Human-induced and natural carbon storage in floodplains of the Central Valley of California

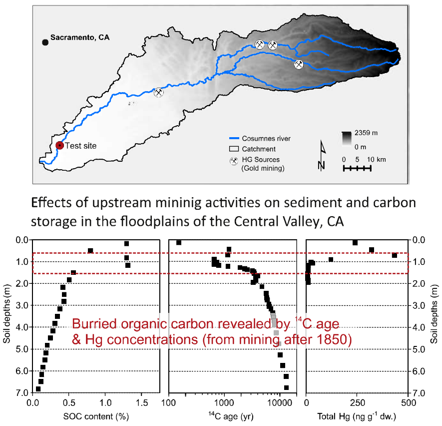
Kristin Steger [a,*,1], Peter Fiener [b,1], Mark Marvin-DiPasquale [c], Joshua H. Viers [d], David R. Smart [a]

[a] College of Agricultural and Environmental Sciences, Department of Viticulture and Enology, University of California, Davis, One Shields Avenue, Davis, CA 95616, USA
[b] Institute of Geography, Augsburg University, Alter Postweg 118, 86159 Augsburg, Germany
[c] US Geological Survey, 345 Middlefield Road, Menlo Park, CA 94025, USA
[d] Environmental Systems, School of Engineering, University of California, Merced, 5200 North Lake Road, CA 95340, USA

HIGHLIGHTS

- Evaluation of depth-dependent SOC contents in a floodplain area of California
- About 60% of the entire SOC stored within the 7 m profiles found in the upper 2 m
- Radiocarbon dating and mercury analysis showed a substantial sedimentation phase.
- This phase was associated with upstream hydraulic gold mining after the 1850s.

GRAPHICAL ABSTRACT



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Anthropogenic influences
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ABSTRACT

Active floodplains can putatively store large amounts of organic carbon (SOC) in subsoils originating from catchment erosion processes with subsequent floodplain deposition. Our study focussed on the assessment of SOC pools associated with alluvial floodplain soils that are affected by human-induced changes in floodplain deposition and in situ SOC mineralisation due to land use change and drainage. We evaluated depth-dependent SOC contents based on 23 soil cores down to 3 m and 10 drillings down to 7 m in a floodplain area of the lower Cosumnes River. An estimate of 266 Mg C ha^{-1} or about 59% of the entire SOC stored within the 7 m profiles was found in the upper 2 m. Most profiles (n = 25) contained discrete buried A horizons at depths of approximately 0.8 m. These profiles had up to 130% higher SOC stocks. The mean $\delta^{13}C$ of all deep soil profiles clearly indicated that arable land use has already altered the stable isotopic signature in the first meter of the profile. Radiocarbon dating showed that the ^{14}C age in the buried horizon was younger than in overlaying soils indicating a substantial sedimentation phase for the overlaying soils. An additional analysis of total mercury contents in the soil profiles indicated that this sedimentation was associated with upstream hydraulic gold mining after the 1850s. In summary, deep alluvial soils in floodplains store large amounts of SOC not yet accounted for in global carbon models. Historic data give evidence that large amounts of sediment were transported into the floodplains

* Corresponding author at: University of Freiburg, 79085 Freiburg i. Br., Germany.
E-mail address: ksteger@ucdavis.edu (K. Steger).
[1] Equal contribution of both authors.

of most rivers of the Central Valley and deposited over organically rich topsoil, which promoted the stabilization of SOC, and needs to be considered to improve our understanding of the human-induced interference with C cycling.

1. Introduction

Riverine floodplains occupy only about 1.3% of Earth's land surface (Sutfin et al., 2016; Tockner and Stanford, 2002), but they deserve growing attention as important sinks of terrestrial carbon (Hasada and Hori, 2016; Hoffmann et al., 2009; Notebaert et al., 2014; Ricker and Lockaby, 2015; Sutfin et al., 2016). Global estimates of carbon (C) storage in floodplains vary between 1.4 and 7735 Mg C ha^{-1} (Appling, 2012; Sutfin et al., 2016; Wigginton et al., 2000). There are two main reasons why floodplains store substantial amounts of organic C: (i) the in situ conditions are often favorable for C sequestration because C inputs via photosynthetic assimilation products are more or less optimal (Baldock and Skjemstad, 2000) and soil organic carbon (SOC) mineralisation is often limited by saturated conditions; (ii) floodplains receive substantial amounts of SOC associated with deposited sediments coming from the entire catchment of such floodplains (Robertson et al., 1999).

