A potential shift from a carbon sink to a source in Amazonian peatlands under a changing climate

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Abstract: Amazonian peatlands store a large amount of soil organic carbon (SOC) and its fate under a future changing climate is unknown. Here we use a process-based peatland biogeochemistry model to quantify the carbon accumulation for peatland and non-peatland ecosystems in the Pastaza-Marañon foreland basin (PMFB) in the Peruvian Amazon from 12,000 years before present to 2100 AD. Model simulations indicate that warming accelerates peat SOC loss while increasing precipitation accelerates peat SOC accumulation at millennial time scales. The uncertain parameters and spatial variation of climate are significant sources of uncertainty to modeled peat carbon accumulation. Under warmer and presumably wetter conditions over the 21st century, SOC accumulation rate in the PMFB slows down to 7.9 (4.3~12.2) g $\mathrm{C\ m^{-2}\ yr^{-1}}$ from the current rate of 16.1 (9.1~23.7) g $\mathrm{C\ m^{-2}\ yr^{-1}}$ and the region may turn into a carbon source to the atmosphere at -53.3 (-66.8~-41.2) g $\mathrm{C\ m^{-2}\ yr^{-1}}$ (negative indicates source), depending on the level of warming. Peatland ecosystems show a higher vulnerability than non-peatland ecosystems as indicated by the ratio of their soil carbon density changes (ranging from 3.9 to 5.8). This is primarily due to larger peatlands carbon stocks and more dramatic responses of their aerobic and anaerobic decompositions in comparison with non-peatland ecosystems under future climate conditions. Peatland and non-peatland soils in the PMFB may lose up to 0.4 (0.32~0.52) Pg C by 2100 AD with the largest loss from palm swamp. The carbon-dense Amazonian peatland may switch from a current carbon sink into a source in the 21st century.

Significance Statement: We use a process-based biogeochemistry model to quantify the carbon accumulation for peatland ecosystems in the Pastaza-Marañon foreland basin in the Peruvian Amazon from 12,000 years before present to 2100 AD. We find that warming accelerates peat carbon loss while increasing precipitation slightly enhances peat carbon accumulation at millennial time scales. With these impacts, our simulations suggest that the basin might lose up to 0.4 Pg C by 2100 AD with the largest loss from palm swamp. If this loss rate is true for all Amazonia peatlands, we project that these carbon-dense peatlands may switch from a current carbon sink into a future source in this century.

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Tropical peatlands cover ~441,025 km^2 and store a large quantity (88.6 Pg C) of soil organic carbon (SOC)[1-4]. These ecosystems occupy ~11% of the global peatland area and account for 15-19% of the total global peat SOC stock[3]. Tropical peatlands are mainly distributed in Southeast Asia (~56%, 247,778 km^2), and South and Central America (~23%, 107,486 km^2)[3]. Recently, additional 145,500 km^2 of tropical peatlands containing 30.6 Pg C was discovered in central Congo basin, Africa[5]. Given their significant C stocks, studying their responses to past climatic trends and to the future climate change is of global importance[6-8].

To date, most studies on the role of tropical peatlands in the global C cycle have focused on Indonesian peatlands, which have been acting during the last decades as a considerable C source to the atmosphere resulting from anthropogenic activities (e.g., land exploitation and fires)[9-11]. Few studies have focused on the Amazon basin, where peatlands remain nearly intact, and have been a long-term C sink[7,8,12,13]. The 120,000 km^2 Pastaza-Marañón foreland basin (PMFB) located in Peru is the most extensive peatland complex in the Amazon basin, with up to 7.5 m thick peat deposits. The basal ages vary from 0.67 to 8.9 ka (1 ka=1000 cal years before present) and the peat SOC accumulation rates range from 26 to 195 $g\ C\ m^{-2}\ yr^{-1}$[8,14]. It is a subsiding foreland basin, resulting from the Cenozoic uplift of the Andes Mountains[15-18] and characterized by meandering (more than 100 m in a year[19]) and avulsions of rivers (abrupt changes in the location of river stretches[20, 21]). Waterlogged conditions due to high precipitation and low lying topography provide a favorable environment for peat accumulation[13,14]. By measuring peat characteristics at several peatland sites within the basin, and using Landsat TM images, Lähteenoja[14] estimated a peatland area of 21,929 km^2 with SOC stock of 3.116 Pg (0.837-9.461 Pg) for the central parts of the PMFB. Further, by incorporating multi-sensor remote sensing and adding more peat core data, Draper[13] mapped the distribution of peatland and non-peatland ecosystems in the PMFB and estimated a peatland area of 35,600$\pm$2133 km^2 with 3.14 Pg C (0.44−8.15 Pg) stored in the vegetation and peat deposits of the whole basin.

