# Carbon degradation in agricultural soils flooded with seawater after managed coastal realignment

Kamilla S. Sjøgaard[1], Alexander H. Treusch[1] and Thomas B. Valdemarsen[2]

[2]Department of Biology. [1]Nordic Center for Earth Evolution, University of Southern Denmark, Odense M, 5230, Denmark

5 *Correspondence to*: Kamilla S. Sjøgaard (kamillasjogaard@biology.sdu.dk)

**Abstract.** Permanent flooding of low-lying coastal areas is a growing threat due to climate change induced sea level rise. An increasingly common solution to protect coastal areas is their intentional flooding by 'managed coastal realignment'. However, the biogeochemical implications of flooding agricultural soils with seawater are still not well understood. We conducted a 1-year mesocosm experiment to investigate microbial carbon degradation processes in soils flooded with

10  seawater. Agricultural soils were sampled on the northern coast of the island Fyn (Denmark) at Gyldensteen Strand, an area that was subsequently flooded in a coastal realignment project. We found rapid carbon degradation almost immediately after experimental flooding and microbial sulfate reduction established quickly as an important mineralization pathway. Nevertheless, no free sulfide was observed as it precipitated as Fe-S compounds with Fe acting as a natural buffer, preventing toxic effects of free sulfide in soils flooded with seawater. The refractory nature of the terrestrial organic carbon,

15  in combination with the anoxic conditions created in the soil after flooding, caused significantly decreased organic carbon degradation after 6 months. During the experiment only 6-7 % of the initial soil organic carbon pools were degraded. On this basis we suggest that flooding of coastal soils through sea level rise or managed coastal realignment, will cause significant preservation of soil organic carbon and create an overall negative feedback on atmospheric carbon dioxide concentrations.

## 1 Introduction

20  Sea level rise driven by global climate change is expected to continue for centuries, impacting about 70 % of the global coastlines (Church et al., 2013). Rising sea level causes higher and more frequent storm surges, which lead to more incidences of floodwaters overtopping and breaking coastal defenses (FitzGerald et al., 2008). Reclaimed coastal areas with low elevation are especially vulnerable to flooding. A low cost strategy of coastal protection is 'managed coastal realignment', whereby old coastal defenses are deliberately breached, and new ones are constructed further inland (Cooper,

25  2003; French, 2008; Roman and Burdick, 2012). The flooded areas created by managed coastal realignment act as buffer zones, protecting populated areas or valuable assets against flooding. Projects where coastal soils are flooded with seawater, by managed costal realignment and similar techniques, occur at an increasing rate (Herbert et al., 2015; Pethick, 2002; Wolters et al., 2005).

Many studies have been performed on freshwater wetlands experiencing salinization and less on diked and drained agricultural soil systems (Ardon et al., 2016; Ardon et al., 2013; Herbert et al., 2015; Portnoy, 1999; Portnoy and Giblin, 1997), where flooding with seawater has dramatic consequences for soil biogeochemistry. Depending on soil porosity and moisture content, soil environments can have deep oxygen penetration (75-100 cm) (Dziejowski et al., 1997; MacDonald et al., 1993; Neira et al., 2015), since oxygen ($O_2$) can rapidly be supplied from the overlying atmosphere via diffusion. Therefore, surface soils are predominantly oxic environments where soil organic matter is degraded by a wide variety of microorganisms, fungi and fauna (Boer et al., 2005; Kalbitz et al., 2000). Aerobic degradation is catalysed by hydrolytic enzymes and reactive oxygen radicals that can break bonds in refractory organic compounds such as lignin and cellulose, and facilitate complete degradation of soil organic carbon (SOC) to $CO_2$ (Canfield, 1994). However, when soils are flooded, $O_2$ penetration is dramatically reduced, since $O_2$ solubility in water is low and $O_2$ diffusion in water is $10^4$ times slower than in air (Neira et al., 2015). $O_2$ will therefore be depleted by microbial and abiotic $O_2$ consuming processes in surface soils, which become anoxic below a few milimeters depth. In anoxic environments mutualistic consortia of microorganisms degrade organic macromolecules into smaller moieties by the excretion of exoenzymes and extracellular hydrolysis, which are then fermented into smaller organic molecules, mainly acetate (Valdemarsen and Kristensen, 2010). The fermentation products are taken up by microorganisms and terminally oxidized to carbon dioxide ($CO_2$) by the consumption of alternative electron acceptors (e.g. nitrate, Mn oxides, Fe oxides and sulfate) (Arnosti, 2011; Glud, 2008). Furthermore, soils flooded with seawater are enriched with sulfate that can be utilized as an alternative electron acceptor, and microbial sulfate reduction (SR) is expected to become a major mineralization pathway (Sutton-Grier et al., 2011; Weston et al., 2011).

While some studies have looked at SOC mineralization pathways in different types of soils introduced to saltwater (Ardon et al., 2016; Chambers et al., 2013; Herbert et al., 2015; Neubauer et al., 2013; Weston et al., 2006; Weston et al., 2011), a lot is still unknown about how the dynamics between initial SOC degradation to DOC and the terminal mineralization are affected by the introduction of saltwater (Herbert et al., 2015). Many soils subject to managed coastal realignment contain considerable amounts of SOC (Franzluebbers, 2010; Wolters et al., 2005). The degradation of SOC after flooding will depend on the rate of establishment of heterotrophic microbial communities and their ability to degrade SOC (Schmidt et al., 2011). Labile organic carbon may be easily degraded by marine microorganisms, while more complex organic carbon, and especially structurally complex organic compounds such as cellulose and lignin, may be virtually non-degradable in anoxic environments (Kim and Singh, 2000; Kristensen and Holmer, 2001). Flooding of coastal soils by sea level rise and coastal realignment may therefore cause significant preservation of stored SOC, implying an overall negative feedback on atmospheric $CO_2$ concentrations.

In this study the fate of SOC after flooding with seawater was investigated in soils collected at Gyldensteen Strand on the northern coast of Fyn, Denmark, an area that was designated to be flooded in a coastal realignment project. We were especially interested in following the temporal establishment of dominating microbial pathways and quantifying the rates and temporal trajectories of SOC degradation in newly flooded soils. We hypothesized that (1) total SOC degradation activity in soils after flooding depends on SOC content and lability, and that (2) most SOC at the time of flooding will, due

to it's terrestrial origin, be non-degradable and hence preserved under the anoxic conditions formed after the flooding. Therefore it is plausible that (3) the majority of SOC will be preserved in the flooded soils, creating a negative feedback on atmospheric $CO_2$ concentrations. To test these hypotheses we performed parallel mesocosm experiments with two different types of soils that were experimentally flooded with seawater. SOC-degradation and other biogeochemical developments in the mesocosms were traced with high temporal and spatial resolution for the next 12 months. The results showed how flooding with seawater impacts C-degradation and soil biogeochemistry and formed the basis for an initial evaluation of potential feedbacks of flooding on atmospheric $CO_2$ concentrations.

## 2 Materials and methods

### 2.1 Study site

This study was conducted in relation to the nature restoration project at Gyldensteen Strand funded by the Danish Aage V. Jensen Nature Foundation. The sampling site (55°34'26.4"N 10°08'17.0"E) was a shallow intertidal habitat until 1871 (size of ~600 ha), where it was diked and continuously drained to create new land for agriculture. The reclaimed area was for the following 140 years mainly used for production of different crops such as onions and grains (Stenak, 2005). As a part of the nature restoration project, selected sections of the dikes were removed in March 2014 and 211 ha of the area were permanently flooded with seawater and turned into a shallow marine lagoon.

### 2.2 Experimental design and Sampling

Sampling for the mesocosm experiment was performed in November 2013, half a year before the flooding of the site, at two different stations representing uncultivated (UC) and cultivated (C) soils (Fig. 1). Station UC was located in an area with low elevation, which never could be properly drained. Station UC was therefore abandoned for agriculture and became a reed swamp that accumulated plant material and litter. Station C, however, resembled the majority of the re-flooded area that was farmed since the land reclamation (fertilized, ploughed and used for monoculture, also illustrated in Fig. 1). From each station, 15 soil cores were sampled in 30 cm long, 8 cm internal diameter stainless steel core liners. The core liners were hammered 25 cm down into the soil, dug up with a spade and closed in both ends with rubber stoppers.

In the laboratory, the headspaces of individual soil cores were gently flooded with 22-26 salinity seawater collected from the shore face directly north of station UC (Fig. 1). Soil cores were then transferred to 70 L incubation tanks filled with seawater. During the whole experiment the flooded cores were maintained at 15 °C and kept in darkness. The water in the tanks was rigorously aerated through air diffuser stones and 10-20 L of the seawater in the tanks was exchanged with fresh seawater (also collected from the shore face) every 14 days. Thus soil cores were incubated under constant environmental conditions, while factors such as diurnal temperature variations, tidal exchange, benthic primary production and bioturbation were omitted by the experimental setup.

The flooded soil cores were incubated for 12 months. Flux experiments were conducted with 3 random soil cores from each station at various times (weekly in the first month, biweekly for the next 3 months and monthly hereafter). Core sectionings were performed on 3 randomly selected soil cores from each station at different times during the experiment (before the flooding, 1 week after and after 2, 4, 6 and 12 months).

5 **2.2.1 Flux experiments**

Fluxes of $O_2$, dissolved organic carbon (DOC) and $TCO_2$ (= $CO_3^{2-}$ + $HCO_3^-$ + $H_2CO_3$) between soil and overlying water were measured regularly as described above. Cores were equipped with stirring magnets, closed with rubber stoppers and placed around a central magnet rotating at 60 rpm and hereafter incubated for about 4 hours in darkness. $O_2$ was measured and water samples were taken in the headspace of the soil cores at the beginning and end of incubations. $O_2$ was measured with
10 an optical dissolved oxygen meter (YSI ProODO). DOC samples were stored at -20 °C until analysis using a Shimadzu TOC-5000 Total Organic Analyzer. Samples for $TCO_2$ analysis were kept in 3 mL gas-tight exetainers for a maximum of 1 week until analysis by flow injection (Hall and Aller, 1992).

**2.2.2 Core sectioning**

Core sectioning was performed by slicing each soil core into 6 depth intervals (0-1, 1-2, 3-5, 5-10, 10-15 and 15-20 cm).
15 Porewater was extracted from each depth interval by centrifugation and GF/C filtration in double centrifuge tubes (500 g, 10 min). The porewater was sampled for various parameters; 500 µL porewater were preserved with 30 µL saturated $HgCl_2$ for $TCO_2$, 250 µL porewater were preserved with 50 µL 1 M zinc acetate (ZnAc) for total dissolved sulfide ($TH_2S = H_2S + HS^-$ + $S_2^-$) analysis, 250 µL porewater were preserved with 100 µL 0.5 M HCl for $Fe^{2+}$ analysis and remaining porewater was stored at -20 °C until analysis for sulfate ($SO_4^{2-}$) and DOC. $TCO_2$ and DOC samples were stored and analyzed as described
20 above. $TH_2S$ samples were analyzed by the method of Cline (1969). $Fe^{2+}$ samples were analyzed by the Ferrozine method (Stookey, 1970). $SO_4^{2-}$ was analyzed by liquid ion chromatography on a Dionex ICS-2000 system.