Under natural conditions floodplains seem to be a continuous long-term carbon sink, as shown by Hoffmann et al. (2009) quantifying the SOC storage in the floodplains of the River Rhine in the entire Holocene. However, human-induced changes like river management, land use along rivers and land use patterns of the catchment have altered an increasing number of floodplains and their catchments (Poff et al., 2007). For instance, along European rivers as much as 79% of the riparian areas are intensively cultivated compared to 46% for North American rivers (excluding Alaska and northern Canada) and 11% for African rivers (Tockner and Stanford, 2002). Anthropogenic changes affecting floodplain SOC storage are complex as in situ changes often are intertwined with changes within the entire catchment. Historically, floodplains with natural riparian vegetation were often converted into arable land following river regulation (i.e., flood control dyking) and wetland drainage (Zedler and Kercher, 2005). Moreover, river regulation fundamentally alters the natural flow regime often resulting in the reduction of floodwater magnitude, duration and frequency, which in turn alters biogeochemical cycling (Poff et al., 1997). The net effect of land conversion, including loss of riparian vegetation and the drainage of floodplains, coupled with flow alteration and its reduction in floodwater inundation and sediment deposition, or in some sites, cessation of peat formation and peat decay, is a reduced C sequestration potential in alluvial soils. However, it is important to note that despite the globally pervasive and persistent decline in natural floodplain areas, there are also efforts underway to restore natural floodplain systems (Bullinger-Weber et al., 2014; D'Elia et al., 2017; Florsheim and Mount, 2003; Schiemer et al., 1999). In contrast to the loss of C due to in situ changes in drainage and use of floodplain soils, these areas might gain more C via lateral input from catchments, in cases where human induced land use change in the upper catchment leads to accelerated erosion and hence sediment input into the floodplains. The importance of this C input was underlined by a recent study of Wang et al. (2017) estimating that globally about 1/3 of eroded SOC due to land use change was stored in colluvial and especially alluvial soils during the last centuries to millennia.

Studying human impacts on C sequestration in floodplains is difficult as these impacts are mostly long lasting, weakly documented and occurring diffusely along the rivers and within the catchments. The Central Valley of California represents a specific situation, which partly allows to unravel the human impact and compare it to natural conditions. Until the 1850s the human impact in the region was relatively minor. This changed with the onset of the Californian Gold Rush, leading to a substantial change in the headwater catchments of the Sierra Nevada. By the mid-1850s and onwards, the headwater catchments were affected by hydraulic mining as the most cost-effective method to recover large amounts of gold in areas with sufficient surface water (Alpers et al., 2005; Gilbert, 1917; James, 2005). This powerful method delivered significant amounts of sediment to the floodplains of the Central Valley, with an estimate of 260×10^6 m^3 of mining debris reaching the San Pablo Bay in the northern part of San Francisco Estuary between 1856 and 1887 (Jaffe et al., 2007). Due to the utilization of mercury (Hg) in processing gold, these sediments were heavily Hg contaminated (Alpers et al., 2005; Drexler et al., 2009; Hornberger et al., 1999). Parallel to the gold mining activities, millions of hectares of wetlands were granted by the newly founded government of California to encourage the drainage of lands (Robinson et al., 2016). Levees were built to constrain anastomosing rivers to single channels and to separate wetlands from tidal waters restraining floods that historically filled the basins. Dense stands of tule (*Schoenoplectus acutus*) and willows (*Salix* spp.) were flattened or burned to make way for farmland and contributed as non-mining waste to sediment loads to the Central Valley (van Geen et al., 1999). By 1930, the majority of the land was virtually entirely cultivated (Whipple et al., 2012).

The major objective of this study is to analyze the human-induced effects of changing river sediment dynamics and land use on carbon storage in relation to the natural processes of SOC storage within a floodplain of the Cosumnes River, California.

2. Materials and methods

2.1. Sampling site

The study area was located approximately 30 km south of Sacramento, California, and north of a restored floodplain-riparian habitat, the Cosumnes River Preserve (38°17′49″N, 121°23′10″W; 5–6 m a.s.l.). In the last 150 years, the site was used for agricultural production (most recently for a variety of row crops) and associated with a significant degradation of the floodplain ecosystem. Prior to this anthropogenic disturbance, the lower Cosumnes River was an anastomosing channel network with perennial floodplain lakes (Constantine et al., 2003; Florsheim and Mount, 2003) (Fig. 1). The Cosumnes River, a tributary to the Sacramento – San Joaquin Delta, drains a 2460 km^2 watershed and is the last major river draining the western Sierra Nevada without any large regulating dams, and as such it rapidly responds to precipitation events (Nichols and Viers, 2017). Basin headwaters are located at an elevation of approximately 2400 m a.s.l. within a complex assemblage of granitic, andesitic, and metamorphic rocks that are part of the Sierra Nevada geomorphic province. The lower Cosumnes River ultimately enters the Great Valley geomorphic province with Pleistocene alluvium and river terraces generated during multiple Plio-Pleistocene episodes of valley incision and filling (Nichols and Viers, 2017). Floodplain restoration efforts have been implemented at the site since fall 2011 with the overall objective to hydrologically restore much of the multichannel system and the associated vegetation (D'Elia et al., 2017).