According to most climate models, mean air temperature of South America has been projected to increase by 1.8-5.1℃ for the PMFB by the end of this century[22-25]. Annual precipitation is projected to increase by up to 500 mm, although a large uncertainty exists[22-25]. The strong dependence of C dynamics on climate suggests that warming in the 21st century may turn the peatlands in the PMFB from a long-term C sink into a source[6,7,9,26]. However, this potential change has not been quantified or modelled in any way in previous studies. Nearly all models focusing on the future C dynamics of the Amazon basin have been applied to non-peatland ecosystems[27-35] with the exception of[29].

Process-based models offer an alternative approach to quantifying peatland C dynamics and providing insights for future projection[36-40]. Recently, a peatland terrestrial ecosystem model (P-TEM) was developed for both peatland and non-peatland ecosystems by combining a hydrology module (HM), a soil thermal module (STM), a methane dynamics module (MDM), and a C and nitrogen dynamics module (CNDM)[39]. P-TEM has been evaluated and used for estimating C stocks across the Alaskan landscape since the last deglaciation[39,40]. Here, we parameterize and evaluate the P-TEM for tropical peatlands and model the C dynamics of the peatlands in the Pastaza-Marañón foreland basin, Peruvian Amazonia (Fig 1) from 9 ka to 2014

AD. The model parameters were optimized using published peat, vegetation and remote-sensing data for the PMFB from[13,14] as well as other published sources (SI Appendix, Tables S1 and S2). The model was then used to 1) quantify past C accumulation from 12 ka to 2014 AD in peatlands, and 2) predict the future trends of C accumulation under different climate scenarios in the 21st century in peatland and non-peatland ecosystems within the PMFB.

Past C Accumulation

The annual comparison between model simulation and measurements[14] at a temporal resolution of a year at the thickest and largest Amazonian peatland site (Aucayacu site) reveals that our model captures the historic peat SOC accumulation rates (Fig 2a) and the peat depth profile (Fig 2b) for most simulation periods, but overestimates the rates between 8 and 6 ka. Simulated total depth reaches 8 m (ranging 6-12 m), slightly higher than the measured 7.5 m[14]. The correlation between simulations and measurements using 500-year bins at multiple sites with different vegetation types indicates that the model well estimates SOC accumulation trajectories at millennial time scales (SI Appendix, Fig S1). The model underestimates the rates between 3 and 2 ka at San Jorge, and between 0.5 and 0 ka at Charo. Although the model underestimates the rates between 2 and 1.5 ka at Rinon (an open peatland site), the starting basal age for the regional transient simulation for the open peatlands is at 1.6 ka (see SI Appendix, Table S3). As indicated by the mean basal age applied in the model, pole forest (PF) has a longer SOC accumulation period than palm swamp (PS) and open peatland (OP) in general (SI Appendix, Fig S1), with peat initiation of PF around 2000 years ahead of PS and OP peat initiation.

Our simulation suggests there were strong relationships between peat C dynamics and climatic change. Temporally, temperature and photosynthetically active radiation (PAR) rose slightly over the whole period (SI Appendix, Figs S2a and e), whereas annual precipitation decreased before 4 ka and subsequently increased (SI Appendix, Fig S2c). Under increasingly warmer and drier conditions before 4 ka, the historic SOC accumulation rate declined at the Aucayacu site (Fig 2a). It started to increase concurrently with the wetter conditions after 3.5 ka (close to 4 ka), despite continuing warming. Overall, the historic SOC accumulation rates of the Aucayacu peat core followed the historic pattern of the precipitation change (SI Appendix, Fig S2c). This suggests that higher rainfall might have accelerated while warming and drought might have decelerated peat SOC accumulation at millennial time scales.