Reactive iron, RFe, was extracted from soil subsamples from every depth interval with 0.5 M HCl for 30 min while shaking (Lovley and Phillips, 1987). After centrifugation (500 g, 10 min) the supernatant was transferred to sampling vials and stored at room temperature until analysis for reactive Fe(II) and Fe(III) [RFe(II) and RFe(III), respectively]. The
25 supernatant was analyzed for $Fe^{2+}$ and RFe by the ferrozine method (Stookey 1970) before and after reduction with hydroxylamine (Lovley and Phillips, 1987). RFe(II) was calculated directly, while RFe(III) was calculated from the difference between RFe and RFe(II). An estimate of total Fe content was obtained by boiling combusted soil subsamples in 1 M HCl for 1 hour at 120 °C. The supernatant was stored at room temperature until analysis by the ferrozine method.

Acid volatile sulfides (AVS) (Rickard and Morse, 2005) and chromium reducible sulfur (CRS) were determined on
30 soil subsamples preserved with 1 M ZnAc and stored at -20 °C until analysis. AVS and CRS were extracted by 2-step distillation as described in Fossing and Jørgensen (1998). Sulfide concentrations in the distillates were analyzed by the method described by Cline (1969).

Soil characteristics were also determined for every depth interval during every core sectioning. Soil density was determined gravimetrically and soil subsamples were dried (24 h, 105 °C) to determine water content and porosity. Soil organic matter content was measured as the weight loss of dry sediment after combustion (520 ˚C, 5 hours). SOC on selected soil samples (samples obtained after 1 week and 6 months) was also measured by elemental analysis on Carlo Erba CHN EA1108 Elemental Analyzer according to Kristensen and Andersen (1987).

### 2.2.3 Anoxic incubations (Jar experiments)

Depth distribution of microbial $TCO_2$ and DOC production and SR were estimated from anoxic soil incubations (Kristensen and Hansen, 1995; Quintana et al., 2013). The excess soil from core sectionings was pooled into 4 depth intervals (0-2, 2-5, 5-10 and 15-20 cm), thoroughly homogenized and tightly packed into 6-8 glass scintillation vials (20 mL). The vials were closed with screw caps, buried head-down in anoxic mud and incubated at 15 °C in darkness. 2 jars from each jar series were sacrificed every week for porewater extraction in the following 4 weeks. The screw caps were changed to a perforated lid containing a GF/C filter and the jars were centrifuged upside-down in a centrifuge tube (10 min at 500 g). The extracted porewater was sampled and analyzed for $TCO_2$, DOC and $SO_4^{2-}$ as described above.

### 2.3 Data analysis

Fluxes of $TCO_2$, DOC and $O_2$ were calculated from the concentration differences between start and end samples. Microbial rates in jar experiments (DOC and $TCO_2$ production and SR) were calculated for 0-2, 2-5, 5-10, 15-20 cm depth intervals by fitting the time dependent concentration changes by linear regressions with checks for obvious outliers. When the slopes were significant ($p < 0.05$), the volume specific reaction rates (nmol $cm^{-3}$ $d^{-1}$) in individual depth layers were calculated from the regression slopes corrected for sediment porosity. Microbial reaction rates, porewater and solid pools were depth integrated over 0-20 cm and converted to area specific units. Linear data interpolation was used to correct for missing data points, e.g. for the depth interval 10-15 cm where microbial rates were not measured. There was a significant linear correlation between organic matter content and SOC for both sampling stations [OC(%) = 0.0649 x LOI(%) + 0.0936, $r^2$=0.9824, n=20]. This correlation was used to convert organic matter into SOC for the time points where SOC was not directly measured. A one-way ANOVA was performed on area specific SOC pools at the different time points to test for significant changes in the SOC pools over time. Depth integrated SR rates were normalized to C-units since an almost 2:1 relationship between $TCO_2$ production and SR (Jørgensen, 2006) was observed throughout the experiment. For soil characteristics, fluxes, porewater and solid pools errors were calculated as standard errors of the mean (SEM). For depth-integrated values of microbial rates and solid pools errors were calculated as standard errors propagation (SEP) of standard deviation (SD) values following Eq. (1):

$$SEP = \sqrt{SD_{0-1\,cm}^2 + \cdots + SD_{15-20\,cm}^2} \qquad (1)$$

In a carbon budget estimating SOC degradation during the experiment, total degradation of SOC (mol m$^2$) was calculated as the sum of the time integrated $TCO_2$ efflux, time integrated DOC efflux and area specific $TCO_2$ and DOC in porewater by the end of the experiment. The percentage of the initial SOC pool degraded during the experiment was calculated from the estimated total degradation of SOC and mean bulk SOC pool. In a time specific carbon degradation budget, total degradation to $TCO_2$ was calculated as the sum of time integrated $TCO_2$ efflux and accumulated porewater $TCO_2$ at different time points after flooding (1 week and 2, 4, 6 and 12 months). Based on the jar experiments, total anaerobic $TCO_2$ production and $TCO_2$ production by SR (according to a 2:1 relationship between TCO2 production and SR) was calculated by time integration at different time points after flooding (1 week and 2, 4, 6 and 12 months). Relative contributions of SR to anaerobic degradation to $TCO_2$ were estimated from $TCO_2$ production and $TCO_2$ production by SR measured in jar experiments.

# 3 Results

## 3.1 Soil characteristics

The two sampled stations had very different soil appearance, as a result of different use after the land reclamation (i.e. no cultivation and cultivation). Station UC was overgrown with mosses and grasses, and a dense layer of roots and litter characterized the upper 5 cm of the soil, while the deeper parts of the soil (>10 cm depth) consisted of clay. At station C only relatively small amounts of grass and root material were evident in the upper 5 cm. Some of the vegetation was still alive 2 months after the flooding, as indicated by long green grass leaves seeking light, but it slowly died out thereafter. The soil at both stations contained partially degraded shell material from gastropods and bivalves remaining from when the area was a marine lagoon before 1871.

There was very little variation in soil characteristics between successive core sectionings, so results were averaged for the whole experiment (Table 1). The water content at station UC decreased with depth from 83 % at the top to 35 % in the bottom, while water content only decreased from 32 % to 20 % at station C. The same depth trend was observed for porosity. The high water content and porosity at station UC was caused by high amounts of plant material (e.g. roots), while the soil at station C was sandy, homogenous and poor in organic debris.

Soil organic content varied greatly with depth at station UC, and the topsoil was enriched with SOC (16 %) compared to the bottom (1 %) (Table 1). SOC varied between 0.8 and 1.4 % at station C with no depth variation. A one-way ANOVA showed no significant difference between the SOC contents at the different time points at either station UC or C (df = 17, F = 1.9, p = 1.16 for both stations).

## 3.2 $CO_2$ and DOC efflux, and $O_2$ consumption

$TCO_2$ effluxes in UC soil were highest in the beginning of the experiment with a maximum of 239±30 mmol m$^{-2}$ d$^{-1}$ measured on day 13 (Fig. 2a). Subsequently it decreased to about 130 mmol m$^{-2}$ d$^{-1}$ 31-199 days after flooding and stabilized

around 67 mmol $m^{-2}$ $d^{-1}$ from day 220 to the end. The $TCO_2$ effluxes in C soil were relatively constant around an average of 29 mmol $m^{-2}$ $d^{-1}$.

High DOC efflux was evident 1 day after flooding at station UC (108±3 mmol $m^{-2}$ $d^{-1}$) (Fig. 2b), while it decreased to around 60 mmol $m^{-2}$ $d^{-1}$ 6-20 days after flooding and to 17 mmol $m^{-2}$ $d^{-1}$ after approximately 2 months to the end. DOC effluxes at station C showed a similar pattern, averaging 25 mmol $m^{-2}$ $d^{-1}$ in the first 2 months after flooding, and decreasing to an average of 5 mmol $m^{-2}$ $d^{-1}$ for the remaining experiment.

$O_2$ consumption decreased almost linearly during the 1-year experiment on both stations (Fig. 2c). At station UC initial $O_2$ consumption was 57±3 mmol $m^{-2}$ $d^{-1}$, 1-45 days after flooding, and then it steadily decreased to 19±3 mmol $m^{-2}$ $d^{-1}$ by the end. At station C there was a less pronounced temporally decreasing trend. $O_2$ consumption was highest initially with about 26 mmol $m^{-2}$ $d^{-1}$ at day 1-13 and then decreased to 9±0.6 mmol $m^{-2}$ $d^{-1}$ by the end.

**3.3 Porewater chemistry**

Porewater DOC was high 1 week after flooding at both stations (on average 10.4 and 3.8 mM at stations UC and C, respectively; Fig. 3a). Over the experiment porewater DOC decreased slightly in UC soil, while it increased slightly in C soil.

Porewater $TCO_2$ concentrations in UC soil were in the range of 5-13 mM between 1 week and 2 months after flooding, and profiles showed a slightly increasing pattern with depth (Fig. 3b). Afterwards an unexpected drop in $TCO_2$ concentrations, especially in the deep soil (>2 cm depth), was observed. This was likely an experimental artifact, however, caused by extremely high $Fe^{2+}$ concentrations >2 mM in the porewater. During sample storage the $Fe^{2+}$ got oxidized to Fe-oxyhydroxides and formed an orange-brown precipitate at the bottom of the sample containers, probably leading to sample-acidification and $TCO_2$ degassing (Moses et al. 1987; Hedin 2006). Porewater $TCO_2$ concentrations in UC soil after 4 months were affected by this artifact. In C soil, porewater $Fe^{2+}$ did not accumulate at the same rate as in UC soil and only exceeded 2 mM in the 10-20 cm depth layer after 6 months. Here porewater $TCO_2$ accumulated gradually over time as expected (Fig. 3b). Rapid $TCO_2$ accumulation occurred in the first 2 months, where $TCO_2$ increased from 3-5 mM to 11 mM below 3 cm depth. After 2 months to the end, $TCO_2$ increased further in the 2-10 cm depth interval, while a decrease occurred below 10 cm depth, which was probably related to $Fe^{2+}$ exceeding 2 mM.

High concentrations of $SO_4^{2-}$ were introduced to the soil when flooded with seawater. Yet the initial water infiltration and diffusion was the only transport mechanism for dissolved $SO_4^{2-}$ in the mesocosm setup and the experimental period was evidently not sufficiently long to achieve equilibrium in $SO_4^{2-}$ in porewater concentrations down to 20 cm depth. As a result, porewater $SO_4^{2-}$ decreased steeply with depth at both stations (Fig. 3c). By the end of the experiment in UC soil, $SO_4^{2-}$ decreased from ~17 mM at the surface to zero below 10 cm depth. In C soil $SO_4^{2-}$ decreased linearly from ~17 mM at the surface to 0-2 mM at the bottom.

After 7 days of flooding the $Fe^{2+}$ depth distribution in porewater was constant with depth, with on average 0.02 and 0.2 mM at station UC and C, respectively (Fig. 3d). Afterwards a progressive increase in porewater $Fe^{2+}$ was observed at

both stations. At station UC $Fe^{2+}$ increased to up to 1.3±0.6 mM at 5-15 cm depth after 2 months and stabilized after 6 months, where $Fe^{2+}$ exceeded 4 mM below 5 cm depth. The same trend was observed at station C, where $Fe^{2+}$ accumulated to up to 3.7 mM at 15-20 cm depth after 12 months.

