



## Field-warmed soil carbon changes imply high 21<sup>st</sup> century modeled uncertainty

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**Abstract.** The feedback between planetary warming and soil carbon loss has been the focus of considerable scientific attention in recent decades, due to its potential to accelerate anthropogenic climate change. The soil carbon temperature sensitivity is traditionally estimated from short-term respiration measurements -- either from laboratory incubations that are artificially  
10 manipulated, or field measurements that cannot distinguish between plant and microbial respiration. To address these limitations of previous approaches, we developed a new method to estimate temperature sensitivity ( $Q_{10}$ ) of soil carbon directly from warming-induced changes in soil carbon stocks measured in 36 field experiments across the world. Variations in warming magnitude and control organic carbon percentage explained much of field-warmed organic carbon percentage ( $R^2=0.96$ ), revealing  $Q_{10}$  across sites of 2.2, [1.6, 2.7] 95% Confidence Interval (CI). When these field-derived  $Q_{10}$  values were  
15 extrapolated over the 21<sup>st</sup> century using a post-hoc correction of 20 CMIP5 Earth system model outputs, the multi-model mean soil carbon stock changes shifted from the previous value of  $83 \pm 156$  Pg-carbon (weighted mean  $\pm 1$  SD), to  $16 \pm 156$  Pg-carbon with a  $Q_{10}$  driven 95% CI of  $245 \pm 194$  to  $-99 \pm 208$  Pg-carbon. On average, incorporating the field-derived  $Q_{10}$  values into Earth system model simulations led to reductions in the projected amount of carbon sequestered in the soil over the 21<sup>st</sup> century. However, the considerable parameter uncertainty led to extremely high variability in soil carbon stock projections within each  
20 model; intra-model uncertainty driven by the measured  $Q_{10}$  was as great as that between model variation. This study demonstrates that data integration may not improve model certainty, but instead should strive to capture the variation of the system as well as mean trends.

### 1 Introduction

The flux of carbon dioxide between the soil and atmosphere is a major control on atmospheric carbon dioxide concentrations.  
25 Warming temperatures, driven by increases in atmospheric carbon dioxide, have the potential to simulate carbon decomposition, accelerating its release into the atmosphere (Davidson and Janssens, 2006). If this is not counterbalanced by an equivalent increase in primary productivity (the opposing carbon flux) then it has the potential to drive a land carbon-climate feedback that will accelerate anthropogenic climate change. Recent global compilations of data from ecosystem warming experiments lend support to this idea (Carey et al., 2016), suggesting that warming alone could drive a loss of carbon



from the upper soil horizons (Crowther et al., 2016). However, these studies addressed the impact of warming in isolation, and it remains unclear how this process will interact with the variety of other global change drivers to affect the global soil carbon stock over the rest of this century. Reflective of such uncertainty, soil carbon changes projected for 2100 under business-as-usual scenario for Coupled Model Intercomparison Project Phase 5 (CMIP5) range from -70 to 250 Pg-carbon across different Earth system models (Todd-Brown et al., 2014), making the land-carbon feedback one of the largest sources of uncertainty in future climate projections (Friedlingstein et al., 2014). Improving the soil carbon component of the Earth system models is essential to predicting the future evolution of the Earth system and thus establishing meaningful greenhouse gas emissions targets.

A fundamental parameter describing soil temperature-sensitivity in soil carbon models is the  $Q_{10}$  – the factor of the change in decomposition rate associated with 10°C of warming from a reference temperature (Davidson and Janssens, 2006; Lloyd and Taylor, 1994). Traditional laboratory incubations have found a wide range of  $Q_{10}$  values, varying from 1.4 (Townsend et al., 1997) to > 3 (Davidson et al., 1998, 2006) with 2 being the most commonly accepted value. Complicating this, theoretical analyses based on chemical kinetics suggest  $Q_{10}$  is itself dependent on temperature (Davidson and Janssens, 2006; Lloyd and Taylor, 1994), though these values are typically very close to 2 in most environmental temperature ranges (Lloyd and Taylor, 1994). More recently, large-scale analyses of field respiration converge on  $Q_{10}$  estimates of 1.4 to 1.5 (Bond-Lamberty and Thomson, 2010; Hashimoto et al., 2015; Mahecha et al., 2010). Unsurprisingly, this temperature response is also critical in Earth system models, where the temperature sensitivity parameter is known to be a major driver of variation (Booth et al., 2012; Jones and Cox, 2001; Jones et al., 2006). However, it is unclear what is driving the lower  $Q_{10}$  estimates in these field-based syntheses compared to the average lab-based estimates from single-site studies, and there appears to be a relatively wide range of ‘typical’  $Q_{10}$  values in the literature. Nevertheless, most Earth system models use values that range from 1.5 (Oleson et al., 2013; Raddatz et al., 2007) to 2 (Bonan, 1996; Cox, 2001).

Traditionally, these  $Q_{10}$  values have been calculated from warming-induced changes in soil respiration rates. However, this approach has two main limitations: 1) respiration rates measured under idealized laboratory conditions fail to reflect the structure, heterogeneity and variability of natural systems, whereas 2) field measurements cannot directly isolate heterotrophic soil respiration from autotrophic root respiration without substantially altering the system. Estimating  $Q_{10}$  directly from warming-induced changes in soil carbon stocks could be a valuable approach to address these limitations, but the variability and relative imprecision of soil carbon stock data necessitates a large sample size to adequately describe variation at the global scale (Bradford et al., 2016). Yet, results from a recent Earth system model meta-analysis indirectly suggests that, with enough sample coverage it may be possible to infer  $Q_{10}$  directly from changes in soil carbon stocks (Todd-Brown et al., 2014).