The Mediterranean climate in the Cosumnes River catchment is characterized by cool wet winters and hot dry summers (mean annual temperature 16 °C). The majority of precipitation occurs as rain in the winter and spring months (between December and March; annual

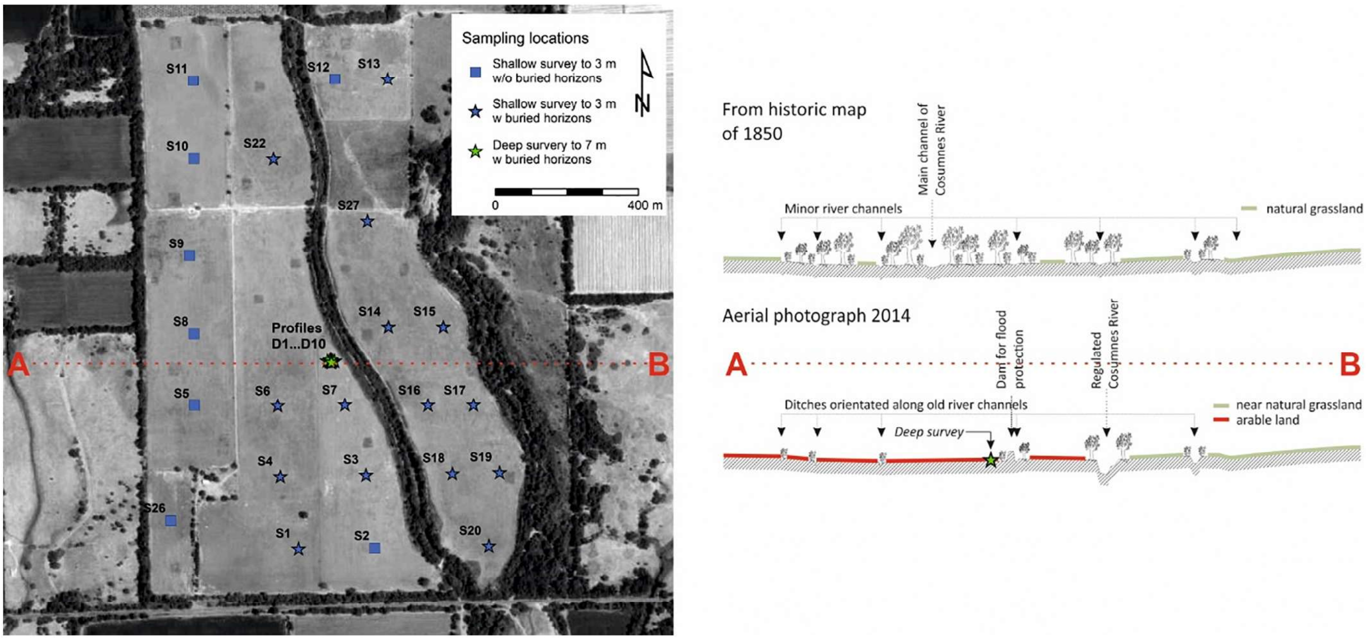

Fig. 1. Cosumnes River test site with sampling locations for shallow (0–3 m), raster survey performed in 2012 (S1…S27) and deep (0–7 m) survey performed in 2015/2016 (D1–D10). Note the cross-section A-B was derived from a recent aerial photograph and a historic map from 1850; the Y-axis of the cross-section is not scaled.

mean 445 mm at nearby Sacramento meteorological station), and flood events happen throughout the wet season with either rain-on-snow or late snowmelt floods (Whipple et al., 2017).

The soils at the site are mapped as Cosumnes silt loams (NRCS, 2013), and all samples were taken in the same mapping unit.

2.2. Soil sampling

A transect of six sites (D1–D6) in west-east direction and a transect of four sites (D7–D10) in north-south direction was established, with soil samples collected in November 2015 and February 2016, respectively. Both transects covered an area of 1 ha, and core sites were drilled down to 7 m, except for D1 = 6 m. All soil samples were taken with a Geoprobe® machine (Salina, Kansas) using the DT22 system with an inner rod string diameter of 28.5 mm. Each sampled liner represented 1 m of soil depth, which was cut and extracted in 0.33 m increments. For soil core D7 smaller increments (e.g., 0.05 m and 0.10 m) were taken below 1 m (see Table 1). After thorough homogenization of the material of each soil increment a portion (ca. 35 g) was collected in 20 mL glass vials for water content measurement and ^{14}C analysis. The remaining soil was stored in paper/plastic bags in a cooler until and during transport to the laboratory.

A previous soil sampling conducted during 2012 surveyed a larger area with 23 sites within the same general study area (Fig. 1). In this case, all soil samples were taken by hand auger to a depth of 3 m (see D'Elia et al., 2017), and similarly sub-sampled as described above. This earlier study focused on SOC stocks and effects of restoration of floodplains. Furthermore, it presented a description about the old river channels and the native vegetation in the floodplain (D'Elia, 2015).