Spatially, this relationship between the peat SOC accumulation and climate was indicated by the patterns of the peat SOC densities distribution and mean historic temperature and precipitation within the PMFB (Figs. 3 and S3). We find that the highest peat SOC density region fell in the northeast with the highest precipitation and relatively low temperature. The secondary highest peat SOC region was located in the northwest with moderate precipitation but the lowest temperature. The lowest peat SOC zone fell within the southwest where the lowest precipitation and highest temperature coincided. This, again, suggests that higher precipitation increased whereas higher temperature reduced peat SOC accumulation at regional scales.

The climatic effects on the long-term peat SOC accumulation in the PMFB can be explained by our simulated C fluxes and hydrological factors. Peat accumulated SOC where the rate of soil C input was higher than the decomposition[41]. Soil C input from litters was largely controlled by and was proportional to plant net primary productivity (NPP). Soil decomposition was modeled as heterotrophic respiration (R_H)[42] (see SI Appendix). Increasing temperature and PAR stimulated the plant C uptake by increasing NPP. However, warming might have created favorable conditions for microbial decomposition[43]. Warming also increased the evapotranspiration, decreasing water table, thereby reducing anaerobic respiration and increasing aerobic respiration[44]. Increasing precipitation had a positive effect on NPP. It also lifted the water table and decreased R_H. This, in turn, enhanced peat C accumulation. In our previous study for the northern (Alaskan) peatlands[40], under the warmer conditions, the stimulation of NPP exceeded the stimulation of R_H, thereby increasing SOC accumulation in northern peatlands during the Holocene Thermal Maximum (HTM). Similarly, we find that R_H within 1 m depth followed the increasing trend of temperature with a decrease at 4 ka when precipitation increased (SI Appendix, Fig S4b). This suggests that warmer condition in the PMFB enhances R_H while wetter condition decreases R_H. The volumetric soil moisture (VSM) (SI Appendix, Fig S4c) and water table (SI Appendix, Fig S4d) started decreasing at 8 ka as precipitation became lower. At the same time, R_H kept increasing under such drier condition. Interestingly, when climate became wetter at ~4 ka (SI Appendix, Fig S2c), the VSM abruptly increased. The water table also stopped dropping and showed an increase pattern. Meanwhile, R_H started decreasing. This again suggests that higher precipitation may decrease R_H and thus slows the peat SOC decomposition by increasing the soil moisture and raising the water table. As warming continued, the increase of VSM and water table were slight, presumably due to the enhanced evapotranspiration.

To examine how temperature and precipitation have impacted NPP in this region. The attributions of these two key drivers and soil water content to NPP for both historical periods and the 21st century were analyzed with the Analysis of Variance table (ANOVA) and the F-test of the multi-variate linear regression between annual mean NPP and climate variables. For the historical simulation at Aucayacu site, we find that, although higher precipitation and higher temperature increase the NPP (SI Appendix, Table S4 and S5), those two factors have limited effects presumably because the temperature exceeds the optimum temperature for photosynthesis and soil water content is already suitable for plant growth. The variable with the highest importance is VSM, indicating the hydrological condition plays the most important role in determining the NPP. Such hydrological condition is modeled by various factors including the temperature, precipitation, solar radiation and others such as soil porosity, soil layers characteristics that are described in our previous hydrological modeling studies[39,60].

Our historical simulations at Aucayacu and in the PMFB suggest that NPP was consistent with the temporal patterns of precipitation and VSM (SI Appendix, Fig S2c). The spatial correlations between NPP, vegetation C density, and mean historic precipitation were detected (SI Appendix, Figs S3b and S5) when observing each peatland vegetation type separately. Pixels with higher vegetation C density and NPP fell within the northeastern wetter region while lower vegetation C density and NPP pixels were in the southwestern drier region. Our explanation is that during the historical period, the point where NPP will no longer positively respond to the increasing precipitation and VSM has not been reached. Still, the wetter condition stimulates the SOC accumulation by increasing NPP and decreasing R_H. However, for the simulations under three future climate scenarios (SI Appendix, Table S6), we find that the F values of the precipitation started decreasing as the precipitation continued to be higher from RCP 2.6 to 8.5. This suggests that precipitation becomes less and less important in the future for NPP as it increases, based on the existing suitable hydrological condition.