Here we present a new approach to estimate the global  $Q_{10}$  value from net changes in soil carbon stocks under warming, rather than soil respiration measurements, and examine the consequences of these estimates –with associated uncertainty– on CMIP5 Earth system model projections of global carbon storage over the rest of the 21<sup>st</sup> century. To do this, we use a global database of soil carbon stock data from 36 field-warming experiments around the world, each of which includes control (ambient) plots, and those which have been warmed for extended (years to decades) periods of time (Crowther et al., 2016) (Tables S11) and



outputs from 20 CMIP5 (Taylor et al., 2011) Earth system models RCP 8.5 business-as-usual experiment (Tables 1 and SI3). These field data were used previously to derive Earth system model independent estimates of global soil carbon temperature sensitivity where the effect of warming was isolated from other global change drivers or the interacting climate system (Crowther et al., 2016). In this study we develop a novel approach that enables us to explore these field results in the context of the temperature sensitivity function ( $Q_{10}$ ) used in integrated Earth system model. We then examine the consequences of the data-driven  $Q_{10}$  estimates, and the associated uncertainty, for CMIP5 Earth system model projections of global carbon storage over the rest of the 21<sup>st</sup> century using a novel post-hoc modification of the CMIP5 simulation outputs.

## 2 Materials and Methods

### 2.1 Field sites

The field sites were drawn from a previous analysis (Crowther et al., 2016). From this initial database of 48 paired case-control studies, we selected 36 studies that were run longer than 2 years to match the metastable state assumption articulated below. 18 of these sites were temperate grasslands, savannas, and shrublands, 10 temperate broadleaf and mixed forests, 6 tundra, 1 boreal forests or taiga, and 1 site was in a Mediterranean forest, woodland and scrub. A traditional statistical analysis of the sites is provided by Crowther *et al.* (2016). For this study, we used the increase in soil temperature due to warming, length of the study, and the percent of soil organic carbon in paired warmed and control plots (Table SII).

### 2.2 $Q_{10}$ calculations

We calculated traditional  $Q_{10}$  estimates based on these warming-induced soil carbon losses, enabling us to embed this temperature sensitivity information into a soil decomposition model framework. Traditional soil decomposition models follow a first order linear decay framework where:

$$\frac{dC(t)}{dt} = u_{in}(t)\mathbf{b} - \mathbf{C}(t)\mathbf{K}\mathbf{Q}_{10}(T, t)\mathbf{A}, \quad (1)$$

where the  $\mathbf{C}$  is a vector of soil carbon pools with unique turnover times,  $t$  time,  $u_{in}$  a scalar of soil carbon inputs,  $\mathbf{b}$  an allocation vector describing how the inputs are divided between the soil carbon pools,  $\mathbf{K}$  is a diagonal matrix representing the decomposition rates of the pools,  $\mathbf{Q}_{10}$  is a diagonal matrix with entries of the form  $q_i^{(T(t)-T_0)/10}$  representing the temperature sensitivity factor,  $T$  a scalar describing the soil temperature and  $T_0$  an arbitrary reference temperature, and  $\mathbf{A}$  the transfer matrix representing movement of carbon between soil carbon pools.

The temperature sensitivity was assumed to be constant across pools. This allows us to collapse the diagonal  $\mathbf{Q}_{10}$  matrix to be collapsed into a single scalar value of the form  $Q_{10}^{(T(t)-T_0)/10}$ . This constant temperature sensitivity assumption is discussed below and follows the structure of the CMIP5 Earth system models.

In general, there are three classes of pool structure for traditional models: *independent* where there was no exchange between soil carbon pools making  $\mathbf{A}$  the identity matrix, *cascade* where pools with faster turnover times passed carbon to pools with



slower turnover times making  $A$  a lower triangular matrix, and fully *feedback* models where carbon was exchanged between faster and slower pools and vis-versa making  $A$  a fully dense matrix. In all cases  $KA$  is an M-matrix, implying there exists an inverse whose entries are all positive. For the independent and cascade pools  $KA$  is diagonalizable, implying it can be broken down into a diagonal matrix  $D$  and an invertible matrix  $P$  such that  $KA = P^{-1}DP$ .

- 5 For most well-developed soils, soil carbon stocks are at a metastable state where soil inputs approximately equal outputs (see Results for discussion of Earth system model outputs). Given that  $KA$  is an M-matrix and this metastable state approximation, we can describe the total soil organic carbon as follows:

$$C = \frac{u_{in}}{kQ_{10}^{(T-T_0)/10}}, \quad (2)$$

- where  $C$  is the total organic carbon stock,  $u$  the sum of the soil inputs, and  $k$  a bulk decomposition rate that can be constructed from the decay matrix  $KA$  and allocation vector of the soil inputs  $b$ . For details see of this derivation the SI: Mathematical Analysis.

We can now describe the soil carbon stock difference between two soils with the same decay rate but different temperatures and inputs. This could either be two time points from a simulation where the soil output is close (within 10%) of the soil inputs, or a warmed treatment and a control:

$$15 \quad C_2 = C_1 \left( \frac{u_2}{u_1} Q_{10}^{(T_1-T_2)/10} \right). \quad (3)$$

For the field sites, we assume that the relative change in inputs due to warming is negligible compared to the effect on the decomposition rate across sites and that the main driver of differences in decomposition rates between control and treatment is the warming treatment. Leading us to:

$$C_w = C_c Q_{10}^{-\Delta T/10}. \quad (4)$$

- 20 Finally, we assume that the bulk density of the soil at a given site was unaffected by the warming treatment. This allows us to use the mass percent soil organic carbon instead of the soil organic carbon density for Eq. 4.

### 2.3 Model-data integration: parameter fitting

- Given the relatively small parameter space, we choose a brute-force model-data integration approach where we iteratively calculated the predicted change in soil carbon stock given the control soil carbon (Eq. 4) across a range of  $Q_{10}$  values from 0.1 to 5 in 0.1 increments. Data-model fits were scored using root mean squared error (RMSE) and linear regression ( $R^2$ , slope and intercept).  $Q_{10}$  values were selected on low bias (slopes and intercepts within 2 standard deviations of 1 and 0 respectively) due to the relative insensitivity of the  $R^2$  and RMSE metrics (see Figure 1). To test for statistical power, we randomly sampled the data 1000 times with sample sizes from 5 to 34 sites and compared this to samples with randomly assigned control vs warming (for each study the percent carbon of control and treatment has a 50% chance of being switched). These random and sample generated  $Q_{10}$  distributions, were compared using a two-sample Kolmogorov-Smirnov test to test that the distributions were statistical distinct.