2.3. Soil physical-chemical parameters

Soil samples in the glass vials were dried at 105 °C for 48–72 h, and the gravimetric water content was determined. For soil core D8 with 21 samples, the hydrometer method was used to measure particle size distribution (Sheldrick and Wang, 1993). Additionally, these 21 soil samples were tested for inorganic C in form of carbonates which can interfere with the measurement of organic ^{13}C (Harris et al., 2001). For this test, 70–100 mg soil was weighed into silver capsules, wetted with small amounts of water and placed in a desiccator containing a beaker with 12 M hydrochloric acid (HCl). Carbonates are released as CO_2 in 6 to 8 h. Prior to ^{13}C analysis the samples were dried at 60 °C. As these HCl treated samples gave the same results as untreated samples we concluded that total C = soil organic C (SOC).

2.4. Carbon, nitrogen and isotopic analyses

A total of 225 soil samples were ground with mortar and pestle to pass a 60-mesh (250 μm) sieve and weighed 50–90 mg into tin capsules (5 × 9 mm for solids, Costech, Valencia, CA) for analysis of total carbon, total nitrogen, and the natural abundance of $\delta^{13}C$ and $\delta^{15}N$ isotopes. These analyses were performed at the UC Davis Stable Isotope Facility using an Elementar Vario Micro Cube elemental analyzer (Elementar, Hanau, Germany) interfaced to Isoprime VisION isotope ratio mass

Table 1
Soil depth profile D7 for SOC concentration, ^{14}C age, nitrogen isotopy as $\delta^{15}N$ and total mercury (Hg) concentration to a depth of 2 m (n.a.: not analyzed/determined).

Soil depth (m)	SOC concentration (g/kg)	Calibrated ^{14}C age (cal yr BP)	$\delta^{15}N$ concentration (‰)	Hg concentration (ng g^{-1} dw.)
0.00–0.30	12.4	152	5.66	242
0.30–0.60	7.42	1226	5.36	322
0.60–0.80	9.44	1155	5.28	434
0.80–1.00	16.5	652	4.11	123
1.00–1.05	17.1	737	3.17	24.6
1.05–1.10	17.1	694	3.14	32.6
1.10–1.15	17.2	652	3.09	n.a.
1.15–1.20	15.0	787	3.52	12.8
1.20–1.25	12.3	1164	3.90	10.3
1.25–1.30	10.5	1922	4.22	11.4
1.30–1.35	8.81	2337	4.46	n.a.
1.35–1.40	7.28	2584	4.78	n.a.
1.40–1.45	5.49	3177	4.98	11.6
1.45–1.50	4.72	3367	5.12	n.a.
1.50–1.60	4.36	3451	4.87	9.70
1.60–1.70	4.18	3945	4.81	n.a.
1.70–1.80	4.85	3628	4.39	9.50
1.80–1.90	5.04	3695	4.37	10.5
1.90–2.00	5.73	3317	4.46	11.1

spectrometer (Isoprime, Manchester, UK). Samples were combusted at 1080 °C in a reactor packed with tungsten (VI) oxide. Following combustion, oxides were removed in a reduction reactor (reduced copper at 650 °C). The helium carrier then flowed through a water trap (magnesium perchlorate and phosphorous pentoxide). N_2 and CO_2 were separated using a molecular sieve adsorption trap before entering the isotope ratio mass spectrometer (IRMS).

Analytical batch runs were interspersed with several replicates of at least three different laboratory reference materials. These laboratory reference materials, which were selected to be compositionally similar to the samples being analyzed, have been previously calibrated against Standard Reference Materials (IAEA-N1, IAEA-N2, IAEA-N3, USGS-40, and USGS-41). A sample's preliminary isotope ratio was measured relative to reference gases analyzed with each sample. These preliminary values were finalized by correcting the values for the entire batch based on the known values of the included laboratory reference materials. The long term standard deviation was 0.2‰ for ^{13}C and 0.3‰ for ^{15}N.

2.5. Radiocarbon dating

^{14}C analyses were conducted at the Center for Accelerator Mass Spectrometry (CAMS) at Lawrence Livermore National Laboratory. An acid-fume pretreatment was not necessary as our samples did not contain carbonates (see above Soil physical-chemical parameters section). Samples were combusted and converted to CO_2 in individually sealed quartz tubes with CuO and Ag. The CO_2 was purified, trapped, and converted to graphite using an iron catalyst, following a method similar to that described by Vogel et al. (1984). The graphite targets were measured on the Van de Graaff FN accelerator mass spectrometer at CAMS. The ^{14}C results were reported as radiocarbon years using the Libby half-life of 5568 years and following the conventions of Stuiver and Polach (1977). The results included a $\delta^{13}C$ correction for isotope fractionation (Stuiver and Polach, 1977), and a blank subtraction based on ^{14}C-free coal. The dates were calibrated with the radiocarbon calibration program CALIB rev 7.1.0 (Reimer et al., 2013) and finally, the ^{14}C age was reported as calibrated years before present (cal years BP).