In our model, GPP is a function of atmospheric CO_2 concentrations in addition to physical variables (see SI Appendix). The CO_2 effects is modeled with a Michaelis-Menten equation considering CO_2 concentrations inside leaves which is assumed to be directly proportional to atmospheric CO_2 concentrations when stomata are fully open. When moisture is a limiting factor, the limitation on CO_2 assimilation is modeled by the modifying the conductance of leaves to CO_2 diffusion. The moisture availability is expressed as the ratio of actual evapotranspiration (EET) to potential evapotranspiration (PET). The relationship between CO_2 concentration inside stomatal cavities (C_i) and in the atmosphere (C_a) is proportional to relative moisture availability:

$$G_V = 0.1 + (\frac{0.9EET}{PET})$$

$$C_i = G_V C_a$$

where G_V is a unitless multiplier that accounts for changes in leaf conductivity to CO_2 resulting from changes in moisture availability. When there is sufficient water in soils, EET will not be limited by water, which will reach its maximum value, GV is close to 1. This suggests that inside of leaves CO_2 will be close to ambient CO_2. When the ecosystem has sufficient precipitation, GPP and NPP will not respond to increasing precipitation.

At northern high latitudes, in addition to CO_2 fertilization effects, warming also enhances photosynthesis, stimulating plant productivity (NPP) and thus increasing SOC accumulation[42,5-47]. In contrast, warming in the tropical regions generally led to temperatures above the optimum level for photosynthesis[48,49], which is also suggested by the ANOVA analysis (SI Appendix, Tables S4 and S6) as increasing temperature in the future has less and less positive effects on NPP. Increasing temperature accelerates R_H, however, at the same time. The less sensitivity of NPP versus R_H to warming might ultimately result in the SOC loss in the PMFB under warmer conditions.

However, hydrology, NPP, and SOC accumulation can also be controlled by autogenic processes of peatlands such as transition from minerotrophic to ombrotrophic conditions[8]. This transition is largely induced by the form and thickness of the peat deposit and less affected by prevailing climatic conditions – as long as the rainfall is sufficient to sustain an rain-fed bog. Interestingly, in the Aucayacu peatland, the transition from minerotrophic to ombrotrophic conditions occurred around 4 ka (3.5 ka) – exactly when precipitation started to increase. It might have been a coincidence, but it is also possible that the increased precipitation enabled the appearance of ombrotrophic bogs. If this is the case, a change in the precipitation did not affect the NPP directly but indirectly by inducing a change in the peatland type. Since our model cannot simulate the paleo-ecological change including the shifts between different peatland ecosystem types through time, our results may only partly explain the observed patterns, with much information still relying on paleo-ecological studies[14,50-52]. The relationship between NPP and precipitation for peatland ecosystems in the region should be further studied in the future.

Another key control of the current distribution of peat depths and SOC densities within the PMFB is the active lateral migration of rivers[14]. The current distribution of peat SOC densities can be explained by both climatic and geological factors. The Amazon river networks can be affected under future climate conditions, which will affect peatland dynamics (e.g., formation and area change). Further, our model did not differentiate the minerotrophic vs. ombrotrophic conditions for the peatland ecosystems, which will introduce biases. Incorporating these dynamics into future analysis shall improve our predictions of SOC for this region.

Our uncertainty analysis suggests that the uncertainty of the simulated past C accumulation rates was mainly due to parameters, spatial variations of climate variables (SI Appendix, Fig S3), and the uncertain peat basal ages (SI Appendix, Table S3). Specifically, using the mean peat basal age by averaging the basal ages of peat samples for each peatland type is a top uncertainty source. The variation of peat characteristics (e.g., bulk density, C content, peat depth) and limited number of samples are also sources of the uncertainty.

Current C stocks

Overall, model simulations of current peatland C stocks are comparable to the field measurements of[13]. Specifically, PF has the SOC density of 1900 $\mathrm{Mg\ C\ ha^{-1}}$, consistent with the field measurements (800-2200 $\mathrm{Mg\ C\ ha^{-1}}$, SI Appendix, Fig S6a)[13]. PS has the next highest SOC density (1100 $\mathrm{Mg\ C\ ha^{-1}}$), which is sufficiently within the measured range of 300-1390 $\mathrm{Mg\ C\ ha^{-1}}$. The SOC density of OP is 535 $\mathrm{Mg\ C\ ha^{-1}}$, also within the measured 392-1492 $\mathrm{Mg\ C\ ha^{-1}}$. The high SOC density of PF corresponds to the longer SOC accumulation period compared to the other types (SI Appendix, Figs S1a, b and S6a). Our simulations are even closer to the field measurements[13] when vegetation C density was examined (SI Appendix, Fig S6b). The simulated lowest vegetation C density was in PF (86 $\mathrm{Mg\ C\ ha^{-1}}$) and within the measured range of 80-100 $\mathrm{Mg\ C\ ha^{-1}}$.