## 2.4 Earth system model analysis

Earth system model simulations were drawn from CMIP5, the Coupled Model Intercomparison Project to support the 5<sup>th</sup> IPCC assessment report (Taylor et al., 2011). We downloaded simulation outputs from the RCP 8.5 scenario, representing the ‘business as usual’ scenario, including heterotrophic respiration ( $rh$ ), soil temperature ( $tsl$ ), and heterotrophic carbon stock ( $cSoil$ ,  $cLitter$ , and  $cCwd$ ) from the CMIP5 repository on the Earth System Federation Grid. Ten-year means were taken at the beginning and end of the 21<sup>st</sup> century for each variable (corresponding to 2006-2015 and 2090-2099). Soil temperature was averaged for the first 10 cm to correspond with experimental surface temperature readings. Soil carbon stock was calculated by adding all heterotrophic-respiring pools (including soil  $cSoil$ , litter  $cLitter$ , and coarse woody debris  $cCwd$ ) where multiple pools were reported. Soil carbon inputs were calculated from the monthly change in soil carbon stock plus the reported heterotrophic respiration. Model variable summaries can be found in Tables SI3 and processing code is documented in SI.

These 20 Earth system models are built from previous models which contain 10 distinct soil sub-models (Table 1). The number of soil carbon pools in these ESMs varied from 1 (INM-CM4) to 8 (BCC-CSM1.1) with most models having 2 to 5 pools. None of the models reported soil carbon with depth although GFDL documents a depth dependent model. There were three classes of pool structure for these models: *independent* where there was no exchange between soil carbon pools, *cascade* where pools with faster turnover times passed carbon to pools with slower turnover times, and fully *feedback* models where carbon was exchanged between faster and slower pools and vis-versa. In this set of models; 2 of these soil models were full feedback models (HadGEM, ISPL-CM), 6 were cascade pool structure (MRI-ESM1, MIROC-ESM, MPI-ESM, CLM4.0 [CESM1, CCSM4, NorESM1], CanESM2, BCC-CSM1.1), and 2 were independent pools (GFDL-ESM2, INM-CM4). Only 1 model documented an explicit constant  $Q_{10}$  (INM-CM2,  $Q_{10}=2$ ), 1 model documented a temperature dependent  $Q_{10}$  (CanESM2), 1 proposed a unique temperature sensitivity (HadGEM2), 4 models documented a temperature sensitivity from Lloyd and Taylor which behaves very similar to  $Q_{10}=2$  under moderate temperatures (Lloyd and Taylor, 1994), and the remaining 3 (ISPL-CM5, GFDL-ESM2, BCC-CM1.1) all used a variation of the temperature sensitivity proposed in CENTURY (Parton et al., 1987, 1988) which also behaves very similar to  $Q_{10}=2$  under moderate temperatures (Lloyd and Taylor, 1994). The ESMs considered had a single global  $Q_{10}$ , or  $Q_{10}$ -formula dependent on temperature, uniformly applied to the decay pools. This documented structure should be approached with caution due to frequent lags between model development and documentation, actual values and functions may differ. For details with citations see Table 1.

Soil carbon stocks at the beginning of Earth system model simulations are typically spun up to close to steady state, and there is numerical support that this holds throughout the simulation (see Results and Figure S3), then Eq. 3 can be extended to the change in soil carbon stock over the 21<sup>st</sup> century. This leads to the following explicit calculation for a  $Q_{10}$  value at each grid cell.

$$\ln(Q_{10}) = \left( \frac{10}{T_m - T_f} \right) \ln \left( \frac{C_f u_m}{C_m u_f} \right), \quad (5)$$

where the  $Q_{10}$  value is related to the modern soil temperature  $T_m$ , future soil temperature at the end of the 21<sup>st</sup> century  $T_f$ , modern soil inputs  $u_m$ , future soil inputs  $u_f$ , modern soil carbon stock  $C_m$  and the future soil carbon stock  $C_f$ .



For soils that are very close to zero soil carbon stocks, have minimal shifts in temperature or have very low soil inputs, the estimated  $Q_{10}$  is not finite. Similarly soils which are not well described by their shift in temperature (for example, if there is a significant shift in the moisture regime) may have non-typical  $Q_{10}$  values that are either less than 0.5 or greater than 5. We examined the amount of shift in soil carbon stocks associated with the four categories of  $Q_{10}$  values (nonfinite, less than 0.5, greater than 5, or typical), as well as the spatial patterns associated with these categories.

To support the assertion that the  $Q_{10}$  value can be calculated from relatively short time scales found in the field experiments, we examined the distribution typical  $Q_{10}$  values associated with similar temperature steps experienced by the field experiments at 1, 5, 10, 15, 20, 50, 75, and 84 year time-scales using 10-year mean gridded values of soil carbon stocks, soil inputs, and soil temperature. It should be noted however that changes in the moisture conditions over the 21<sup>st</sup> century may complicate this analysis of the Earth system model simulations, thus it is not an exact proxy for the field experiments where the control and treatment plots experienced similar baseline climate conditions and a more or less constant offset throughout the experiment. Finally, the  $Q_{10}$  distribution was scaled to reflect the best estimate and uncertainty from the field data. This distribution shift was done by normalizing the  $Q_{10}$  map to the mean of the distribution and multiplying it by the experimentally derived values. The  $Q_{10}$  correction was only applied to grids with typical  $Q_{10}$ 's (non-typical  $Q_{10}$ 's were considered to have predominately non-temperature driving variables and their soil carbon stocks were not altered). This normalization shifted the global  $Q_{10}$  distribution within the models to match the most common (geographically likely)  $Q_{10}$  with the data-drive  $Q_{10}$  value, yet by preserving the distribution we preserved other factors affecting changes in decomposition rate (i. e. moisture shifts) in the model. We then recalculated the change in soil carbon for each grid cells with this modified  $Q_{10}$  according to Eq. 3 and calculated the global area-weighted totals.