### 3.4 Anaerobic net DOC production in jar experiments

Net DOC production after 1 week of flooding was high in the surface 0-2 cm at station UC (2666±695 nmol $cm^{-3}$ $d^{-1}$; Fig. 4a) and decreased exponentially with depth to 203±23 nmol $cm^{-3}$ $d^{-1}$ at 15-20 cm depth. A gradually decreasing net DOC production was observed in all depth layers over the experiment, and by the end significant net DOC production (121-172 nmol $cm^{-3}$ $d^{-1}$) was only detected in the upper 0-5 cm. A similar pattern in net DOC production was observed at station C, although rates were much lower than at station UC. After 1 week of flooding, net DOC production at station C was 1155±158 nmol $cm^{-3}$ $d^{-1}$ in the upper 0-2 cm of the soil but only 66-83 nmol $cm^{-3}$ $d^{-1}$ below. After 4 months it had decreased to 135 nmol $cm^{-3}$ $d^{-1}$ in the top 0-2 cm and no net DOC production was detected below 5 cm depth. Very low rates (21-25 nmol $cm^{-3}$ $d^{-1}$) were detected in the top 0-5 cm by the end.

Depth integrated net DOC production at station UC was initially 118-133 mmol $m^{-2}$ $d^{-1}$ in the first 2 months after flooding and then gradually declined to 8 mmol $m^{-2}$ $d^{-1}$ after 12 months (Fig. 5). Initial depth integrated net DOC production at station C was 4-fold lower than at station UC. Net DOC production in C soil decreased by 75 % in the first 2 months after flooding and almost no net DOC production occurred after 6 months.

### 3.5 Anaerobic TCO$_2$ production in jar experiments

Initial depth trends in $TCO_2$ production were generally similar to those observed for DOC, but temporal trends were markedly different (Fig. 4b). At station UC, $TCO_2$ production was initially almost 1000 nmol $cm^{-3}$ $d^{-1}$ in the top 0-2 cm and decreased to 380 nmol $cm^{-3}$ $d^{-1}$ at 15-20 cm depth. After 2 months, $TCO_2$ production had increased in the surface 0-2 cm to 6250 nmol $cm^{-3}$ $d^{-1}$, while rates below 10 cm depth remained relatively low. After 4 months, $TCO_2$ production decreased to about 2500 nmol $cm^{-3}$ $d^{-1}$ in the top 0-2 cm, while it was not possible to determine $TCO_2$ production rates directly for soil deeper than 5 cm due to the problem with extremely high porewater $Fe^{2+}$ described above. As seen below, porewater $SO_4^{2-}$ concentrations were not affected by the high porewater $Fe^{2+}$ concentrations. For the affected data points $TCO_2$ production was calculated as rate of SR x 2, assuming that SR was the dominating $CO_2$ producing process in the anoxic soil (Jørgensen, 2006). The calculations showed that $TCO_2$ production had decreased further after 6 and 12 months in the top 5 cm (600-1000 nmol $cm^{-3}$ $d^{-1}$) and was quite stable below (0-85 nmol $cm^{-3}$ $d^{-1}$). $TCO_2$ production rates were generally much lower in C soil, while relative trends for $TCO_2$ production and their development over time were quite similar between stations. Maximum $TCO_2$ production rates occurred at 0-2 cm depth, where $TCO_2$ production varied from 400 to 780 nmol $cm^{-3}$ $d^{-1}$ between 1 week and 2 months and then gradually decreased to 110 nmol $cm^{-3}$ $d^{-1}$ by the end. Similar trends were observed in the deeper soil, where $TCO_2$ production decreased from 180-310 nmol $cm^{-3}$ $d^{-1}$ after 7 days to 7-53 nmol $cm^{-3}$ $d^{-1}$ after 12 months.

Area specific $TCO_2$ production at station UC was initially 115-200 mmol $m^{-2}$ $d^{-1}$ in the first 2 months, and decreased to 40 mmol $m^{-2}$ $d^{-1}$ after 6 months (Fig. 5). At station C area specific $TCO_2$ production was relatively stable around 44 mmol $m^{-2}$ $d^{-1}$ for the first 4 months and decreased to 21 and 10 mmol $m^{-2}$ $d^{-1}$ after 6 and 12 months, respectively.

### 3.6 SR in jar experiments

Significant SR was measured in the top 0-5 cm (470 nmol $cm^{-3}$ $d^{-1}$) in UC soil 1 week after flooding, while no SR was detected below (Fig. 4c). After 2 months, high SR was only measured in the top 0-2 cm (3128±190 nmol $cm^{-3}$ $d^{-1}$). After 4 months SR was still highest in the topsoil (1217±147 nmol $cm^{-3}$ $d^{-1}$), while significant SR was detected down to 10 cm depth. From 4 months to the end, SR gradually decreased at all depths to 338±147 and 43±6 nmol $cm^{-3}$ $d^{-1}$ at 0-2 and 5-10 cm depth, respectively. Since $SO_4^{2-}$ did not reach the bottom (15-20 cm) during the experiment at station UC, no SR occurred here. In C soil SR occurred at considerably lower rates than in UC soil. After 1 week SR was 177±25 nmol $cm^{-3}$ $d^{-1}$ at 0-2 cm depth and decreased exponentially with depth to zero at 15-20 cm depth. By month 2 and 4, SR occurred at all depths (20-159 nmol $cm^{-3}$ $d^{-1}$). Afterwards SR decreased in the upper 15 cm while no SR was detected in the 15-20 cm depth interval.

Depth integrated SR at station UC increased from 24 to 63 mmol $m^{-2}$ $d^{-1}$ between week 1 and month 2, corresponding to 48 and 126 mmol $m^{-2}$ $d^{-1}$ carbon mineralization, respectively (Fig. 5). SR had decreased to 27.7 mmol $m^{-2}$ $d^{-1}$ after 12 months. SR increased during the first 4 months in C soil (6 to 12 mmol $m^{-2}$ $d^{-1}$) and then decreased to 4 mmol $m^{-2}$ $d^{-1}$ after 12 months.

### 3.7 Solid pools of Fe and S

Before flooding, RFe(II) in UC soil increased with depth from 4 μmol $cm^{-3}$ at 0-1 cm depth to 13 μmol $cm^{-3}$ at 15-20 cm depth, while a corresponding increase in RFe(III) occurred from 19 to 44 μmol $cm^{-3}$ (Fig. 6). The RFe pools at station C were relatively constant with depth, on average 2.5 and 23 μmol $cm^{-3}$ for RFe(II) and RFe(III), respectively. Twelve months after flooding, RFe(II) in UC soil had increased to 34-59 μmol $cm^{-3}$, while RFe(III) had accumulated to 134.5±85 μmol $cm^{-3}$ in the top and decreased to an average of 4 μmol $cm^{-3}$ below. A similar trend was obtained in C soil with RFe(III) accumulating to 51.9±1.4 μmol $cm^{-3}$ on the surface. In UC and C soil, total RFe initially consisted of 78 and 92 % Fe(III), respectively, while it was reduced to 19 and 10 % by the end. Clearly, RFe(III) became reduced to RFe(II) during the experiment due to the anoxic conditions created by flooding.

The RFe content was quite heterogeneous at the study sites and there were large variations between soil cores. Based on all the depth profiles obtained over the experiment, average total Fe content in UC and C soil was 19.3±2.8 mol $m^{-2}$ and 26.7±1.8 mol $m^{-2}$, respectively.

Although jar experiments suggested high SR in both soil types, dissolved sulfide ($TH_2S$) was never detected in the porewater. Instead, a large fraction of the sulfide produced during SR accumulated as AVS and CRS in both soil types (Fig. 7). One week after flooding, AVS and CRS in UC soil were low (0.2-2.7 μmol $cm^{-3}$), except at 2-5 cm depth where AVS content was slightly elevated. 12 months after flooding, AVS and CRS had increased to 25±10 and 41±11 μmol $cm^{-3}$ at 2-5

cm depth, respectively, while no accumulation occurred below 10 cm depth. A similar pattern was observed in C soil, where AVS and CRS were initially constant with depth averaging 0.1 and 3.5 $\mu mol\ cm^{-3}$, respectively, and accumulated to 6.4±1 and 8.4±0.7 $\mu mol\ cm^{-3}$ after 12 months of flooding, respectively. Over the whole experiment total sulfide accumulated as AVS and CRS gradually increased, from 0.5 $mol\ m^{-2}$ before flooding to 4.7 $mol\ m^{-2}$ after 12 months in UC soil, and from 0.63 to 2 $mol\ m^{-2}$ in C soil.

## 3.8 Budgets for SOC degradation

Area specific SOC pools were 710.9±54 and 232.5±22 $mol\ m^{-2}$ (n = 18) in UC and C soil, respectively (Table 2). Total SOC degradation estimated as the sum of $TCO_2$ and DOC effluxes, and porewater accumulation over the 1-year experiment was 49.6 and 14.8 $mol\ m^{-2}$ at station UC and C, respectively, corresponding to 7 and 6 % of the SOC pools.

Total SOC mineralization to $TCO_2$ was estimated as the sum of $TCO_2$ efflux and porewater accumulation during the whole experiment (Table 3), which was 40.0 and 12.0 $mol\ m^{-2}$ at station UC and C respectively. The importance of anaerobic SOC degradation for total $TCO_2$ mineralization could be calculated from jar experiments, and a total of 32.6 and 10.8 $mol\ m^{-2}$ SOC was converted to $TCO_2$ anaerobically, corresponding to 82 and 90 % of flux-based total $TCO_2$ production at station UC and C, respectively. The SR measured in jar experiments corresponded to 25.3 and 4.3 $mol\ m^{-2}$ $CO_2$ production at station UC and C during the experiment. Thus 63 and 36 % of the flux-based total $TCO_2$ production was driven by SR in UC and C soil, respectively, starting at 30-40 % after 1 week and gradually increasing up to 100 % by the end of the experiment. This means that the remaining 19 and 54 % of the flux-based total $TCO_2$ production was produced by other anaerobic processes than SR in UC and C soil, respectively (e.g. nitrate or Fe reduction).

## 4 Discussion

### 4.1 Temporal trends in SOC degradation

The UC and C soil had very different organic content. UC soil had not been used for agriculture and organic matter consisting of dead and alive plant matter had accumulated in the topsoil (Table 1), while lower organic matter content was evident in C soil due to lower plant cover and regular mechanical soil reworking during agricultural cultivation (Benbi et al., 2015; Six et al., 1998). Consequently, the bulk SOC pool was 3 times higher in UC soil than in C soil. The source of soil organic matter at both stations was terrestrial plants such as grasses and herbs rich in cellulose and lignified tissues (Arndt et al., 2013; Sullivan, 1955). Such organic matter is refractory towards degradation in anaerobic marine sediments (Kristensen, 1990, 1994) compared to structurally simple phytoplankton, microphytobenthos and macroalgae, which are common organic carbon sources in coastal marine sediments (Dubois et al., 2012; Fry et al., 1977). It was therefore uncertain to which extent the SOC at Gyldensteen Strand could serve as substrate for developing microbial communities after the flooding with

seawater. Nevertheless, we observed high heterotrophic activity (e.g. $O_2$ uptake and $TCO_2$ production) right after the flooding, indicating that at least part of the SOC in both soil types was readily available for microbial degradation.