2.6. Mercury analysis

Soil total mercury (THg) was quantified by a standard U.S. Geological Survey method (Olund et al., 2004), with modifications to the sample digestion. A sub-sample of freeze dried soil (0.07–0.3 g, exact weight determined) was initially digested with aqua regia (2 mL of concentrated HNO_3 and 6 mL of concentrated HCl) in Teflon digestion bombs overnight at room temperature. Subsequently, 22 mL of 5% bromine monochloride (BrCl) was added to each sample, and the bombs were heated to 50 °C in an oven overnight. Once cooled, a sub-sample of the digestate (0.1–0.5 mL) was analyzed on an Automated Mercury Analyzer (Tekran Model 2600, Tekran, Inc., Canada), according to USEPA Method 1631, Revision E (USEPA, 2002). Quality assurance consisted of (i) the percent deviation of analytical duplicates (mean = 2.7%, n = 2 sample pairs); (ii) percent recovery of certified reference material (PACS-3 marine sediment, 88% of certified value, n = 2); and (III) and matrix spike recoveries (111 ± 3%, n = 3).

2.7. Other soil analyses

A buried soil horizon in this study is defined as a soil layer below 0.3 m that did not have significantly different SOC concentration than the current soil surface, but at least 50% more C than the overlaying segment (Wills et al., 2014).

Bulk soil density was calculated using organic matter, sand and clay contents following the equations developed by Saxton and Rawls (2006). Finally, the SOC stocks (Mg C ha^{-1}) were calculated for each depth increment (C_i) using the following equation according to D'Elia et al. (2017):

$$C_i = D_b * z_i * C_w * 10,000 \quad (1)$$

where D_b is bulk density (g cm^{-3}), z_i is depth (m), and C_w is the measured C concentration (% w/w). A correction for stone content was not necessary with these soils.

All statistical analyses were done using R for statistical computing 3.2.2 (R Development Core Team, Austria).

3. Results

3.1. Organic carbon dynamics in floodplain soils (0–7 m)

Our deep soil survey (Fig. 1) gave an example for the large carbon storage in floodplain soils of the Central Valley (Fig. 2a). Within the 7 m profiles of our floodplain soils, slightly >58% (or approximately 266 Mg C ha^{-1}) of the entire SOC was stored in the upper 2 m. The profiles below 2 m showed a continuous development with a linear decrease in SOC and N concentrations with increasing soil depths (Fig. 2a, b). The decline in SOC concentration below 2 m was more pronounced than the decline in N concentration resulting in a steeper reduction of the C/N ratio (Fig. 2c), especially below 3 m depth. Below 2 m the calibrated ^{14}C ages also revealed a more or less continuous decline ranging from about 4700 to 13,000 yrs. BP (Fig. 3c).

In comparison to the soil layers below 2 m, the SOC and N concentrations and the calibrated ^{14}C ages in the upper 2 m exhibited a substantial discontinuous development with increasing soil depths. The carbon rich topsoil layer (0–0.33 m) was followed by a carbon depleted layer (0.33–0.66 m), but between 0.66 and 1.33 m SOC and N values showed again similar concentrations as the topsoil (Fig. 2a, b). This soil layer had a noticeably darker color and a much higher silt and clay content compared to the overlaying soil layer (Fig. 3a). Median calibrated ^{14}C ages declined also discontinuously from topsoil to a depths of 2.0 m (Table 1), where the more sandy material between 0.3 and 0.8 showed higher ^{14}C ages (approx. 1200 yrs. BP) compared to the underlying more clay and silt containing horizon (approx. 650 yrs. BP).

3.2. Anthropogenic effects upon the organic carbon sedimentation in floodplain soils

The SOC, N, ^{14}C and texture data clearly revealed a phase of substantial sedimentation occurring in the younger history of the floodplains (Figs. 2a, b and 3a, c). Interestingly, the deposited material (0.3–0.8 m at D7) had much older ^{14}C ages (Table 1) indicating that the source of the sediment was not topsoil from the catchment, which should had similar ages as the buried A horizon. The median calibrated ^{14}C age of the buried horizon (between 0.8 and 1.0 m at D7) was 652 years. Assuming that the calibrated bulk ^{14}C age of the topsoil layer at the time of burial was similar to the today's ^{14}C topsoil age allows a first estimate that the buried horizon was not much older than 400 years. However, substantial anthropogenic changes have not occurred prior to 1850 making the ^{14}C ages only an indication but not a proof of anthropogenic impacts. Historic data give evidence that large amounts of sediment were transported into the floodplains of most rivers of the Central Valley (Gilbert, 1917; Jaffe et al., 2007; James, 2005) giving rise to the idea that the massive sedimentation resulted mostly from hydraulic mining and to a smaller extent from accelerated topsoil erosion due to land use change in the catchment. Gold mining is associated with mercury (Hg) application used extensively to amalgamate fine gold in both hard rock and placer mining in California during the Gold Rush era (Alpers et al., 2005; Nriagu, 1994; Nriagu and Wong, 1997). Hence, we used Hg concentrations in two profiles (D7 and D8) as fingerprints of the origin of theses sediments (as done in Bouse et al., 2010, Marvin-DiPasquale et al., 2003). Both profiles were evidently contaminated