The full analysis script and those used to generate the figures are available in the supplemental.

### 3 Results

From the changes in soil carbon stocks across field studies, we find a global  $Q_{10}$  of 2.2 (90%CI 1.6, 2.7;  $R^2 > 0.95$ , root mean squared error  $< 2$ ; Figure 1, Figure S2). The model-data fit was evaluated using a linear regression and root mean squared error (Figure 1). While the  $R^2$  of the model-data comparison was relatively insensitive to the  $Q_{10}$  value, there was a notable improvement in the bias with  $Q_{10}$  (as defined as the slope within 2 standard deviations of 1 and intercept within two standard deviations of 0). This bias-driven criteria was used to select  $Q_{10}$  values from a prior range of (0.1, 5), see Methods for details. The  $Q_{10}$  distribution was compared with a random null distribution and was significantly distinct (Kolmogorov-Smirnov  $D=0.441$ ,  $p < 2e-16$ , See Table S2, Figures 2 and S1). Randomly selecting 5 to 34 sites from the full dataset were compared to a null distribution where control vs warmed labels were randomized. The quartiles of the data subsets notably converged at a sample size of 25 where the null distribution was relatively invariant across sample size (Figure 2). The distribution of the  $Q_{10}$  values under null appeared log-normal, centered around 1 demonstrating no temperature effect (Figure S1). The distribution of the  $Q_{10}$  range for the data subsets converged to around 2.2 (Figure S1).



The balance between gridded soil inputs and heterotrophic respiration at both the initial and final 10-year mean for the 21<sup>st</sup> century was within 10% for over 93% of the grid cells across all models. Most models had 95% and two models consistently had 100% of their grid cells within 10% -- the absolute value of the net flux was within 10% of the highest primary flux (Figure S3).

- 5 The inferred  $Q_{10}$  values in the Earth system models derived from 10-year mean changes across different time steps (1, 5, 10, 15, 20, 50, 75, and 84 years) had similar distributions in most of the models (Figure S4). Differences in inputs and heterotrophic respiration drove some differences in the  $Q_{10}$  distribution in the MIROC-ESM models. There were minor shifts in the mode of most models which could be attributable to changes in the moisture conditions or other (non-temperature or input) environmental variables in the simulation. Models aggregated across common land models showed marked similarity in their
- 10  $Q_{10}$  distributions (Figure S5). There was also an extremely high correlation between  $Q_{10}$  values derived from soil inputs compared to those derived for heterotrophic respiration across all models (Figure S6).

The inferred  $Q_{10}$  values in the Earth system models from the decadal average across the 21<sup>st</sup> century fell into four categories (nonfinite, less than 0.5, greater than 5, or typical; Figure S7), however most of the change in soil carbon stocks over the 21<sup>st</sup> century occurred in grid cells with typical  $Q_{10}$  values between 0.5 and 5 (Figure S6). A notable exception to this trend was the

15 MRI-ESM1 model where roughly half of the change in carbon stocks occurred in grid cells with  $Q_{10}$  values greater than 5 (Figure S6). Spatially the  $Q_{10}$  categories showed strong geographical patterns (Figure S7). The GFDL-ESM2 models were dominated by non-finite values in high northern latitudes (Figure S7). MIROC-ESM, CCSM4, CESM1, and NorESM1 models were dominated by  $Q_{10}$  values above 5 in the high northern latitudes (Figure S7). Unless otherwise noted, only typical  $Q_{10}$  values are addressed for the remainder of this study.

- 20 The inferred  $Q_{10}$  values for the decadal average across the 21<sup>st</sup> century, also showed strong geographic patterns (Figures 3) and was typically unimodal (Figure S6). MIROC-ESM and MIROC-ESM-CHEM showed the weakest spatial patterns with high grid-to-grid variation (Figure 3). Mean  $Q_{10}$  values fell within the 90% CI of the field data  $Q_{10}$ , ranging between 1.8 (GFDL-ESM2G) and 2.6 (MIROC-ESM-CHEM), with the multi-center  $Q_{10}$  values at  $2.0 \pm 0.2$  (Tables 2).

When the inferred  $Q_{10}$  values were modified to reflect the data-driven  $Q_{10}$  range, resulting variation in the multi-center mean

25 was almost as large as the variation across model projections (Figure 4, Table 2). Re-centering the global  $Q_{10}$  distribution to reflect the range of field-driven  $Q_{10}$  values (Figure S8) resulted in changes in soil carbon stocks over the 21<sup>st</sup> century of between -452 Pg-carbon (MPI-ESM-MR) and 285 Pg-carbon (HadGEM2-CC) with a best-estimate  $Q_{10}$  ( $Q_{10} = 2.2$ ) resulting in  $16 \pm 156$  Pg-carbon (multi-center mean  $\pm 1$  SD) and field-drive bound ( $Q_{10} = 1.6, 2.7$ ) of  $[245 \pm 194, -99 \pm 208]$  Pg-carbon (Figure 4, Table 2).