Cleavage of particulate organic carbon to DOC by extracellular enzymes is the primary degradation step in waterlogged anoxic soils and sediments (Arnosti, 2011; Weiss et al., 1991). The produced DOC is hereafter converted into short chain fatty acids and acetate, by microbially mediated fermentation and hydrolysis, which then are terminally oxidized to $CO_2$ by e.g. SR (Canfield et al., 2005; Valdemarsen and Kristensen, 2010). DOC production can therefore generally be considered the rate-limiting step for organic carbon degradation. However, a small proportion of produced DOC is recalcitrant and may accumulate in soil pore water over time in an experimental setup without advective porewater transport. In this experiment we observed high DOC concentrations in porewater and highest DOC production in jar experiments already 7 days after flooding with seawater (Fig. 3a & 5). Part of this DOC may have leached to the porewater as a result of flooding (Kalbitz et al., 2000), while the rest was produced by microbial degradation of particulate SOC (Kim and Singh, 2000). Microbial degradation of soil organic matter to DOC was initiated immediately after flooding irrespective of the shift to anoxic conditions. Differences in DOC production rates indicated that the availability of degradable SOC was clearly highest in UC soil compared to C soil following the overall difference in total SOC content. However, total DOC production ceased rapidly in both soil types and was close to zero after 1 year. Ardon et al. (2016) also observed reduced DOC concentration and export due to introduction of saltwater. It therefore appears that only a minor portion of SOC (6-7 %; Table 2) is available for microbial degradation under the present conditions (flooded with seawater and anoxic conditions). The low degradability of SOC after flooding probably reflects limitations of the anaerobic microbial communities to degrade complex organic matter of terrestrial origin (Fors et al., 2008; Yucel et al., 2013).

Heterotrophic DOC oxidizing microbes were also active immediately after flooding as shown by initial $TCO_2$ effluxes and high $TCO_2$ production in the jar experiments 7 days after flooding (Fig. 2a & 5). Rapid microbial $CO_2$ production has previously been observed in experiments with experimentally flooded soils (Chambers et al., 2011; Neubauer et al., 2013; Weston et al., 2011). In both soil types, $TCO_2$ production in the surface soil increased over the first 2 months, peaked, and then decreased gradually towards the end. These temporal dynamics were out of phase with DOC availability, indicating that terminally oxidizing microbes may adapt slower to flooded conditions than hydrolyzing microbes. Similar cases of initial substrate hydrolysis outpacing fermentation and SR has been observed before (Arnosti et al., 1994), maybe due to lag response in the microbial community (Bruchert and Arnosti, 2003). Nevertheless, the majority (~80 %; Table 2) of produced DOC over the whole experiment was oxidized completely to $TCO_2$, while the rest effluxed to the overlying water (~19 %) or accumulated in porewater (~1 %).

**4.2 SOC degradation pathways**

$SO_4^{2-}$ was an important electron acceptor in both soils and SR accounted for 63 and 36 % of the total $TCO_2$ production during the experiment in UC and C soil, respectively (Table 3). One week after flooding, active SR corresponding to 30-40 % of anaerobic $TCO_2$ production was detected in the jar experiment. The relative importance of SR increased gradually over

the experiment and by the end accounted for up to 100 % of the anaerobic $TCO_2$ production in both soil types. This is in accordance with Weston et al. (2006) who measured SR in freshwater marsh soil exposed to saltwater in anoxic flow through reactors, and found that the relative importance of SR for total $TCO_2$ production increased from 18 % initially to >95 % after 4 weeks. The delay in SR probably reflects a lag phase for the community of $SO_4^{2-}$ reducing microbes to respond to elevated $SO_4^{2-}$ levels. The delay in SR could also reflect initial competition with other $TCO_2$ producing pathways (e.g. $NO_3^-$ and Fe reduction) in the time right after flooding when $NO_3^-$ and oxidized Fe might have been abundant. However, as the soil became reduced due to increased SOC degradation activity and limited $O_2$ supply, electron acceptors other than $SO_4^{2-}$ were rapidly depleted and SR became the dominant respiration pathway.

By combining results from flux and jar experiments it was possible to confine the relative importance of different microbial respiration pathways in flooded soils. The difference between $TCO_2$ effluxes (aerobic + anaerobic processes) and $TCO_2$ production in jar experiments (anaerobic processes) suggested that aerobic respiration only played a minor role in the flooded soils (18 and 10 % in UC and C soil, respectively). On the other hand, SR was quantitatively a very important pathway, constituting 63 and 36 % of total C-mineralization to $TCO_2$ in UC and C soil, respectively. Hence 19 (UC) to 54 % (C) of $TCO_2$ production occurred by respiration processes not directly accounted for. Weston et al. (2006) found that Fe reduction was responsible for about 60 % of $CO_2$ production in the first 4 days after saltwater intrusion in coastal soils. When considering the high initial concentrations and the rapid decrease in soil RFe(III) in our experiment (Fig. 6), respiratory Fe-reduction was probably an important respiration process initially. However, based on this experiment it was not possible to distinguish between biological and chemical Fe-reduction.

**4.3 Will newly flooded coastal habitats be hotspots for SOC burial?**

In this study we observed that only 6-7 % of the total SOC pool in coastal soils was degraded by microbial processes in the first year after flooding with seawater. The low final SOC degradation rates (especially the very low DOC production) suggested that remaining SOC would be permanently buried due to the limited ability of anaerobic microbial communities to degrade complex organic matter of terrestrial origin (Burdige, 2007; Canfield, 1994; Hedges and Keil, 1995). Neubauer et al. (2013) similarly found long-term reduction of degradation rates and lability of SOC pools in a tidal freshwater marsh experiencing saltwater intrusion, which also support preservation of the SOC pool. Considering that terrestrial non-flooded vegetated soils generally have $CO_2$ effluxes in the order of 0.1 to >1 mol $m^{-2}$ $d^{-1}$ (Chirinda et al., 2014; Fang and Moncrieff, 2001; Hursh et al., 2017; Rustad et al., 2001), which is much higher than measured in the flooded soils in this study (Table 4), it appears that flooding of coastal soils with seawater, due to either sea level rise or mitigation techniques such as coastal realignment, will cause reduced $CO_2$ efflux and a negative feedback on atmospheric $CO_2$ concentrations. This negative feedback may constitute a marked C-sink. For instance, a detailed investigation of the soil characteristics in the topsoils (down to 20 cm) at Gyldensteen Strand suggests $48\pm6\cdot10^3$ kg SOC $ha^{-1}$ (average $\pm$ SEM, n = 30, T. Valdemarsen, unpublished results). Hence, when assuming that about 10 % of the soil organic matter will be degraded after flooding, the

nature restoration project at Gyldensteen Strand (211 ha) constitutes an immediate C-sink of about $9 \cdot 10^6$ kg SOC (or about 0.02 % of the total annual $CO_2$ emission of Denmark).

## 4.4 Efficient Fe-driven sulfide buffering in flooded soils

Accumulation of free $H_2S$ is often seen in metabolically active organic enriched marine sediments, where it has toxic effects on benthic fauna (Hargrave et al., 2008; Valdemarsen et al., 2010). It was therefore a concern if free $H_2S$ would accumulate in the soils from Gyldensteen after flooding, since this could hamper the succession of benthic fauna as well as overall ecological developments. However, despite the extremely high initial SR rates in the flooded soils, comparable to SR measured beneath fish farms (Bannister et al., 2014; Holmer et al., 2003) no accumulation of free $H_2S$ occurred in any of the soil types. Dent (1986); Portnoy and Giblin (1997); Weston et al. (2011) also observed a similar lack of $H_2S$ accumulation in soils introduced to saltwater, suggesting that newly flooded soils have a high capacity to buffer $H_2S$. Budget considerations suggest that most of the produced $H_2S$ was immediately re-oxidized, e.g. with $O_2$ in the surface soils, while a significant proportion (37 and 93 % in UC and C soil, respectively) precipitated as different Fe-S compounds, for instance FeS and $Fe_3S_4$ in AVS and $FeS_2$ and $S^0$ in CRS (Reddy and DeLaune, 2008; Rickard and Morse, 2005; Valdemarsen et al., 2010). The depth profiles of solid Fe and S showed that sulfide precipitation occurred at the same depths where active SR was measured, i.e. in the upper 10 cm in UC soil and down to 20 cm depth in C soil. The decreasing microbial activity and increasing Fe(II) over time will create a long term sulfide buffering capacity in the soil (Schoepfer et al., 2014).

## 5 Conclusions

In this study a rapid stimulation of heterotrophic microbial degradation of SOC was observed in two different soils (uncultivated or cultivated) following flooding with seawater. Degradation rates peaked in the first 2 months after flooding, and hereafter gradually declined to low levels after 1 year. Microbial SR was rapidly established in both soil types and was the dominating respiration pathway. Nevertheless, despite extremely high SR rates, $H_2S$ did not accumulate in the soils as it was re-oxidized with $O_2$ at the soil-water interphase or precipitated with Fe to form AVS and CRS. All three hypotheses stated initially were confirmed. Total SOC degradation activity in the tested soils clearly did depend on SOC content (hypothesis 1) and was 3-fold higher in organic rich uncultivated soil compared to the organic poor cultivated soil. However, only a small proportion of SOC (6-7 %) was degraded in the first year after flooding, and when considering the low final SOC degradation rates, it appears that most SOC is non-degradable under anoxic marine conditions and will be preserved after flooding (hypothesis 2). Hence this study suggests that in soils flooded with seawater the majority of SOC will be permanently preserved in comparison to non-flooded soils, therefore creating an overall negative feedback on atmospheric $CO_2$ concentrations (hypothesis 3).

*Acknowledgements.* We thank technician Birthe Christiansen for help with chemical analyses. Further we thank Erik Kristensen and Marianne Holmer for valuable discussions and for initiating research at Gyldensteen Strand. This work was supported by a grant from the Danish National Research Foundation (grant number: DNRF53) and the Aage V. Jensen Nature Foundation.

5  **References**

Ardon, M., Helton, A. M., and Bernhardt, E. S.: Drought and saltwater incursion synergistically reduce dissolved organic carbon export from coastal freshwater wetlands, Biogeochemistry, 127, 411-426, 2016.

Ardon, M., Morse, J. L., Colman, B. P., and Bernhardt, E. S.: Drought-induced saltwater incursion leads to increased wetland nitrogen export, Glob Chang Biol, 19, 2976-2985, 2013.

10  Arndt, S., Jorgensen, B. B., LaRowe, D. E., Middelburg, J. J., Pancost, R. D., and Regnier, P.: Quantifying the degradation of organic matter in marine sediments: A review and synthesis, Earth-Science Reviews, 123, 53-86, 2013.

Arnosti, C.: Microbial extracellular enzymes and the marine carbon cycle, Ann Rev Mar Sci, 3, 401-425, 2011.

Arnosti, C., Repeta, D. J., and Blough, N. V.: Rapid bacterial degradation of polysaccharides in anoxic marine systems, Geochimica et cosmochimica acta, 58, 2639-2652, 1994.

