Pg N, 0.03 Pmol N) will be denitrified in the coastal zone, this would yield an additional ocean CO2 uptake of 0.3 Pg C yr-1 through alkalinity generation (after Thomas et al., 2009), i.e., 15% of the overall oceans CO2 uptake of assumed 2 Pg C yr-1 or in N-terms threefold higher denitrification in the oceans’ shelf regions (18 Tmol N yr-1 Seitzinger et al., 2006). Integrated from 2024 until 2100, i.e., over 74 years, this yields an additional CO2 uptake potential by the oceans of 0.3 Pg C \* 74 = 22.2 Pg C (81.4 Pg CO2, 1.85 Pmol C). While SSP5 is a high-energy scenario and while admittedly this estimate is subject to large uncertainty and error, it reveals the potentially massive scale of the issue if implemented globally. For comparison, the current effect of denitrification of anthropogenic nitrogen in the marine realm, estimated at 250 Tg N yr−1 (18 Tmol N yr−1; Seitzinger et al., 2006) can be estimated to 15 Tmol CO2 (or 0.6 mol CO2 m−2 yr−1, assuming a Revelle factor of 11, and an area for shelf seas of 25\*106 km², 7% of the global ocean area).

However, at least two further aspects have to be considered here, which might reveal the potential to (partially) outweigh the expected increase in ocean carbon storage. Firstly, the eutrophication-driven release of N2O and CH4 associated with anoxic sediments counteracts the enhanced CO2 storage effect. Secondly the overall greenhouse gas balance, i.e., the CO2 footprint of the fertilizer production itself, which can be estimated to approximately 4 t CO2-eqv per t N (4:1). This approximates the respective CO2 storage effect which in turn could only be realised if fertilizer was produced using regenerative energy sources.

**Affect of ocean acidification on calcifying organisms**

Increasing atmospheric CO2 leads to ocean acidification and a reduction in carbonate ions, increasing the thermodynamic cost of calcification for calcareous organisms. There is a potential negative feedback on increasing atmospheric CO2 as alkalinity removal from calcification reduces or stops (e.g., Hoffman and Schellenhuber, 2009; Zhang et al., 2016). Second, order feedbacks relating to mineral ballasting may partially offset this negative feedback but in any case such feedbacks are small relative to anthropogenic carbon release (Heinze, 2004; Hoffman and Schellnhuber, 2009) and the principal concern is the biological impact of ocean acidification changes on calcifying organisms.

Caldiera and Rau (2000) estimate that up to 1.5 moles of carbonate dissolution is required for a mole of CO2 permanently sequestered so we apply thisconversion factor to the calculations presented below to determine the net CO2 effect in Table 1 of the main manuscript.

**Coral reefs**

Coral reefs occupy 284,300 to 600,000 km2 of the World’s oceans (Spalding et al., 2001), and even under more optimistic scenarios for CO2 emissions are critically threatened (e.g., Silverman et al., 2009; Hoegh-Guldberg et al., 2017), although it is unclear to what extent or at what atmospheric CO2 concentration significant impacts on coral calcification will occur (e.g., Pandolfi et al., 2011). Silverman et al., (2009) predict that coral reefs will become net sinks for CO2 (i.e., net dissolution of calcium carbonate) at around 700 ppm of atmospheric CO2 but conclude that the impact on the global carbon cycle would be small (0.017 Pg C yr-1) which equates to 1.34 Pg C by 2100 (4.9 Pg CO2). This is our lower estimate.

Davis et al. (2021) predict a shift in coral ecosystems from net calcification to net dissolution mid-century (2054). Currently, global net calcification rates averaged over their study are of the order of 100 mmol (CaCO3) m2 d-1, which over 6 x 1011 m2 equates to 2.2 x 1013 moles CaCO3 yr-1, equating to 0.64 Pg CO2 yr-1. Assuming a linear decrease to 2054 and the same rate of change continuing to 2100 (i.e., linear increase in net dissolution) this yields a total difference in CO2 balance (compared to continued calcification at the current rate) of 62 Pg less CO­2 production. We take this as an upper estimate.

**Calcifying plankton**

Extrapolating from a global biomass of 444–505 Tg C (Bednarsek et al., 2012) and assuming a PIC:POC of 1 and a turnover of pteropod biomass annually, pteropod carbonate production is estimated to be 222–252 Tg C yr-1. Pteropods exhibit a 28% decrease in calcification when exposed to 2100 BAE scenario pH levels (Comeau et al., 2010). This will lead to 61 to 71 Tg C less calcification by 2100 or approximately. The reduced calcification of all calcifying plankton (pteropods, coccolithophores, foraminifera etc) is predicted in an earth system modelling study to be between 120 and 640 Tg C yr-1 (Gangstø et al., 2011). Taking 70–640 as the range of annual net decrease and extrapolating to 2100 this gives a total estimated reduction of CO2 production of between 2.7 and 25 Pg C (10 to 92 Pg CO2).

However, there are conflicting studies on the extent to which pteropods and other calcifying organisms will be affected by acidification and temperature, ranging from prediction of complete loss of stocks, to the adaptation and the maintenance of calcified shells by some species. Changes seem to be dependent on location and species. For instance, pteropod stocks are expected to actually increase or at least be maintained in the Antarctic because warming will result in a release of available nutrients from sea ice, which will allow pteropods to grow/maintain shells despite undersaturation (Bednarsek et al., 2012), although this is not backed up by more recent studies (e.g. Gardner et al., 2018).

**Non-coral shallow water carbonates**

Morse et al. (2006) investigate the potential for Mg-rich calcite (which is more readily soluble than aragonite or pure calcite) to dissolve. Depending on the dissolution model used they predict between 0 and 1016 moles of Mg-rich calcites will dissolve by 2100 under old IPCC scenario IS92a, which is comparable to RCP 4.5 rather than 2.6 – obviously the atmospheric CO2 is a key driver of dissolution so a lower pathway would mean a lower uptake capacity due to Mg-calcite dissolution. Nonetheless, we adopt the upper estimate as the potential maximum response.

### **Ocean carbon pump response**

Various biological and abiological mechanisms result in the export of carbon from the atmosphere, via the surface ocean, to the deep ocean (Buesseler et al., 2020; Nowicki et al., 2022), with a large stock of carbon stored out of contact from the atmosphere for hundreds of years as dissolved inorganic carbon and potentially up to tens of thousands of years as refractory organic carbon. Estimates of the strength of the ocean carbon pump range from 2.5 to ~10 Pg C yr-1 globally (Gruber and Sarmiento, 2002; Henson et al., 2011; Nowicki et al., 2022).

The relative importance of the various pump mechanisms, and their likely response to global change are poorly understood (e.g., Nowicki et al., 2022) and differences in response between different ocean bioregions may be important (e.g., subtropical gyres, where community shift to more picoplankton will have an unknown effect on net C export; e.g., Lomas et al., 2011).

Given the complexities of the ocean carbon pump and its response to a changing climate there are a wide range of estimates of the magnitude and even net direction of change in the carbon pump in future. The state-of-the-art in ocean modelling (Nowicki et al., 2022) predicts a reduction in storage of 65 Pg C (238 Pg CO2) in the deep ocean by the end of the century. However, overall the direction, let alone the magnitude of the response is yet unclear (Henson et al., 2022), with net changes in global carbon export strength of –41 to +1.8%. On this basis we adopt the range of 10 to –238 Pg CO2 as the possible range of responses..

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