#### 30 4 Discussion

By capturing information about warming induced changes to relatively undisturbed field soil carbon stocks directly rather than inferring this from soil respiration rates, this is the first study to generate field  $Q_{10}$  estimates of soil carbon losses without



- needing to correct for belowground autotrophic respiration. Using a simplified version of a traditional decomposition model with a temperature sensitivity function, we estimate that the global  $Q_{10}$  value is 2.2 ([1.6, 2.7] 95%CI, Figure 1, S2). This  $Q_{10}$  is notably higher than previous global estimates based on field soil respiration data ( $Q_{10} = 1.4$  to  $1.5$  (Bond-Lamberty and Thomson, 2010; Mahecha et al., 2010)), yet well within the range of estimates from laboratory-based studies (Davidson and Janssens, 2006) as well as close to documented temperature sensitivity parameters ( $\sim 2$ ) of Earth system models (Table 1). This  $Q_{10}$  range is statistically significant. Resampling the 36-study data set demonstrates the need for over 25 sites to distinguish the  $Q_{10}$  range from random (Figure 2 and S1). While the  $Q_{10}$  distribution for the 34-study subset is distinct from the null (Kolmogorov-Smirnov  $D=0.441$ ,  $p < 2e-16$ ), there appears to be some minor drift in the range suggesting that more study sites could be informative.
- 10 Inferring a decadal-scale environmental sensitive from an annual-scale experiment is controversial but support by both traditional model structure and numerical trends in the Earth system model. In the traditional model structure the temperature sensitivity function is applied as a single scaler to multi-pool models causing the relative decomposition response in both fast and slow pools to be the same (for example, (Parton et al., 1987)). Examining the inferred gridded  $Q_{10}$  values from annual means across time scales from 1 year to 84 years in Earth system models shows a strong similarity in the distribution most
- 15 models (Figure S4), with the notable exception of MIROC-ESM which is explained by unusual differences in soil inputs and outputs (Figure S4). Differences in the  $Q_{10}$  distribution across time scales are likely driven then by interaction with other sensitivity functions like moisture.
- If soils are more sensitive to warming than previously expected, then how would this affect future soil carbon stocks over the 21<sup>st</sup> century? To address this question, we turned to the CMIP5 Earth system models run under RCP 8.5 (Taylor *et al.*, 2011).
- 20 In order to modify the Earth system model output to reflect the data-driven  $Q_{10}$  we applied similar assumptions used in the field-data analysis. We first examine the temperature sensitivity of CMIP5 Earth system model simulated soil carbon stocks. In contrast to the field data, we take into account the effect of the change in soil inputs on soil carbon stocks in the Earth system models because these coupled simulations include  $CO_2$  fertilization and other climate effects known to influence primary production (see Methods, Eq. 5). Though these inferred  $Q_{10}$  values ( $Q_{10} = 1.8, 2.6$ ) fall within the uncertainty of the field
- 25 derived  $Q_{10}$  values ( $Q_{10} = 1.6, 2.7$ ), most EMS- $Q_{10}$  modes (most common gridded  $Q_{10}$  value within each ESM) fell under the median data- $Q_{10}$  value of 2.2 (Table 2) implying ESMs were, on average, less sensitive to temperature shifts than the field warmed data would imply. It should be noted that this inferred  $Q_{10}$  value is not exactly the parameterized  $Q_{10}$  value, and is instead a combination of the temperature sensitivity and other environmental sensitivities. If there were, on average, an additional constraint on respiration (such as moisture) we might expect the inferred  $Q_{10}$  parameter to be lower than the model
- 30 parameterized  $Q_{10}$ .
- There were notable regional patterns across all but 2 of the Earth system model inferred  $Q_{10}$  (Figure 3, S7). High northern latitudes tended to have either large or non-finite  $Q_{10}$  values suggesting that something other than temperature and input shifts were driving changes in soil carbon stock. This alternative driver could be a shift in moisture regimes or dynamics driven by thaw thresholds, this should be a focus of future research.



Propagating this field- $Q_{10}$  range into the ESM projections resulted in greater carbon losses from the soil by the end of the 21<sup>st</sup> century (multi-center means the of soil carbon change, shifted from 83 to 16 Pg-carbon) with large uncertainties; ESM multi-center standard deviation was initially 156 Pg-carbon which is half of the range in multi-model mean attributed to  $Q_{10}$  95% CI [245, -99] Pg-carbon (Figure 4, Table 2). To calculate these modified projections, means of the model specific  $Q_{10}$  distributions were re-centered to reflect the best estimate  $Q_{10}$  and associated 95% CI from the field data analysis. By preserving the distribution within the model, we attempted to propagate soil moisture sensitivities and other model-specific effects into the modified projections. We also did not modify grid cells with non-typical  $Q_{10}$  values (non-finite, below 0.5 or above 5) since those grids likely governed other non-temperature drivers. The large range of carbon shifts in each ESMs driven by this  $Q_{10}$  CI confirms the importance of considering parameter uncertainty in the land carbon component of Earth system model projections. The post-hoc correction that we present provides an innovative way to account for this parameter variation without the computational burden of additional ensemble runs.

This analysis includes several basic assumptions and caveats. Specifically, we assume that the difference between treatment and control is driven entirely by the soil warming effect, and those warming effects are uniform across soil carbon quality. Though warming-induced changes in soil inputs are, on average, relatively small, they have been shown to be highly variable in similar sites (Lu et al., 2013). The analysis of field data could be extended to account for these changes in inputs in follow-up studies (Eq. 3). A large increase in soil inputs would cause an underestimation of the  $Q_{10}$  value, while a decrease in soils inputs would cause an overestimation of the  $Q_{10}$  value (see Eq. 3). While there is some evidence to support temperature sensitivity dependency on soil carbon quality (Knorr et al., 2005), there is also evidence for a uniform temperature sensitivity (Hicks Pries et al., 2017). as is represented in the Earth system models considered in this study (Table 1). A quality dependent  $Q_{10}$  would not be separable from the bulk decay term and thus a one pool model would be inappropriate in this case (see SI). In addition, the dataset has acknowledged biases (see Crowther *et al.*, 2016), which are typical of field studies.

#### 4.1 One pool simplification

We find that multi-year soil carbon dynamics can be well-described by a one pool model at a specific time scale in both the Earth system models and field experiment. If we restrict the decomposition models to those with either independent or cascade pool structures (that is no carbon passed from the slow to the fast pools) then the temporal dynamics of the total soil carbon of the system at a specific time scale can be approximated by a single pool due to the fact that the lower triangular decomposition-transfer matrix is diagonalizable (see SI for details). While this diagonalizable property does not hold for full feedback models where carbon is transferred from the slower to faster carbon pools, all decomposition-transfer matrices are M-matrices. If we combine the positive-inverse properties resulting from this M-matrix structure and assume that the soils are close to metastable state (that is soil inputs are roughly equal to the heterotrophic respiration outputs, as we show in Figure S3 for the Earth system models considered and would expect for soils from intact systems). Then the total soil carbon can be described by a bulk decay rate that is a linear combination of the transfer coefficients, decay rates, and input allocations of the component pools (see SI



for analytical details). This provides analytical support for the one pool simplification seen numerically in the Earth system models in the CMIP5 project (Todd-Brown *et al.*, 2013, 2014).