15  Bannister, R. J., Valdemarsen, T., Hansen, P. K., Holmer, M., and Ervik, A.: Changes in benthic sediment conditions under an Atlantic salmon farm at a deep, well-flushed coastal site, Aquaculture Environment Interactions, 5, 29-47, 2014.

Benbi, D. K., Brar, K., Toor, A. S., and Singh, P.: Total and labile pools of soil organic carbon in cultivated and undisturbed soils in northern India, Geoderma, 237, 149-158, 2015.

Boer, W., Folman, L. B., Summerbell, R. C., and Boddy, L.: Living in a fungal world: impact of fungi on soil bacterial niche development,
20  FEMS Microbiol Rev, 29, 795-811, 2005.

Bruchert, V. and Arnosti, C.: Anaerobic carbon transformation: experimental studies with flow-through cells, Mar Chem, 80, 171-183, 2003.

Burdige, D. J.: Preservation of organic matter in marine sediments: controls, mechanisms, and an imbalance in sediment organic carbon budgets?, Chem Rev, 107, 467-485, 2007.

25  Canfield, D. E.: Factors influencing organic carbon preservation in marine sediments, Chem Geol, 114, 315-329, 1994.

Canfield, D. E., Kristensen, E., and Thamdrup, B.: Aquatic geomicrobiology, Advances in marine biology, 48, 1-599, 2005.

Chambers, L. G., Osborne, T. Z., and Reddy, K. R.: Effect of salinity-altering pulsing events on soil organic carbon loss along an intertidal wetland gradient: a laboratory experiment, Biogeochemistry, 115, 363-383, 2013.

Chambers, L. G., Reddy, K. R., and Osborne, T. Z.: Short-Term Response of Carbon Cycling to Salinity Pulses in a Freshwater Wetland,
30  Soil Science Society of America Journal, 75, 2000-2007, 2011.

Chirinda, N., Elsgaard, L., Thomsen, I. K., Heckrath, G., and Olesen, J. E.: Carbon dynamics in topsoil and subsoil along a cultivated toposequence, Catena, 120, 20-28, 2014.

Church, J. A., Clark, P. U., Cazenave, A., Gregory, J. M., Jevrejeva, S., Levermann, A., Merrifield, M. A., Milne, G. A., Nerem, R. S., Nunn, P. D., Payne, A. J., Pfeffer, W. T., Stammer, D., and Unnikrishnan, A. S.: Sea Level Change. In: Climate Change: The Physical
35  Science Basis. Contribution of Working Group I to the Fifth Assessment Report of the Intergovernmental Panel on Climate Change, Stocker, T. F., Qin, D., Plattner, G.-K., Tignor, M., Allen, S. K., Boschung, J., and A. Nauels, Y. (Eds.), Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA., 2013.

Cline, J. D.: Spectrophotometric Determination of Hydrogen Sulfide in Natural Waters, Limnology and Oceanography, 14, 454-458, 1969.

Cooper, N. J.: The use of 'managed retreat' in coastal engineering, Proceedings of the Institution of Civil Engineers, 156, 101-110, 2003.

40  Dent, D.: Acid sulphate soils: a baseline for research and development, International Institute for Land Reclamation and Improvement, Netherlands, 1986.

Dubois, S., Savoye, N., Gremare, A., Plus, M., Charlier, K., Beltoise, A., and Blanchet, H.: Origin and composition of sediment organic matter in a coastal semi-enclosed ecosystem: An elemental and isotopic study at the ecosystem space scale, Journal of Marine Systems, 94, 64-73, 2012.

45  Dziejowski, J. E., Rimmer, A., and Steenhuis, T. S.: Preferential movement of oxygen in soils?, Soil Science Society of America Journal, 61, 1607-1610, 1997.

Fang, C. and Moncrieff, J. B.: The dependence of soil CO2 efflux on temperature, Soil Biology & Biochemistry, 33, 155-165, 2001.

FitzGerald, D. M., Fenster, M. S., Argow, B. A., and Buynevich, I. V.: Coastal impacts due to sea-level rise, Annual Review of Earth and Planetary Sciences, 36, 601-647, 2008.

Fors, Y., Nilsson, T., Risberg, E. D., Sandström, M., and Torssander, P.: Sulfur accumulation in pinewood (Pinus sylvestris) induced by bacteria in a simulated seabed environment: Implications for marine archaeological wood and fossil fuels, International Biodeterioration & Biodegradation, 62, 336-347, 2008.

Fossing, H. and Jørgensen, B.: Measurement of Bacterial Sulfate Reduction in Sediments: Evaluation of a Single-Step Chromium Reduction Method, Biogeochemistry, 8, 205-222, 1998.

Franzluebbers, A. J.: Soil organic carbon in managed pastures of the southeastern United States of America. In: Grassland carbon sequestration: management, policy and economics, Abberton, M., Conant, R., and Batello, C. (Eds.), Integrated Crop Management, Rome, 2010.

French, J. R.: Hydrodynamic modelling of estuarine flood defence realignment as an adaptive management response to sea-level rise, Journal of Coastal Research, 24, 1-12, 2008.

Fry, B., Scalan, R. A., and Parker, P. L.: Stable carbon isotope evidence for two sources of organic matter in coastal sediments: seagrasses and plankton, Geochimica et cosmochimica acta, Vol. 41, 1875-1877, 1977.

Glud, R. N.: Oxygen dynamics of marine sediments, Marine Biology Research, 4, 243-289, 2008.

Hall, P. O. J. and Aller, R. C.: Rapid, Small-Volume, Flow Injection Analysis for $\sum CO_2$ and $NH_4 +$ in Marine and Freshwaters, Limnology and Oceanography, 37, 1113-1119, 1992.

Hargrave, B. T., Holmer, M., and Newcombe, C. P.: Towards a classification of organic enrichment in marine sediments based on biogeochemical indicators, Mar Pollut Bull, 56, 810-824, 2008.

Hedges, J. I. and Keil, R. G.: Sedimentary organic matter preservation: an assessment and speculative synthesis, Mar Chem, 49, 81-115, 1995.

Herbert, E. R., Boon, P., Burgin, A. J., Neubauer, S. C., Franklin, R. B., Ardón, M., Hopfensperger, K. N., Lamers, L. P. M., and Gell, P.: A global perspective on wetland salinization: ecological consequences of a growing threat to freshwater wetlands, Ecosphere, 6, art206, 2015.

Holmer, M., Duarte, C. M., Heilskov, A., Olesen, B., and Terrados, J.: Biogeochemical conditions in sediments enriched by organic matter from net-pen fish farms in the Bolinao area, Philippines, Mar Pollut Bull, 46, 1470-1479, 2003.

Hursh, A., Ballantyne, A., Cooper, L., Maneta, M., Kimball, J., and Watts, J.: The sensitivity of soil respiration to soil temperature, moisture, and carbon supply at the global scale, Global Change Biol, 23, 2090-2103, 2017.

Jørgensen, B. B.: Bacteria and Marine Biogeochemistry. In: Marine Geochemistry, Schulz, H. D. and Zabel, M. (Eds.), Springer, Berlin Heidelberg New York, 2006.

Kalbitz, K., Solinger, S., Park, J. H., Michalzik, B., and Matzner, E.: Controls on the dynamics of dissolved organic matter in soils: A review, Soil Science, 165, 277-304, 2000.

Kim, Y. S. and Singh, A. P.: Micromorphological characteristics of wood biodegradation in wet environments: A review, Iawa Journal, 21, 135-155, 2000.

Kristensen, E.: Characterization of Biogenic Organic Matter by Stepwise Thermogravimetry (STG), Biogeochemistry, 9, 135-159, 1990.

Kristensen, E.: Decomposition of Macroalgae, Vascular Plants and Sediment Detritus in Seawater: Use of Stepwise Thermogravimetry, Biogeochemistry, 26, 1-24, 1994.

Kristensen, E. and Andersen, F. O.: Determination of organic carbon in marine sediments: a comparison of two CHN-analyzer methods, Journal of Experimental Marine Biology and Ecology, 109, 15-23, 1987.

Kristensen, E. and Hansen, K.: Decay of plant detritus in organic-poor marine sediment: Production rates and stoichiometry of dissolved C and N compounds, Journal of Marine Research, 53, 675-702, 1995.

Kristensen, E. and Holmer, M.: Decomposition of plant materials in marine sediment exposed to different electron acceptors ($O_2$, $NO_3-$, and $SO_4^{2-}$), with emphasis on substrate origin, degradation kinetics, and the role of bioturbation, Geochimica et cosmochimica acta, 65, 419-433, 2001.

Lovley, D. R. and Phillips, E. J.: Rapid assay for microbially reducible ferric iron in aquatic sediments, Applied and environmental microbiology, 53, 1536-1540, 1987.

MacDonald, J. D., Costello, L. R., and Berger, T.: An evaluation of soil aeration status around healthy and declining oaks in an urban environment in California, Journal of Aboriculture, 19, 209-219, 1993.

Neira, J., Ortiz, M., Morales, L., and Acevedo, E.: Oxygen diffusion in soils: Understanding the factors and processes needed for modeling, Chilean Journal of Agricultural Research, 75, 35-44, 2015.

Neubauer, S. C., Franklin, R. B., and Berrier, D. J.: Saltwater intrusion into tidal freshwater marshes alters the biogeochemical processing of organic carbon, Biogeosciences, 10, 8171-8183, 2013.

Pethick, J.: Estuarine and tidal wetland restoration in the United Kingdom: Policy versus practice, Restoration Ecology, 10, 431-437, 2002.

Portnoy, J. W.: Salt Marsh Diking and Restoration: Biogeochemical Implications of Altered Wetland Hydrology, Environ Manage, 24, 111-120, 1999.

Portnoy, J. W. and Giblin, A. E.: Biogeochemical effects of seawater restoration to diked salt marshes, Ecological Applications, 7, 1054-1063, 1997.

Quintana, C. O., Kristensen, E., and Valdemarsen, T.: Impact of the invasive polychaete Marenzelleria viridis on the biogeochemistry of sandy marine sediments, Biogeochemistry, 115, 95-109, 2013.

Reddy, K. R. and DeLaune, R. D.: Biogeochemistry of Wetlands: Science and Applications, CRC Press, 2008.

Rickard, D. and Morse, J. W.: Acid volatile sulfide (AVS), Mar. Chem., 97, 141-197, 2005.

5   Roman, C. T. and Burdick, D. M.: Tidal Marsh Restoration: A Synthesis of Science and Management, Island Press, Washington, DC, 2012.

Rustad, L. E., Campbell, J. L., Marion, G. M., Norby, R. J., Mitchell, M. J., Hartley, A. E., Cornelissen, J. H. C., Gurevitch, J., and Gcte-News: A meta-analysis of the response of soil respiration, net nitrogen mineralization, and aboveground plant growth to experimental ecosystem warming, Oecologia, 126, 543-562, 2001.

10  Schmidt, M. W. I., Torn, M. S., Abiven, S., Dittmar, T., Guggenberger, G., Janssens, I. A., Kleber, M., Kogel-Knabner, I., Lehmann, J., Manning, D. A. C., Nannipieri, P., Rasse, D. P., Weiner, S., and Trumbore, S. E.: Persistence of soil organic matter as an ecosystem property, Nature, 478, 49-56, 2011.