The model estimates a total SOC of 3.922 (2.208~5.777) Pg in the PMFB including 3.519 (1.833~5.344) Pg in the peatland soils, which is higher than the measured total peat SOC, 2.844 Pg (SI Appendix, Table S7)[13]. The simulated vegetation C stock of 1.104 (1.097~1.137) Pg with 0.34 (0.338~0.369) Pg on the PMFB peatlands is also higher than the measured value (0.293 Pg C)[13]. Our model may overestimate the soil and vegetation C stocks. The uncertainty of the simulated C stocks is mainly due to the spatial variations of the interpolated mean temperature (25-29℃) and precipitation (2200-2900 mm) (SI Appendix, Fig S3).

Future Projection

Under the RCP 2.6 scenario (see below for climate description), the SOC accumulation rate in all ecosystem types within the PMFB decreases from 16 (9~24) to 7.9 (4.3~12.2) $\mathrm{g\ C\ m^{-2}\ yr^{-1}}$ and the SOC accumulation rate in the peatlands dramatically decreases from 56

(29~85) to 23 (15~32) g $C\ m^{-2}\ yr^{-1}$ (Table 1). The PS exhibits the biggest drop from 65 to 26 g $C\ m^{-2}\ yr^{-1}$. Spatially, the majority of pixels within the PMFB have positive SOC accumulation and vegetation C change, but some areas with PS have SOC loss (Figs 4 and S7). Overall, 0.067 (0.037~0.108) Pg SOC, including 0.06 (0.03~0.1) Pg SOC in the peatlands, will be accumulated in the PMFB by the end of the 21st century under moderately warmer and wetter conditions of this climate scenario (Table 1). There will be 0.0148 Pg C accumulated in vegetation, including 0.0048 Pg C in peatland vegetation.

Under the RCP 8.5 scenario, the SOC accumulation rate declines from 16 to -53 (-67~-41) $C\ m^{-2}\ yr^{-1}$ and the rate in peatlands declines from 56 to -123 (-152~-91) $C\ m^{-2}\ yr^{-1}$ (Table 1). Again, the highest decline of the rate is for PS, from 65 to -135 g $C\ m^{-2}\ yr^{-1}$. The pixels with SOC and vegetation C loss dominate the region (Figs 4 and S7). Under this climate scenario, the PMFB will act as a C source of 0.413 (0.319~0.518) Pg C by 2100 AD (Table 1). Peatlands will lose 0.31 (0.23~0.38) Pg C compared with 0.1 Pg C loss from non-peatland ecosystems. Vegetation will lose 0.07 Pg C, including 0.02 Pg C from peatland vegetation. Among all peatland ecosystem types, PS could be severely affected by the climate due to its large area within the PMFB and within the whole Amazon Basin[53]. It must be taken into account that the tendency of the model to overestimate the current soil and vegetation C stocks in the PMFB (SI Appendix, Table S7) might affect these values to some extent.

Under the intermediate RCP 4.5 scenario, the SOC accumulation rate declines from 16 to -19 $C\ m^{-2}\ yr^{-1}$ and the SOC accumulation rate in peatlands declines from 56 to -45 $C\ m^{-2}\ yr^{-1}$. Peatlands will lose 0.12 Pg C compared with 0.034 Pg C from non-peatlands.

Three extra simulations were conducted as sensitivity tests to examine the effects of potential drier climate in the PFMB on SOC accumulation rates. We assume that (1) The future precipitation will decrease 5% in our study region over the century, but holding air temperature change as in the original RCP 2.6; (2) The future precipitation will decrease 10% but holding air temperature change as in RCP 4.5; (3) The future precipitation will decrease 15% but holding air temperature change as in RCP 8.5. The precipitation was manually decreased at monthly step for each grid cell from 2014 to 2100 AD to achieve the certain percentage total reduction at the end of 2100 AD. Our simulations show that the C accumulations are +0.027 (0.02~0.068), -0.203 (-0.349~-0.167), and -0.594 (-0.731~-0.51) Pg C under the three sensitivity simulations (Table 1). These extra simulations suggest that the slightly drier condition will decrease but will not have significant effects on the C accumulation in this region.