The one pool simplifications described above are controversial assertions. The one pool model has proven inadequate to describe laboratory incubations where heterotrophic respiration over time is compared to the soil carbon stock (Thornton, 1998; Weng and Luo, 2011). This is due to the multiple time scales considered (daily, monthly and annual) and, more importantly, the fact that these laboratory incubations are by their nature not at steady state since any inputs to the system are generally removed. Thus this analysis would not be expected to hold for laboratory incubation and we would further expect the bulk decay rate to change with time scales (in other words, the bulk decay rate inferred at a 1 year time step would not match the 100 year time step). However, the scalar multiplier representing environmental sensitivities are independent of pools in most models (ex (Parton *et al.*, 1988)). These scalar multipliers (like the  $Q_{10}$  temperature sensitivity examined in this study) would be invariant to time scale.

## 5 Conclusion

It is still unclear how the terrestrial carbon cycle in general, and soils in particular, will respond to climate change over the 21<sup>st</sup> century. The CMIP5 models, representing our best coupled climate models to date, have a wide range of soil carbon responses over the 21<sup>st</sup> century (Todd-Brown *et al.*, 2014). While it would be nice to have all the models agree on a tightly bound answer, the question we should be asking scientifically is: Does the variation in the models reflect our best scientific understanding? Models must capture not only mean trends but also system variance and accurately represent scientific uncertainty.

Post-hoc correction of simulation results can provide some insight into known gaps in Earth system models without the computational hurdle of re-running simulation results. Previous studies have address nutrient limitations on net primary production (Wieder *et al.*, 2015) and this study demonstrates the high level of uncertainty that can be driven by the soil temperature response parameter. This study suggests that soil carbon response to warming is highly variable in the field and model studies should increase their variability to capture this. Future studies explaining field moisture and applying that understanding to a post-hoc Earth system model analysis is a logical next step.

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The authors have no conflict of interest to declare.

### 7 Author Contributions

- 10 KTB and TWC both contributed to the writing of this manuscript. KTB was responsible for the analysis. TWC coordinated the field data set. BZ was responsible for the mathematical analysis and made contributions to the revisions of the manuscript.

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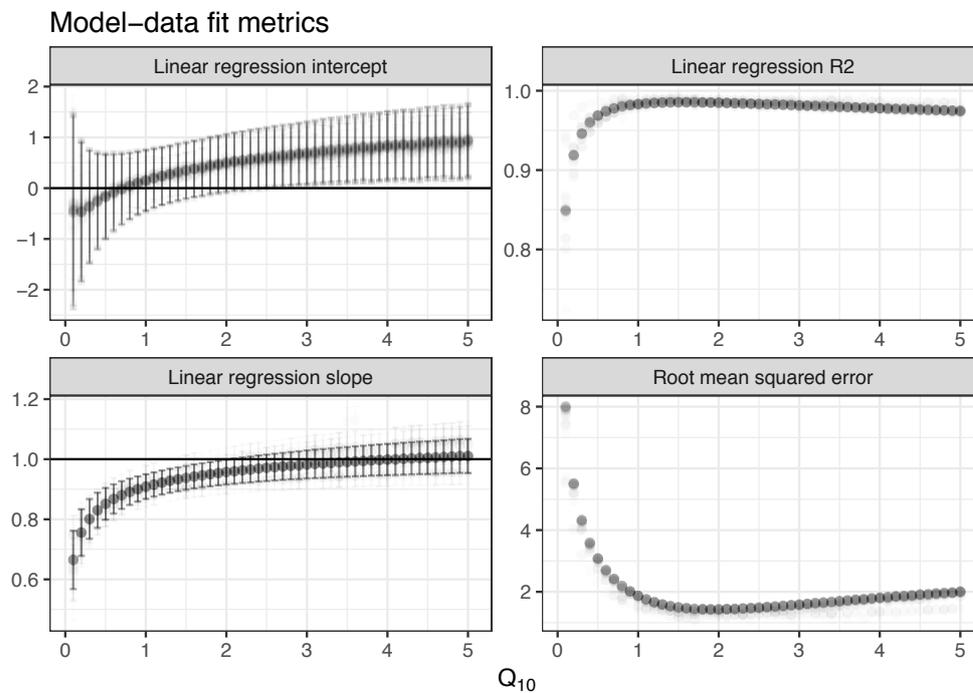
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**Figure 1: The model-data fits across different  $Q_{10}$  values for random subsets of 34 sites including the root mean squared error, and linear regression metrics  $R^2$ , slope, and intercept. The model is taken from Eq. 4 ( $C_w = C_e Q_{10}^{-\Delta T/10}$ ). Slope and intercept values are shown with 2 standard deviation error bars.**