Schoepfer, V. A., Bernhardt, E. S., and Burgin, A. J.: Iron clad wetlands: Soil iron-sulfur buffering determines coastal wetland response to salt water incursion, Journal of Geophysical Research-Biogeosciences, 119, 2209-2219, 2014.

15  Six, J., Elliott, E. T., Paustian, K., and Doran, J. W.: Aggregation and soil organic matter accumulation in cultivated and native grassland soils, Soil Science Society of America Journal, 62, 1367-1377, 1998.

Stenak, M.: Inddæmningerne på Nordfyn. In: De inddæmmede landskaber - en historisk geografi, Landbohistorisk Selskab, 2005.

Stookey, L. L.: Ferrozine - A New Spectrophotometric Reagent for Iron, Analytical Chemistry, 42, 779-781, 1970.

Sullivan, J. T.: Cellulose and Lignin in Forage Grasses and Their Digestion Coefficients, Journal of Animal Science, 14, 710-717, 1955.

20  Sutton-Grier, A. E., Keller, J. K., Koch, R., Gilmour, C., and Megonigal, J. P.: Electron donors and acceptors influence anaerobic soil organic matter mineralization in tidal marshes, Soil Biology & Biochemistry, 43, 1576-1583, 2011.

Valdemarsen, T. and Kristensen, E.: Degradation of dissolved organic monomers and short-chain fatty acids in sandy marine sediment by fermentation and sulfate reduction, Geochimica et cosmochimica acta, 74, 1593-1605, 2010.

Valdemarsen, T., Kristensen, E., and Holmer, M.: Sulfur, carbon, and nitrogen cycling in faunated marine sediments impacted by repeated
25  organic enrichment, Marine Ecology Progress Series, 400, 37-53, 2010.

Valdemarsen, T., Quintana, C. O., Kristensen, E., and Flindt, M. R.: Recovery of organic-enriched sediments through microbial degradation: implications for eutrophic estuaries, Marine Ecology Progress Series, 503, 41-58, 2014.

Weiss, M. S., Abele, U., Weckesser, J., Welte, W., Schiltz, E., and Schulz, G. E.: Molecular architecture and electrostatic properties of a bacterial porin, Science, 254, 1627-1630, 1991.

30  Weston, N. B., Dixon, R. E., and Joye, S. B.: Ramifications of increased salinity in tidal freshwater sediments: Geochemistry and microbial pathways of organic matter mineralization, Journal of Geophysical Research, 111, 2006.

Weston, N. B., Vile, M. A., Neubauer, S. C., and Velinsky, D. J.: Accelerated microbial organic matter mineralization following salt-water intrusion into tidal freshwater marsh soils, Biogeochemistry, 102, 135-151, 2011.

Wolters, M., Garbutt, A., and Bakker, J. P.: Salt-marsh restoration: evaluating the success of de-embankments in north-west Europe,
35  Biological Conservation, 123, 249-268, 2005.

Yucel, M., Galand, P. E., Fagervold, S. K., Contreira-Pereira, L., and Le Bris, N.: Sulfide production and consumption in degrading wood in the marine environment, Chemosphere, 90, 403-409, 2013.

**Table 1 Mean values of water content, porosity and soil organic carbon (SOC) for all core sectionings. Error indicated as SEM (n = 15).**

|  | Depth (cm) | Water content (%) | | | Porosity | | | SOC (%) | | |
|---|---|---|---|---|---|---|---|---|---|---|
|  | 0.5 | 82.9 | ± | 0.7 | 0.82 | ± | 0.04 | 16.2 | ± | 0.8 |
|  | 1.5 | 75.5 | ± | 1.6 | 0.97 | ± | 0.02 | 16.1 | ± | 1.2 |
| Station UC | 3.5 | 60.5 | ± | 1.8 | 0.79 | ± | 0.01 | 11.0 | ± | 0.8 |
|  | 7.5 | 39.3 | ± | 0.9 | 0.60 | ± | 0.01 | 5.2 | ± | 0.2 |
|  | 12.5 | 33.0 | ± | 0.7 | 0.54 | ± | 0.01 | 3.5 | ± | 0.2 |
|  | 17.5 | 34.5 | ± | 0.8 | 0.56 | ± | 0.01 | 3.5 | ± | 0.2 |
|  | 0.5 | 32.0 | ± | 0.6 | 0.58 | ± | 0.02 | 1.4 | ± | 0.0 |
|  | 1.5 | 24.8 | ± | 0.5 | 0.53 | ± | 0.01 | 1.1 | ± | 0.0 |
| Station C | 3.5 | 21.6 | ± | 0.3 | 0.40 | ± | 0.01 | 1.0 | ± | 0.0 |
|  | 7.5 | 18.9 | ± | 0.4 | 0.35 | ± | 0.01 | 0.8 | ± | 0.1 |
|  | 12.5 | 17.9 | ± | 0.3 | 0.34 | ± | 0.00 | 0.9 | ± | 0.0 |
|  | 17.5 | 19.8 | ± | 0.4 | 0.37 | ± | 0.01 | 1.0 | ± | 0.0 |

**Table 2 Carbon budget table showing mean soil organic carbon (SOC) ± SEP (n = 18) in uncultivated (UC) and cultivated (C) soil. Total time integrated efflux and accumulation of total carbon dioxide (TCO$_2$) and dissolved organic carbon (DOC) in porewater are also shown.**

| Carbon budget (mol m$^{-2}$) | Station UC | Station C |
|---|---|---|
| Initial SOC pool | 710.9 ± 54 | 232.5 ± 22 |
| TCO$_2$ efflux | 39.9 | 11.2 |
| DOC efflux | 8.9 | 2.4 |
| TCO$_2$ porewater accumulation | 0.1 | 0.8 |
| DOC porewater accumulation | 0.7 | 0.5 |
| Total SOC degradation | 49.6 | 14.8 |
| Percentage of SOC pool degraded | 7 % | 6 % |

**Table 3 Budget table based on cumulative time integrated total degradation to carbon dioxide ($TCO_2$), and $TCO_2$ production and sulfate reduction (SR) measured in jar experiments (anaerobic rates) at different times after flooding (1 week [1W] and 2, 4, 6 and 12 months [2M, 4M, 6M and 12M, respectively). Relative contributions of SR and other anaerobic respiration processes to total anaerobic $TCO_2$ production (jar experiments) at different time points after flooding.**

| | | Station UC | | | | | Station C | | | | |
| --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- | --- |
| | | 1W | 2M | 4M | 6M | 12M | 1W | 2M | 4M | 6M | 12M |
| Cumulative degradation to $TCO_2$ (mol m$^{-2}$) | Total degradation to $TCO_2$ | 2.07 | 10.4 | 18.8 | 27.4 | 40.0 | 0.5 | 2.7 | 5.0 | 6.6 | 12.0 |
| | Anaerobic degradation to $TCO_2$ | 0.8 | 8.7 | 19.9 | 24.2 | 32.6 | 0.3 | 2.5 | 6.0 | 8.0 | 10.8 |
| | $TCO_2$ production by SR | 0.3 | 4.7 | 12.9 | 16.9 | 25.3 | 0.1 | 0.8 | 2.2 | 3.1 | 4.3 |
| Relative contributions (%) | SR | 42 | 63 | 91 | 100 | 100 | 29 | 30 | 54 | 23 | 93 |
| | Other anaerobic respiration | 58 | 37 | 9 | 0 | 0 | 71 | 70 | 46 | 77 | 7 |