The modeled current C stocks agree with the field observations at the Aucayacu site, which is a PF site. However, instead of PF, PS is the dominant peatland type in the study area and its SOC accumulation rates at Charo site is underestimated. Thus, using PS as representative peatland types for regional simulations under future climate scenarios may evolve uncertainty due to the underestimation during the parameterization.

In conclusion, the warming in the 21st century may weaken the C sink function of the Amazonian peatlands in the PMFB or may entirely switch them from a long-term carbon sink into a source, depending on the severity of the warming. The same has also been predicted for the Amazonian rainforest in general[26]. The vegetation and SOC density changes (future total C stock changes (Table 1) divided by the corresponding areas (SI Appendix, Table S7) of peatlands and non-peatland) were calculated to compare with other studies. Our model estimation of vegetation C change for the non-peatland (mainly flooded forest) ecosystem in the 21st century (+0.23~-1.17 kg C m^{-2}) is well within the range of other studies (+0.6~-1.2 kg C m^{-2}) on the future vegetation C change from Amazonian rainforest dieback[26,54] (SI Appendix, Table S8). Our estimation of SOC change for the non-peatland ecosystem in the 21st century (+0.18~-3.35 kg C m^{-2}) is also comparable to -3.88 kg C m^{-2} from other studies[26,54]. Further, we find that the ratio of SOC density changes for peatlands and non-peatland ecosystems in the next 100 years ranges from 3.9 to 5.8 (SI Appendix Table S8). This indicates that future warming is likely to affect the Amazonian peatlands more dramatically than non-peatland ecosystems, although the total area of peatlands is much smaller than that of non-peatland ecosystems within the PMFB (31000 km^2 vs. 47000 km^2, see SI Appendix for areas). The high vulnerability of peatland ecosystems to future climate is presumably due to its large amount of existing SOC stock (3.5 Pg C) compared with non-peatland ecosystems (0.4 Pg C). Another possible reason is that, in addition to the non-linear function defining the volumetric soil moisture (VSM) effect on heterotrophic respiration within the unsaturated zone, there is also a linear relation between water-table depth and aerobic respiration in the model (see SI Appendix for decomposition calculation). Future warming increases evapotranspiration that subsequently decreases VSM and increases aerobic respiration for both peatland and non-peatland ecosystems. For peatlands, it further lowers the water table which, in turn, increases aerobic respiration. This suggests that peatland ecosystems may suffer larger SOC decomposition under the changing climate and may help explain its vulnerability. In addition, increased land use change, expansion of commercial agriculture, transport infrastructure, and hydropower development form a threat to the persistence of the considerable C stock[55]. The most carbon dense ecosystems of the whole Amazon basin may turn into C sources.

Methods

The peatland biogeochemistry model and its parameterization

In P-TEM, peat SOC accumulation is determined by the difference between NPP and aerobic and anaerobic respiration R_H at a monthly step (see[1,2] and SI Appendix for model details). Parameters in P-TEM were first optimized with data of annual C fluxes and stocks in the Amazon basin taken from literature (SI Appendix, Table S1) in order to obtain the prior distribution of the parameter space for peatland ecosystems (see SI Appendix for details) Specifically, site-level measurements of tree biomass from Amazonian peatlands[13] were used to compare with model simulations to optimize parameters. Due to the lack of NPP measurements, NPP values used in the model are field measurements from neighboring white-sand forests (for pole forest peatlands) and seasonally flooded forests (for palm swamp peatlands and flooded forests) (SI Appendix, Table S1). Second, a Bayesian approach was used to optimize parameters (SI Appendix, Table S2) with Monte Carlo ensemble simulations driven by the extracted paleo climate data (SI Appendix, Fig S2) at five peatland sites (SI Appendix, Fig S1).