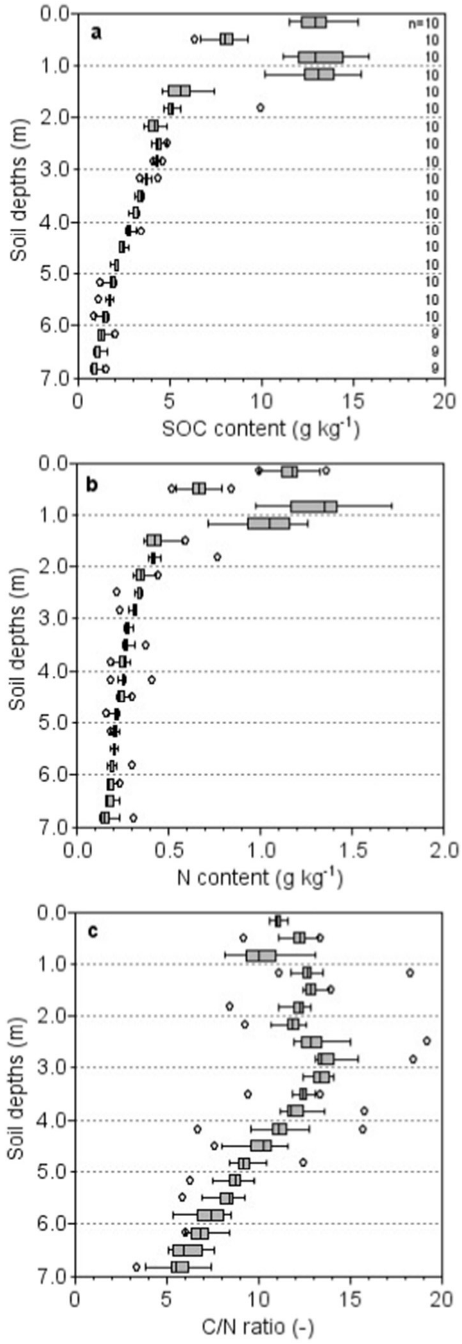

Fig. 2. Box plots of all soil depth profiles of 0–7 m for (a) soil organic carbon (SOC), (b) nitrogen (N) concentration, and (c) C/N ratio (soil depth indicates the mid of each layer).

with Hg in depths between 0 and 0.8 m (Fig. 4). The peak Hg concentrations between 300 and 450 ng g^{-1} coincided with the layer of relatively old C in the soil (Table 1), whereas below 0.8 m with the highest C concentrations in profile D7 (analyzed in 0.05 m increments) Hg sharply decreased and reached background concentrations (<50 ng g^{-1}; Hornberger et al., 1999, Singer et al., 2013) below 1.1 m (Table 1).

The results of the Hg analysis verify our assumption that the soil material above the buried horizon resulted from substantial sedimentation initiated by anthropogenic activities following the 1850s. To estimate the change in carbon stocks due to this burial it was necessary to identify reference sites without buried horizons. The data of an earlier shallow sampling (0–3 m) at the raster points S1 to S27 indicated that 8 out of 23 profiles did not show buried horizons below 0.66 m (Fig. 1). Comparing the mean SOC stocks of all shallow and deep profiles with and without buried horizons

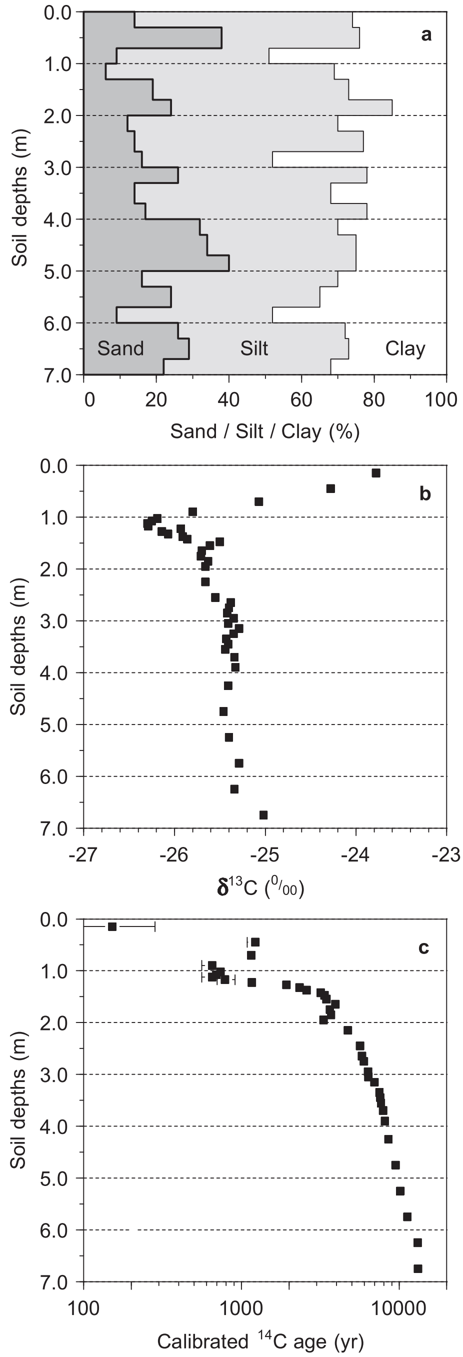

Fig. 3. Particle size distribution at profile D8 (a), mean carbon isotopy as $\delta^{13}C$ at all D profiles (b), and radiocarbon ages ^{14}C at profile D7 (c) (location see Fig. 1). Error bars of radiocarbon dating representing the calibrated C age range (min–max).