**Table 4**

|  | $CO_2$ efflux (mmol m$^{-2}$ d$^{-1}$) | Study |
|---|---|---|
| Danish coastal marine sediments | 33 | Valdemarsen et al. (2010) |
|  | 24-45 | Valdemarsen et al. (2014) |
| Agricultural soils | 46-167 | Chirinda et al. (2014) |
|  | 69-111 | Fang and Moncrieff (2001) |
|  | 386-1374 | Rustad et al. (2001) |
|  | 182±96 | Hursh et al. (2017) |
| Soils flooded with seawater | 29-67 | This study |
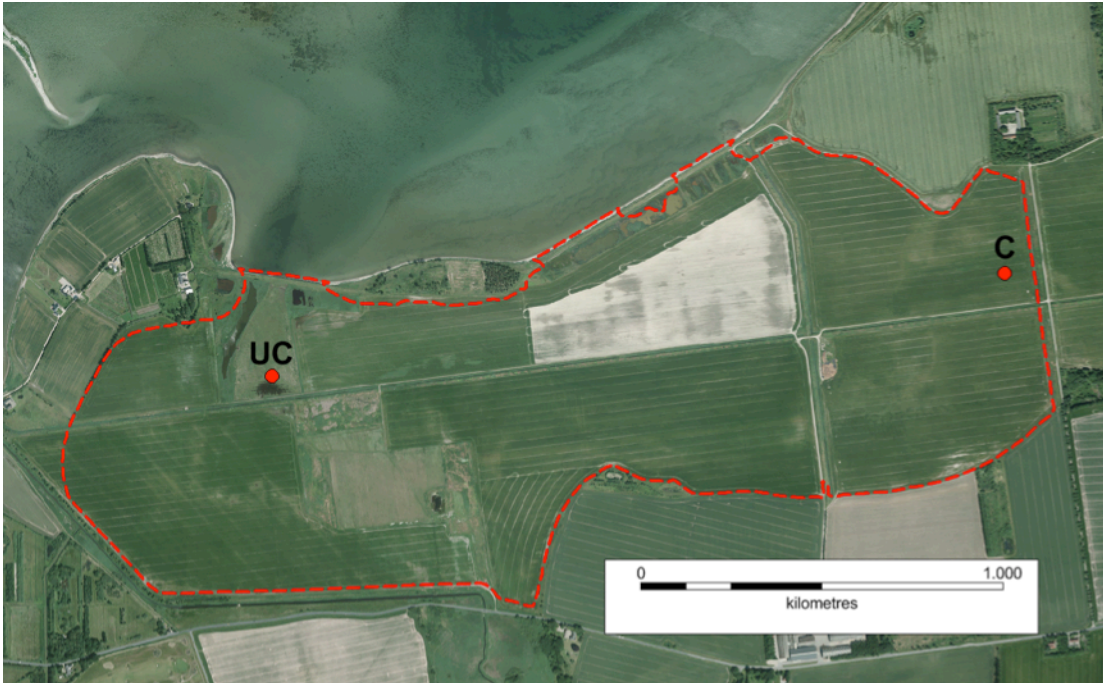
**Figure 1 Map of Gyldensteen Strand with the location of the 2 sampling stations for collecting uncultivated (UC) and cultivated (C) soil cores. The dashed red line indicates the area flooded with seawater in March 2014.**
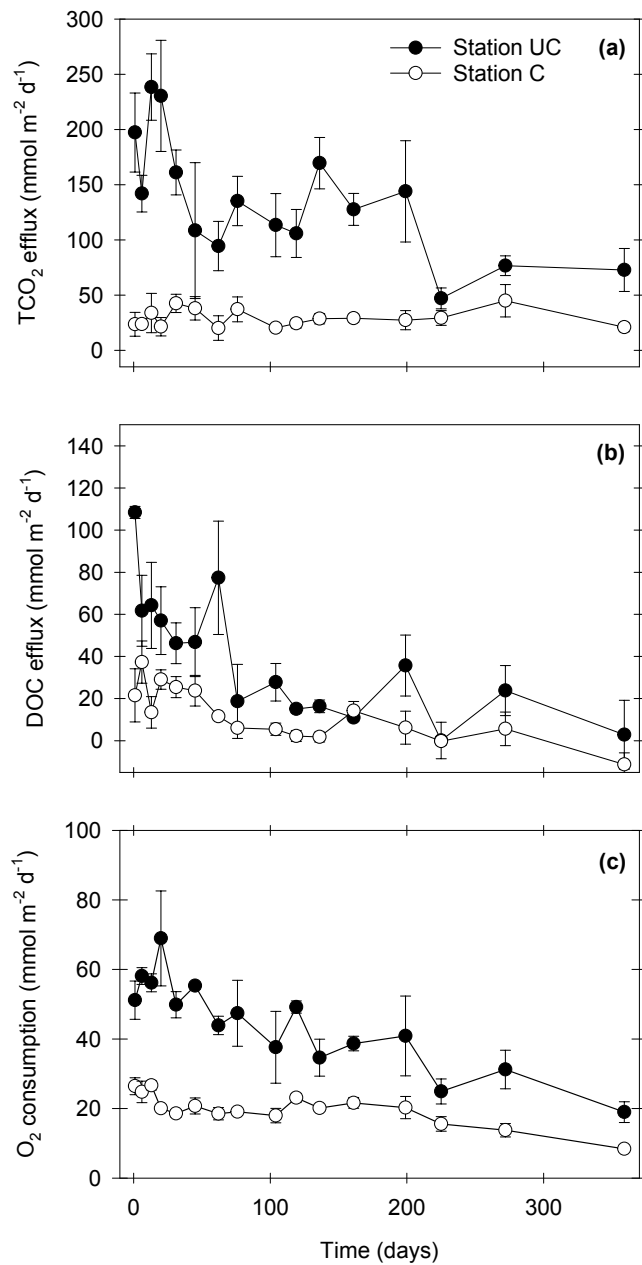
**Figure 2 Fluxes of total carbon dioxide (TCO₂, A), dissolved organic carbon (DOC, B) and oxygen (O₂) consumption (C) in soil cores with uncultivated (UC) and cultivated (C) soil after flooding. Error bars indicate SEM (n = 3).**
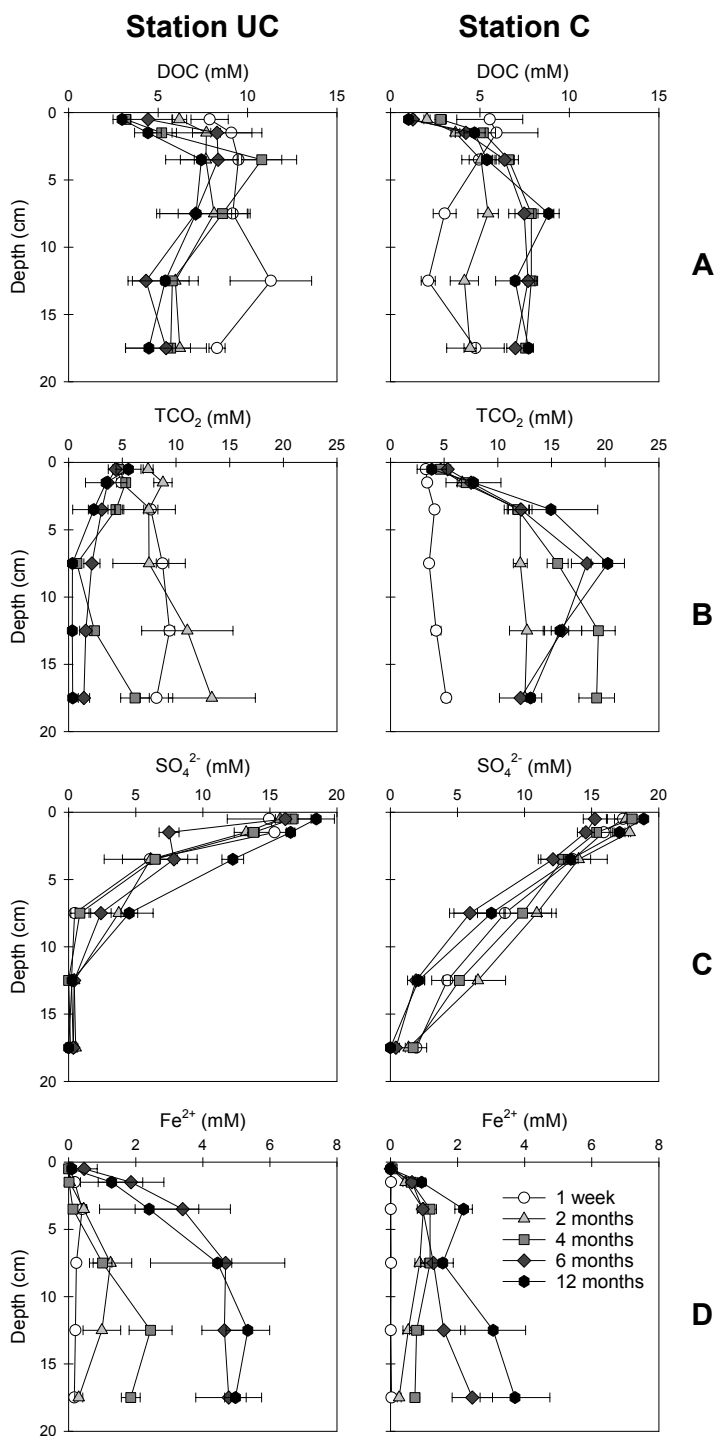
**Figure 3 Porewater profiles for dissolved organic carbon (DOC, A), total carbon dioxide (TCO$_2$, B), sulfate (SO$_4^{2-}$) (C) and Fe$^{2+}$ (D) in uncultivated (UC) and cultivated (C) soil flooded with seawater. Error bars indicate SEM (n = 3).**
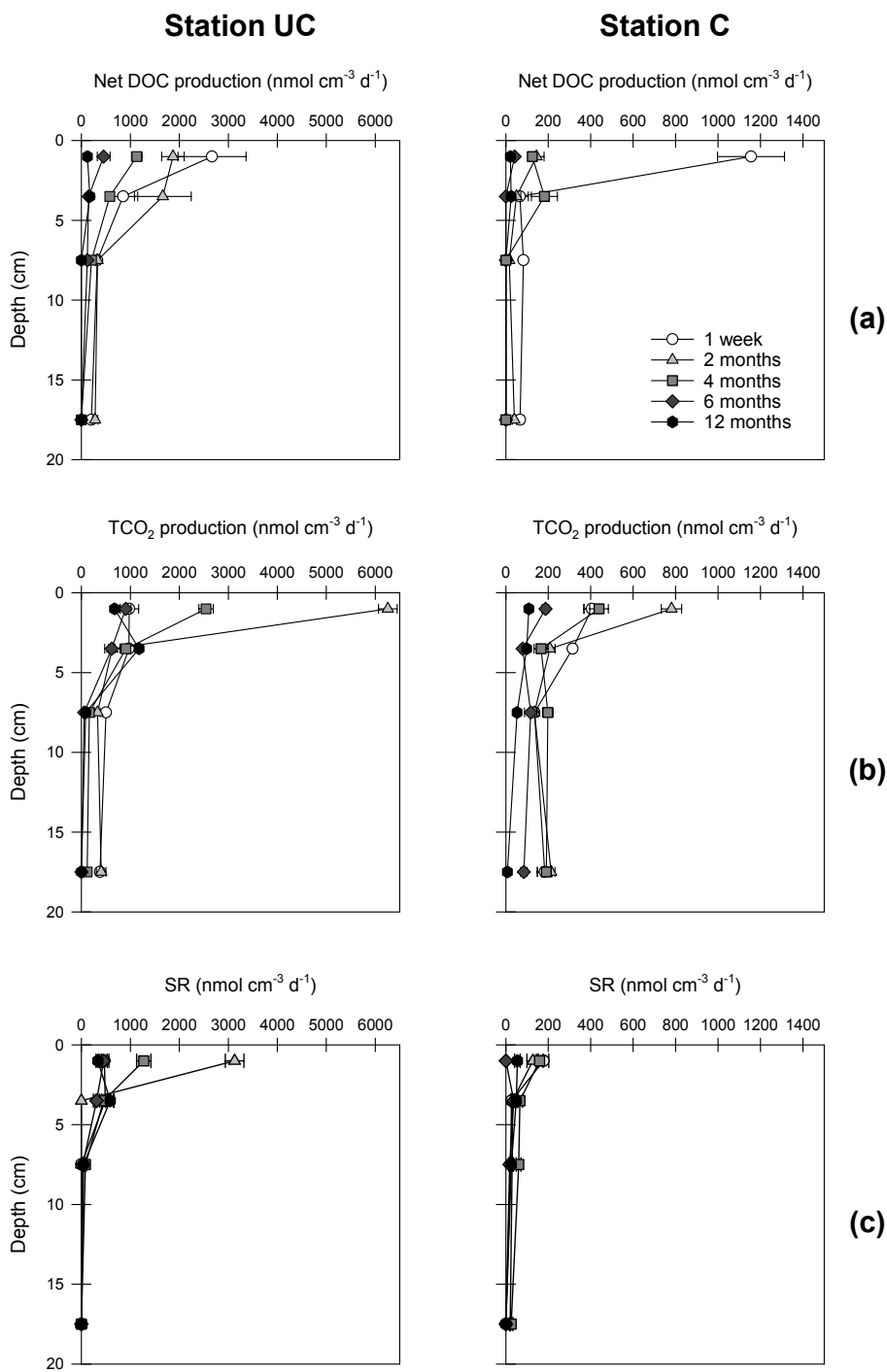
**Figure 4 Temporal and spatial variability in production of dissolved organic carbon (DOC, A) and carbon dioxide (TCO₂, B) and sulfate reduction (SR) measured in jar experiments with uncultivated (UC) and cultivated (C) soils flooded with seawater. Note the different x-axis scaling for station UC and C measurements. Error bars indicate SEM.**
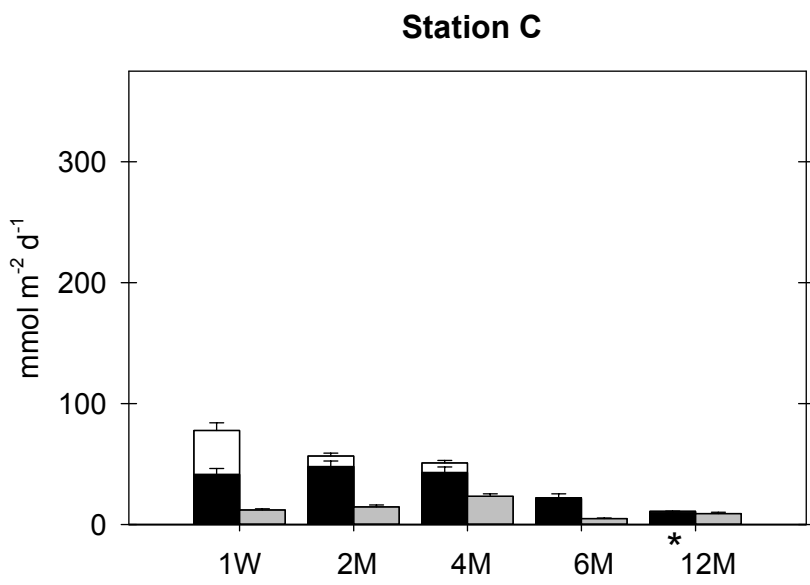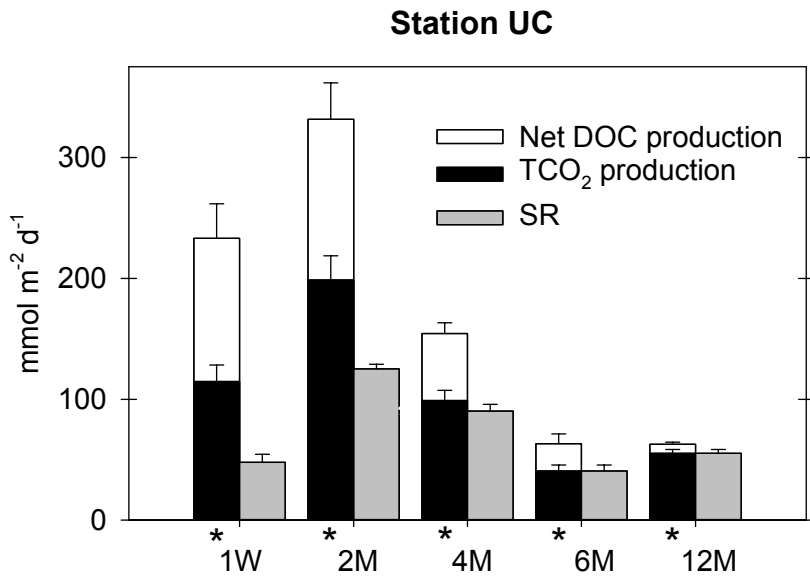