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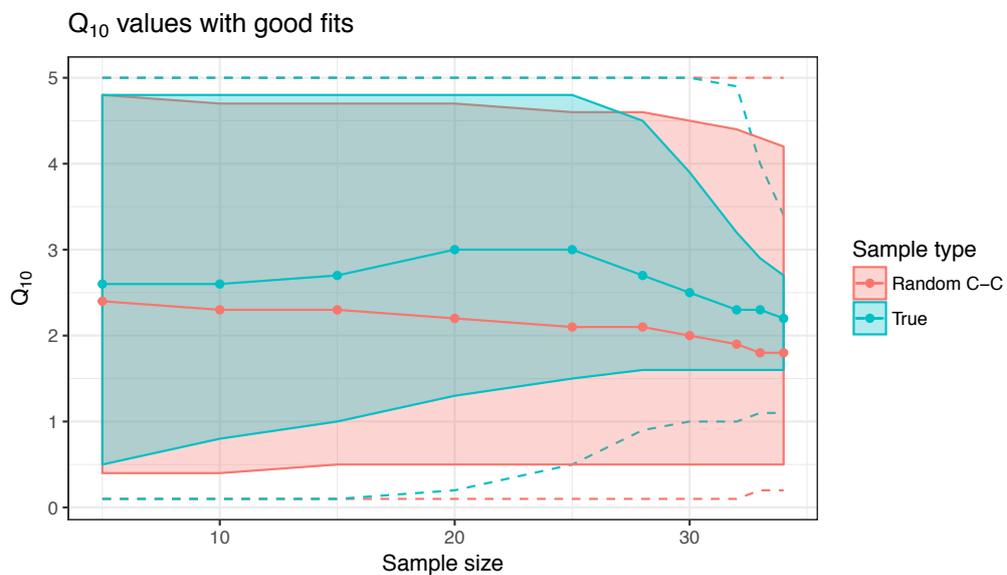


Figure 2: The Q<sub>10</sub> with good 1-to-1 model-data fits defined in Figure 1, at 90% confidence interval (band) with minimum and maximum values (dotted line) and median value (solid line), across 10 different sample sizes ranging from 5 to 34, for the original data set (True: blue) and randomized case-control (Random C-C: red).

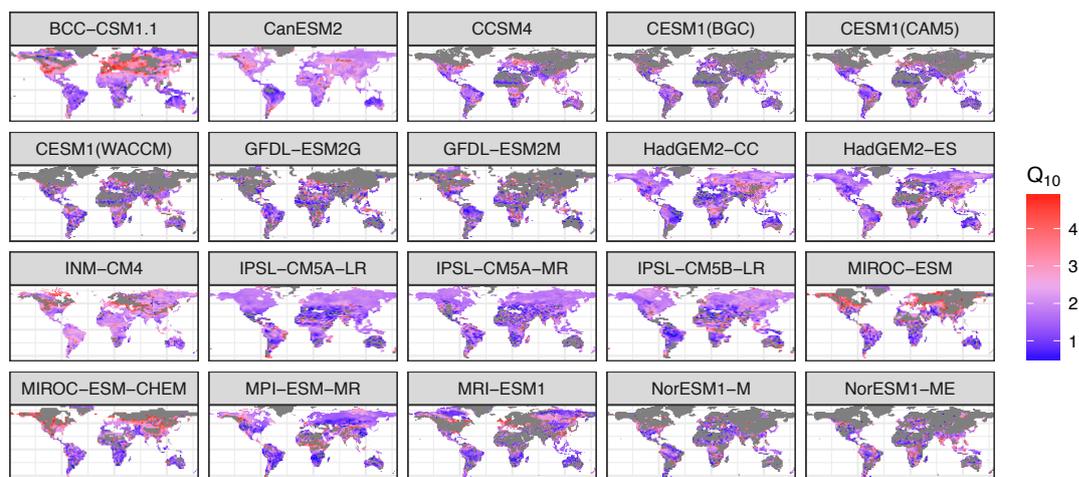


Figure 3: Inferred  $Q_{10}$  values from the Earth system models (CMIP5, RCP 8.5). The colour scheme is centered around the field-driven  $Q_{10}$  median value of 2.2. Grey indicates non-typical  $Q_{10}$  values that were either non-finite, less than 0.5 or greater than 5.

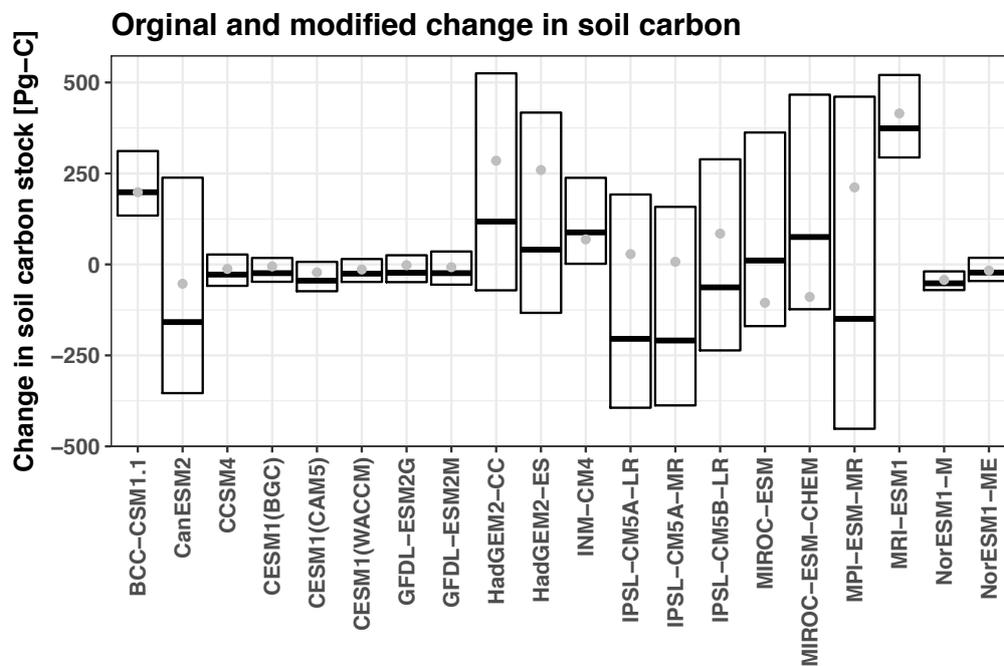


Figure 4: Changes in soil carbon stock (10 year means) over the 21<sup>st</sup> century from Earth system models (RCP 8.5). Grey dots are the original estimates, the open box is the soil carbon loss after the  $Q_{10}$  is rescaled using the 2.5%, 50%, and 97.5% quartiles from the field data