showed substantially higher SOC stocks in the profiles with buried horizons (Fig. 5). SOC stocks in the soil layers below 0.66 m were up to 130% larger for profiles with buried horizons. On average, the estimated SOC stocks of these 3-m profiles were 83 Mg C ha^{-1} higher than profiles without buried horizons indicating a significant ($p < 0.001$) SOC gain as a result of anthropogenic sediment production.

4. Discussion

4.1. Organic carbon dynamics in (natural) floodplain soils (0–7 m)

A previous study in our floodplain site has revealed that agricultural cultivation results in loss of soil and the limitation of carbon

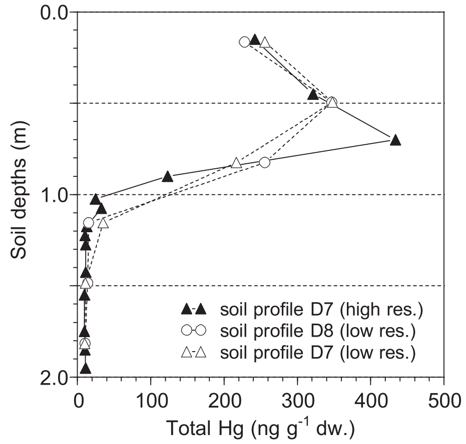

Fig. 4. Total mercury (Hg) concentrations in two soil profiles (D7 and D8) along a depth of 2.0 m (soil depth indicates the mid of each soil layer). The high resolution profile of D7 was numerically degraded to a low resolution profile in order to compare with soil profile D8.

sequestration in these soils (D'Elia, 2015). Moreover, it has been shown that soils down to 3 m contain significantly more SOC than typical carbon stock calculations. Nonetheless, our deep soil survey has revealed even larger carbon storage potential in floodplain soils of the Central Valley (Fig. 2a). Converting C concentrations to C stocks for estimating SOC pools at the landscape scale, D'Elia et al. (2017) reported 47 Mg C ha^{-1} in undisturbed reference sites in the 0–15 cm surface layer, whereas our floodplain soils revealed an estimate of 52 Mg C ha^{-1} in the 0–33 cm surface layer. Moreover, an estimate of 266 Mg C ha^{-1} or only 59% of the entire SOC stored within the 7 m profiles was found in the upper 2 m. Considering that most common reports only address the upper 20–40 cm of soil, substantial amounts of SOC have not been yet accounted for in global carbon models. These at least 40% of additional SOC stocks might be even a very conservative estimate, as the used pedotransfer functions of Saxton and Rawls (2006) do not take into account a potential increase in bulk density and hence carbon stocks with soil depths. Nevertheless, such increase was documented in several studies in alluvial plains, e.g. Ricker and Lockaby (2015) found an increase in bulk density of about 10% between the upper meter and the second meter in alluvial profiles with similar soil texture.

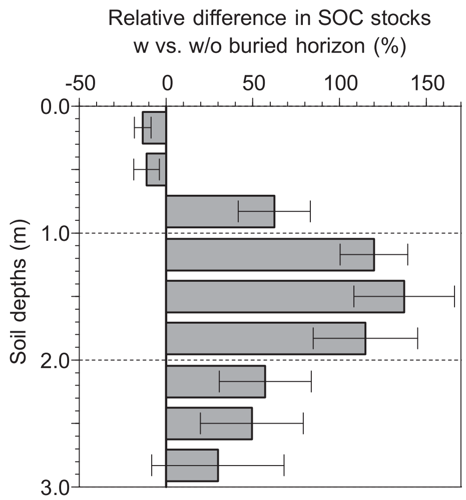

Fig. 5. Soil depth-dependent, relative differences between SOC stocks in profiles with buried horizons and without buried horizons. Error bars indicate the 95%-confidence interval of data from buried horizons.

In general, our estimates of SOC stocks were comparable with studies in New England and the Midwest U.S. (Ricker et al., 2012). For example, Grossman et al. (1998) showed that >50% of the Holocene age C in floodplain soils was deep in the profile, whereas almost 30% of the entire SOC pool in our study is still stored in a depth below 3 m, which verified the concept that alluvial soils contain substantial amounts of SOC at greater depths. That these SOC pools are microbially available in these depths has been shown in other studies (e.g. Gurwick et al., 2008) and at the time of burial, C availability may be determined by the abundance and quality of organic matter. However, further discussion about C availability is beyond the scope of this study. To summarize, our deep soil study of 7 m profiles underlines the importance of natural river floodplains for SOC accumulation during the Holocene.