**Figure 5 Results from jar experiments showing area specific net production of dissolved organic carbon (DOC) and total carbon dioxide (TCO$_2$), and sulfate reduction (SR, based on SR rate measurements converted to C-units) in uncultivated (UC) and cultivated (C) soil at different times after flooding (1 week [1W] and 2, 4, 6 and 12 months [2M, 4M, 6M and 12M, respectively). In columns marked with *, TCO$_2$ production was corrected with 2 x SR. Error bars indicate SEP (n = 4).**
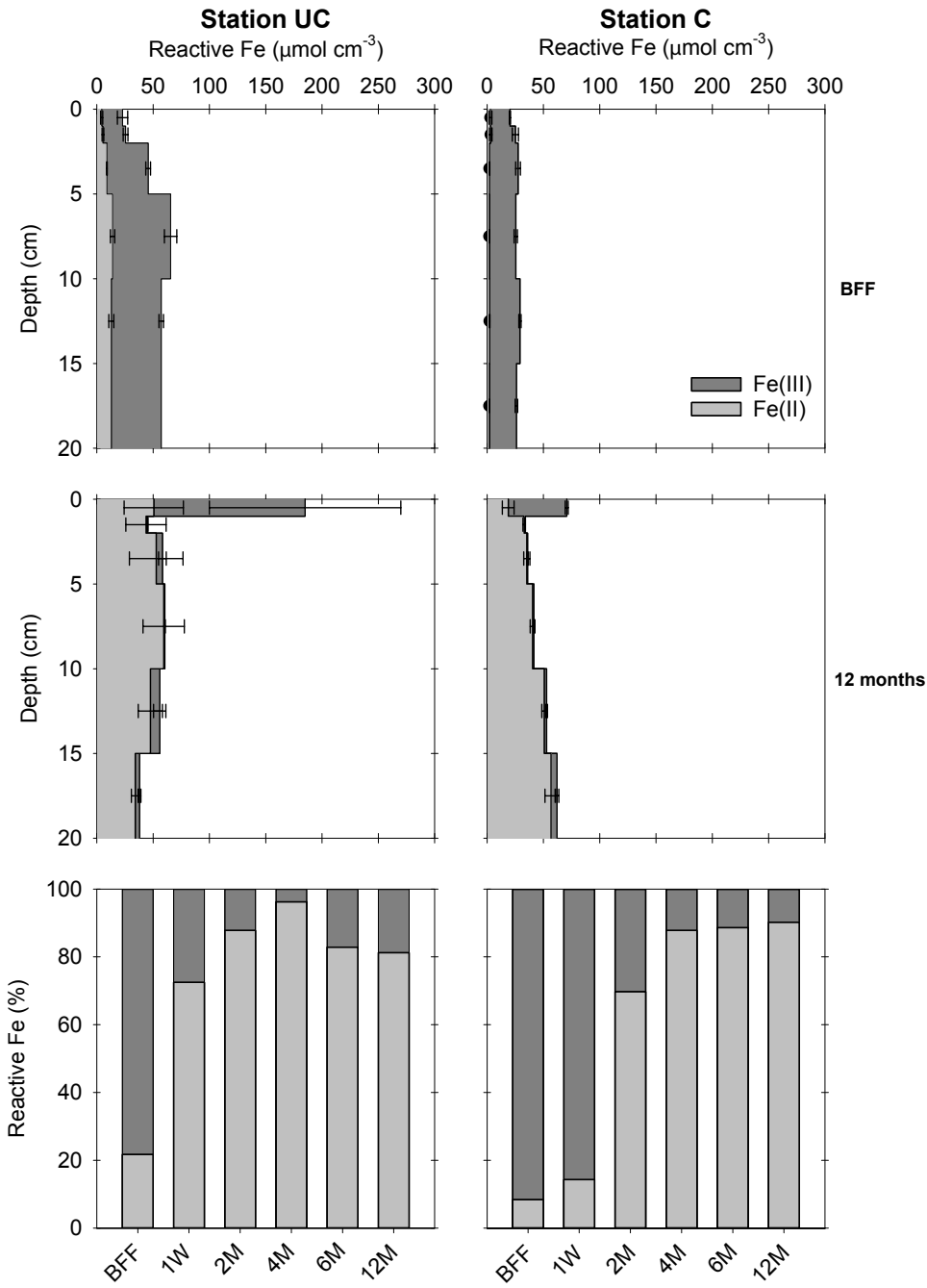
**Figure 6 Upper panels show concentration of reactive Fe(II) and Fe(III) in uncultivated (UC) and cultivated (C) soils before flooding (BFF) and 12 months after flooding. Lower panels show the relative contributions of reactive Fe(II) and Fe(III) in the upper 20 cm at various times after flooding (1 week [1W] and 2, 4, 6 and 12 months [2M, 4M, 6M and 12M], respectively). Error bars indicate SEM (n = 3).**
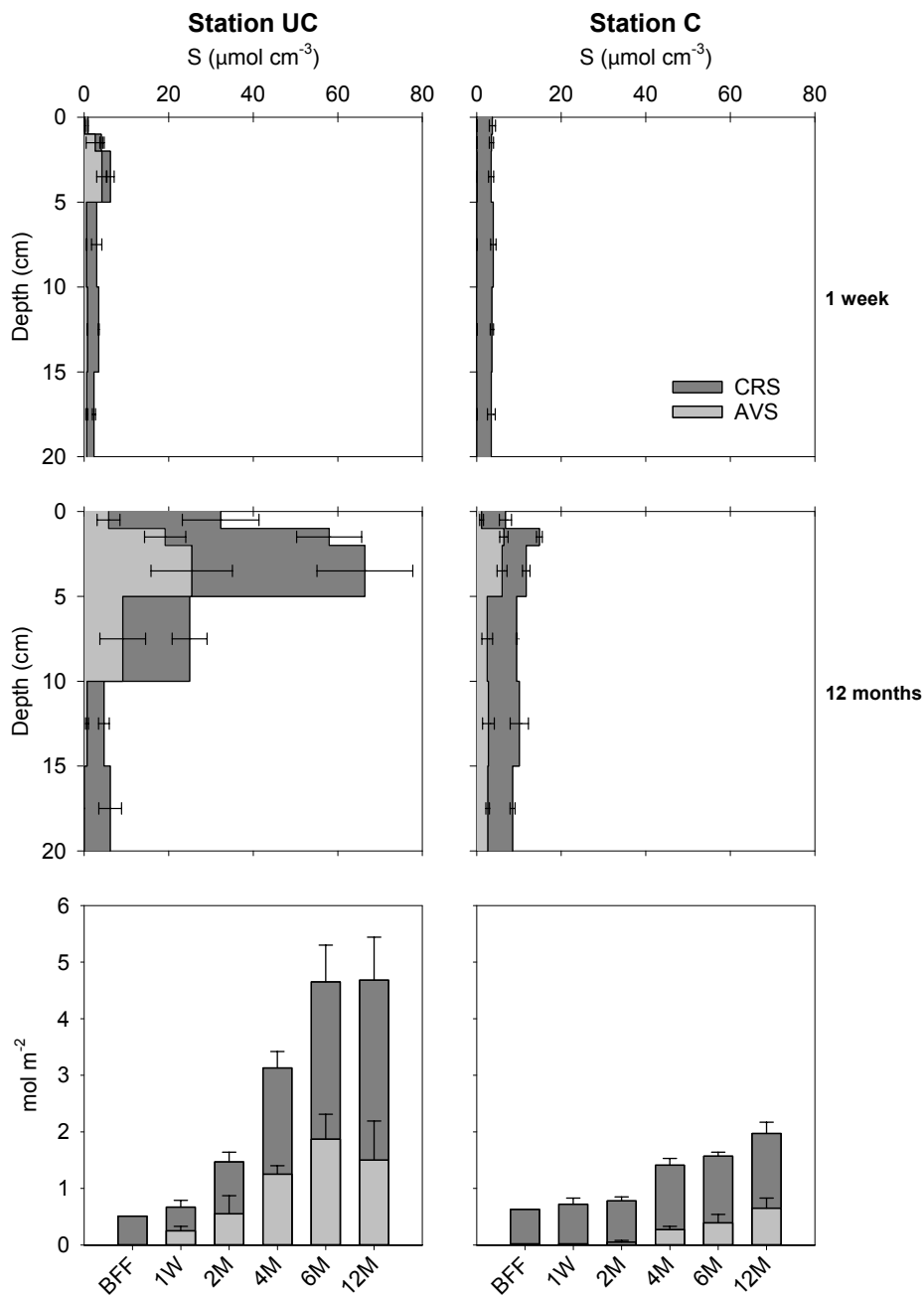
**Figure 7** Upper panels show concentration of chromium reducible sulfides (CRS) and acid volatile sulfides (AVS) in uncultivated (UC) and cultivated (C) soils before flooding (BFF) and 12 months after flooding. Lower panels show the depth integrated pools of AVS and CRS in the upper 20 cm at various times after flooding (1 week [1W] and 2, 4, 6 and 12 months [2M, 4M, 6M and 12M], respectively). Error bars indicate SEM (n = 3).