The distribution of vegetation types was taken from[13] at a resolution of 90 m×90 m and was resized to 1.69 km×1.69 km. Vegetation types in the region include three peat-forming vegetation types (pole forest (PF), palm swamp (PS), and open peatland (OP, peatland lacking closed canopy)) and a non-peat forming type (flooded forest (FF)) (Fig 1). OP was assumed to have minimal NPP and vegetation biomass during the simulation[13].

Climate Data

The climate forcing data for historic simulations include temperature, precipitation, photosynthetically active radiation (PAR), vapor pressure at a monthly step and CO_2 at an annual step from 12 ka to 1990 AD, simulated by CCSM3 (TraCE-21ka) at a spatial resolution of 3.75° ×3.75°. Climate forcing data for modern simulations is from Climate Research Unit (CRU2.0) at a monthly step from 1990 to 2014 AD at a resolution of 0.5° ×0.5°. For future simulations, we applied the Representative Concentration Pathway (RCP) 2.6 (mean annual temperature in the PMFB has the smallest increase (by ~0.5 ℃), mean annual precipitation increases by ~260 mm, and CO_2 increases by ~80 ppm at 2050 AD and decreases by ~30 ppm at 2100 AD), RCP 4.5 (by ~1.5 ℃, ~290 mm, and CO_2 increases by ~150 ppm at 2100 AD), and RCP 8.5 (by ~2.7 ℃, ~350 mm, and ~600 ppm at 2100 AD) at a monthly step from 2014 to 2100 AD at 0.5° ×0.5° as possible future climate scenarios. The CRU data together with the modern digital elevation data at 1.69 km×1.69 km were input into interpolation software ANUSPLIN4.4. We then downscaled the paleo-climate data (TraCE-21ka, 3.75° ×3.75°) and the RCP data (0.5° ×0.5°) based on the spatial variations of the interpolated CRU data (1.69 km×1.69 km) by assuming that the spatial variations of CRU to be the same as that of paleo and RCP data.

Model Application and Uncertainty Analysis

A 500-year run was conducted for each peatland ecosystem type ahead of the basal age using parameters of non peat-forming FF to determine the initial SOC within the upper 1 m mineral soil underlying the peat deposit. The model was first run from 12 ka to 2014 AD for validation at five peatland sites (SI Appendix, Fig S1). The simulated SOC accumulation rates of PS, OP, and PF were firstly compared with measured SOC accumulation rates[7,14] annually in 500-year bins. Second, we applied the model to a regional simulation with interpolated monthly paleo-climate data for the PMFB. We averaged all the measured basal ages of each peatland type to determine the mean basal age of each peatland ecosystem type (SI Appendix, Table S3). Basal ages at sites where they have not been measured were calculated using mean SOC accumulation rates, bulk density, peat depth and C content of each peatland ecosystem type derived from[7,13,14], following the equations in[7]. We conducted the simulation from 1 ka to 2014 AD for FF. Finally, we conducted the simulations for future projection using the interpolated RCP 2.6, RCP 4.5 and RCP 8.5. A series of simulations were conducted to examine the effects of the spatial variation of climate and the posterior distribution of the parameter on the estimated C accumulation rates and stocks (see SI Appendix for details). Twenty sets of parameters were randomly drawn from the posterior parameter space. All pixels in the study area were then assigned with the same climate forcing data which were random combinations between temperature and precipitation.

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Figure Legends

Fig 1. Distribution of peat- and non peat-forming vegetation in the PMFB at the resolution of 90 m×90 m[13]. The map was resized to 1.69 km×1.69 km. Colors represent vegetation types: open peatland (pink), palm swamp (red), pole forest (green), and flooded forest (dark blue). Yellow represents open water and light blue represents other. See the Figs 1 and 4 of[13] for original map.

Fig 2. Comparison between simulation and measurement[14] at the Aucayacu site. (a) SOC accumulation rates and (b) peat depth. Shaded areas represent the range due to uncertainties from the posterior distributions of the parameters after the parameterization. 0 cm at ~9 ka indicates no peat accumulation.

Fig 3. Current (2014 AD) SOC density of flooded forest, palm swamp, open peatland, pole forest and their combination in the PMFB.

Fig 4. Changes of SOC density from 2014 to 2100 AD under RCP 2.6 and RCP 8.5 future climate scenarios in flooded forest, palm swamp, pole forest, open peatland, and their combination in the PMFB. Blue and green represent the SOC accumulation. Yellow and red represent the SOC loss.

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