Model Center	Earth system model	Soil/land carbon sub-model	Number of soil carbon pools	Pool structure	Temperature sensitivity	Citations
BCC	BCC-CSM1.1	BCC_AVIM1.0; AVIM2; CEVSA	8	Cascade	CENTURY	(Cao & Woodward, 1998; Ji <i>et al.</i> , 2008; Wu <i>et al.</i> , 2013)
CCCma	CanESM2	CTEM	2	Cascade	$Q_{10}(T)$	(Arora, 2003; Arora & Boer, 2010; Arora <i>et al.</i> , 2011)
NCAR	CCSM4	CLM 4.0; CN	7	Cascade	Lloyd and Taylor	(Thornton & Rosenbloom, 2005; Thornton <i>et al.</i> , 2007; Oleson <i>et al.</i> , 2010; Gent <i>et al.</i> , 2011; Lawrence <i>et al.</i> , 2011; Tjiputra <i>et al.</i> , 2013)
NSF-	CESM1(BGC)					
DOE-	CESM1(CAM5)					
NCAR	CESM1(WACCM)					
NCC	NorESM1-M					
	NorESM1-ME					
NOAA	GFDL-ESM2G	LM3.0; ED	2	Independent	CENTURY	(Moorcroft <i>et al.</i> , 2001; Shevliakova <i>et al.</i> , 2009; Dunne <i>et al.</i> , 2013)
GFDL	GFDL-ESM2M					
MOHC*	HadGEM2-CC	ROTHC	4	Feedback	f(T)	(Coleman & Jenkinson, 1999; Collins <i>et al.</i> , 2011)
	HadGEM2-ES					
INM	INM-CM4	LSM 1.0	1	Independent	$Q_{10}=2$	(Bonan, 1996; Volodin, 2007)
IPSL	IPSL-CM5A-LR	ORCHIDEE	5	Feedback	CENTURY	(Krinner <i>et al.</i> , 2005; Dufresne <i>et al.</i> , 2013)
	IPSL-CM5A-MR					
	IPSL-CM5B-LR					
MIROC	MIROC-ESM	SEIB-DGVM	3	Cascade	Lloyd and Taylor	(Sato <i>et al.</i> , 2007; Watanabe <i>et al.</i> , 2011)
	MIROC-ESM-CHEM					
MPI-M	MPI-ESM-MR	JSBACH	5	Cascade	Lloyd and Taylor	(Giorgetta <i>et al.</i> , 2013; Schneck <i>et al.</i> , 2013)
MRI	MRI-ESM1	LPJ-DGBM	3	Cascade	Lloyd and Taylor	(Sitch <i>et al.</i> , 2003; Obata & Shibata, 2012; Adachi <i>et al.</i> , 2013)

**Table 1:** This is a summary of the soil decomposition sub-models for the ESMs used in this study and includes the number of pools, structure of the carbon exchange between those pools, temperature sensitivity function, and citations. Temperature sensitivity function is either denoted as: CENTURY: borrowed from the Century model (Parton *et al.*, 1987, 1988), Lloyd and Taylor: taken from the recommended form from (Lloyd & Taylor, 1994),  $Q_{10}(T)$ : a temperature dependent  $Q_{10}$  as defined by (Arora, 2003), f(T):



a unique temperature sensitivity function defined by (Coleman & Jenkinson, 1999), or a  $Q_{10}$  parameter for the  $Q_{10}$  function as defined in this manuscript.



	SOC	Rel.	dT	dSOC		dSOC	dSOC	dSOC
	[Pg-C]	Inputs	[°C]	Q <sub>10</sub>	[Pg-C]	Q <sub>10</sub> =1.6	Q <sub>10</sub> =2.2	Q <sub>10</sub> =2.7
BCC-CSM1.1	1050	1.40	3.7	2.2	198	312	198	134
CanESM2	1541	1.29	7.1	2.0	-53	239	-158	-354
CCSM4	575	1.32	4.2	2.0	-12	27	-28	-59
CESM1(BGC)	575	1.29	3.8	1.9	-6	18	-24	-48
CESM1(CAM5)	619	1.30	4.6	1.9	-22	7	-45	-74
CESM1(WACCM)	559	1.32	3.9	2.0	-14	15	-25	-48
GFDL-ESM2G	1422	1.41	5.1	1.9	-2	25	-23	-49
GFDL-ESM2M	1278	1.38	4.5	2.0	-8	36	-24	-56
HadGEM2-CC	1122	1.55	8.4	1.9	285	525	118	-71
HadGEM2-ES	1129	1.56	8.3	1.8	259	417	41	-133
INM-CM4	1688	1.27	3.3	2.3	69	238	88	2
IPSL-CM5A-LR	1361	1.48	8.2	1.8	28	192	-205	-394
IPSL-CM5A-MR	1403	1.43	7.6	1.8	7	158	-209	-387
IPSL-CM5B-LR	1274	1.41	7.6	1.9	85	289	-63	-236
MIROC-ESM	2586	1.35	7.2	2.5	-105	363	11	-170
MIROC-ESM- CHEM	2588	1.30	7.3	2.6	-89	467	75	-123
MPI-ESM-MR	3110	1.31	6.3	1.8	212	461	-150	-452
MRI-ESM1	1452	1.52	4.4	2.0	415	521	374	294
NorESM1-M	610	1.31	3.7	2.0	-43	-19	-52	-71
NorESM1-ME	618	1.32	3.6	2.1	-17	18	-22	-46
Multi-center mean	1419	1.37	5.4	2.0	83	245	16	-99
Multi-center sd	775	0.09	1.8	0.2	156	194	156	208

Table 2: Global model summary with multi-center mean and standard deviation for modern soil organic carbon (SOC) stocks [Pg-C], relative shift in soil inputs ( $\frac{u_r}{u_m}$ ), absolute change in soil temperature (dT) [°C], inferred mean of Q<sub>10</sub> as calculated by grid cell (see Eqn 5), the change in soil organic carbon (dSOC) over the 21<sup>st</sup> century [Pg-C], and the change in soil organic carbon with rescaled Q<sub>10</sub> values (1.6, 2.2, and 2.7).

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