4.2. Anthropogenic effects on soil organic carbon storage in floodplain soils

Our data clearly indicate that substantial effects of anthropogenic induced sediment mobilisation happened in the headwaters of the Cosumnes River and altered the SOC stocks at most profiles (see buried horizons in Fig. 2). Quantifying this potential SOC gain would call for a larger number of profile data along the entire river to determine buried horizons and burial depths. Such data are not available; nevertheless, our data can be used to estimate the potential increase in SOC storage in case of the accelerated sedimentation that occurred along most rivers of the Central Valley of California since the 1850s. On average of all shallow and deep profiles with buried horizons ($n = 25$) the burial depths of the former A horizon was ranging between 0.66 and 1.33 m. The SOC added on top of the buried horizon represents the gain of SOC due to human-induced sediment accumulating to 114 Mg C ha^{-1}. Additionally, the sedimentation may have conserved SOC due to restrictions in microbial mineralisation in the buried horizon and below. To quantify this effect we compared the mean SOC stocks in depths of 0.8 to 3.0 m in all profiles with buried horizons with the SOC stocks in the upper 2.2 m of the profiles without buried horizons. There was no significant difference (approx. 1 Mg C ha^{-1}) in SOC stocks indicating firstly, the stored SOC was more or less stable within the last 150 years or secondly, the SOC stocks were initially larger under the natural vegetation but some carbon was already mineralized. The latter would be in line with existing studies (Campbell et al., 2000; Engel et al., 2017; Wu et al., 2008) revealing that most of the carbon loss from agricultural soils occurs in the upper meter during the first decade after cultivation (Bruce et al., 1999).

4.3. Anthropogenic effects upon in situ organic carbon dynamics in floodplain soils

In case of our test site, the mean δ^{13}C signature of all deep soil profiles (Fig. 3b) clearly indicated that the arable land use has already altered the δ^{13}C signature in the first meter of the profile. Since corn was an important part of the crop rotation with δ^{13}C values ranging from −6 to −19‰ as a typical C4 plant (Deines, 1979) it influenced the natural ecosystem prior to the period of arable land use. C3 plants with more depleted δ^{13}C values dominated the natural vegetation and therefore, the surface and subsurface soil layers revealed mixed signals ranging between −23 and −25‰ (Fig. 3b). Despite this clear indication that the SOC composition was altered by the shift from natural vegetation to crop land, our data do not allow to quantify the potential loss of SOC due to this land use change occurring in the second half of the 19th century. Such quantification is especially difficult as parallel to land use also soil drainage was affected, indirectly via changes in ground water depths and directly via drainage ditches often installed. However, to balance the impacts of anthropogenic activities upon SOC storage in the floodplains of the Central Valley some estimates are required, how land use change and drainage of the floodplains affect SOC stocks in situ. In the literature, the SOC loss within the upper meter of the soil profile can vary between 0.009 and 0.09 Mg C ha^{-1} yr^{-1} (De Gryze et al., 2010) or range from 0.14 to 1.10 Mg C ha^{-1} over a period of 10 years (Engel et al., 2017). If we consider an estimated SOC loss of 10 Mg C ha^{-1} after 100 years of arable land use, this loss is much smaller than the C gain through the extensive sedimentation following hydraulic gold mining in the headwater catchment (114 Mg C ha^{-1} in the upper 0.8 m of all buried horizons). In consequence, the anthropogenic alteration within catchment and floodplain result in a net C gain of approx. 104 Mg C ha^{-1}, for those areas of the floodplains that exhibit substantial sedimentation. For a full quantification of the anthropogenic impacts on SOC storage in floodplains of the Central Valley more data would be needed about the extent of sedimentation in the floodplains. However, accounting for the data of Jaffe et al. (2007), who estimated that 260×10^6 m^3 of mining debris reached the San Pablo Bay in the northern part of San Francisco Estuary between 1856 and 1887, gives evidence that a substantial area of the floodplains of the Central Valley was affected by a pulse of sediments covering large areas. For a wider study of the entire Central Valley the Cosumnes River floodplains could be a reasonable showcase as they represent the typical situation in many of the other main river floodplains affected by sediment coming from the gold mining areas in the Sierra Nevada. However, it is worth to note that for a full catchment SOC balance taking lateral C fluxes into account, it would be necessary to quantify not only the organic carbon in the floodplain soils but also the lateral SOC loss at erosional sites and their potential in situ dynamic replacement (Berhe et al., 2012; Harden et al., 1999).

4.4. Conclusions

This study gave evidence that human-induced changes of river sediment dynamics and land use within the floodplain have an enormous impact on SOC storage. With the onset of the Californian Gold Rush and the powerful hydraulic mining technique, sedimentation rates increased and deposited huge layers with low SOC contents, but simultaneously deposited over organically rich topsoil which promoted the stabilization of SOC. As this phenomenon is so widespread in the entire Central Valley of California it has a substantial effect on soil organic carbon storage in the entire region and needs to be taken into account to improve our understanding of the human-induced interference with C cycling.

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