Point-by-point response to reviews

Response to Referee #1

1. We focused on global relationships (a) because the methane and methane-isotope data, which we use to support our assumption that the charcoal records provide a global signal of biomass burning, are by their nature globally integrated and (b) because quantifying the strength of a global feedback requires global data; climate feedbacks cannot be assessed on a regional basis. However, we agree that it is interesting to test whether there are regional differences in the nature of the response of fire to specific drivers. There have been multiple papers examining the controls on fire as recorded by charcoal data, both at a global scale and at regional scales, only some of which are cited in this paper (e.g. Marlon et al., 2008; Harrison et al., 2010; Daniau et al., 2012). We think that it would be useful to cite some more of these papers (e.g. Power et al., 2008: Marlon et al., 2009; Mooney et al., 2011; Marlon et al., 2013; Marlon et al., 2016) and to expand our discussion of the charcoal/temperature relationship as discussed in these papers, and particularly focusing on the results from Daniau et al. (2012) and Marlon et al. (2013). Daniau et al. (2012) showed similar positive relationships between temperature and fire, and unimodal relationships with moisture (precipitation minus evaporation), at a global scale in both charcoal data covering the last 21,000 years and satellite (GFED3.1) data. Marlon et al. (2013) showed positive relationships between temperature and charcoal on centennial to millennial times scales over the Holocene (past ca 12,000 years) for data-rich regions, including Europe, North America and eastern Asia. The strength of this relationship varies, from an r^2 of 0.85 in North America to 0.33 in eastern Asia, showing that other factors also play a role, but nevertheless the nature of the relationship itself (higher temperatures, more fire) does not change from region to region.

We also agree that it would be useful, in the context of the present paper, to include some additional information about regional relationships between temperature and fire. Many recent papers have already described spatial and temporal patterns and correlates of biomass burning based on GFED products. Given the fact that the global relationships in the satellite-era are statistically weak (even at a global scale) because of the short length of the record and the strong anthropogenic impact on recent global emissions, we have focused these analyses on the charcoal record. Because the data coverage is uneven across continents we have confined the analysis to broad latitudinal bands, i.e. N and S tropics and extratropics. This analysis strongly supports our treatment of biomass burning variability as a function of temperature. We propose adding it to Section 3.4.

The key point here is that, as we discuss in the paper, fire initiation and spread are controlled by multiple variables including ignitions, temperature and moisture-related climate variables, vegetation properties, and anthropogenic influences on landscape fragmentation and land management. Analyses that have separated the independent role of each of these variables (e.g. Krawchuk et al., 2009; Bistinas et al., 2014) show that the apparent effect of any one variable in a specific region or at a specific time is not the same as the fundamental effect. For example, the fundamental impact of climate on the fire regime can be overwhelmed by changes in vegetation properties or by human activities, but nevertheless, increased temperature will always lead to an increase in burning (all other things being equal). At the global scale, the impact of temperature is paramount because changes in temperature influence other aspects of the climate system: e.g. the equator-to-pole temperature gradient controls atmospheric circulation patterns and wind strengths, and there is a strong dependence of rainfall patterns on global temperature changes. Thus, for a global feedback

analysis, there are excellent reasons to quantify fire feedbacks in terms of global temperatures, as has also been done for other feedbacks (e.g. Arneth et al., 2010).

2. We discuss (in paragraph 2 of the Discussion) the reasons why the feedback strength might be different in the pre-industrial and recent periods, specifically because of the impact of human activities on deforestation, land-use, landscape fragmentation and fire suppression. We will add a further comment on this in the abstract, to the effect that *although the feedback estimates from palaeo and satellite-era data are in agreement, this is likely fortuitous because of the pervasive influence of human activities on fire regimes during recent decades.*

3. Thank you for drawing our attention to the Petrenko et al. (2017) article, which shows considerably lower methane emissions (15.4 Tg CH_4 a^{-1}) based on the ice core record of the Younger Dryas. This is considerably lower than the figure given by Schwietzke et al. (2016). However, both papers make the assumption that the geologic flux is constant. We do not use any estimate of this flux in our calculations; it is simply assumed to be constant. Thus, the quantitative disagreement between Petrenko et al. and Schwietzke et al. is immaterial to our argument. However, we will add the reference to the text because the difference in the estimate of this assumed constant flux may be of interest to readers.

4. We agree completely, and indeed this is what we say (page 1 lines 17-18, page 8, lines 3-8, page 10, lines 19-25) in the manuscript: the modern signal is dominated by fires associated with deforestation and peatlands, and although both of these vary with climate they are primarily an anthropogenic signal.

Response to Referee #2

We thank the reviewer for their positive comments about the novelty of this work, and for recognizing the value of the palaeo-record in addressing issues that are not satisfactorily resolved using contemporary data alone.

1. Yes human activities do affect ignitions, and we propose to further emphasize this in the Introduction, but the primary impact is through modifying fuel availability and fire spread.

2. We use GFED4s in this paper, but we agree that since this is derived from GFED4, it is appropriate to include the van der Werf (2017) reference here.

3. The Arora et al. paper was cited correctly in the reference list as 2013, but incorrectly as 2014 at some places in the text. We will correct this.

4. Yes, we can (and propose to) provide a more detailed breakdown of the total land climate-carbon cycle feedback from different "no-fire" models as given in Arora et al. (2013). The spread of values among all models is large, but this is partly due to the inclusion of two CMIP5 Earth System models (using the same, fire-enabled land model) that have been shown to greatly underestimate the strength of this feedback based on the observed relationship between tropical land temperatures and the growth rate of atmospheric CO_2 (Wenzel et al. 2014). After removing these two models we are left with only two that explicitly represent fire, and their feedback strengths are in the same range as those for the five models that do not explicitly represent fire. Thus, the main purpose of citing Arora et al. (2013) here is simply to give a general idea of the magnitude of this feedback as represented in Earth System models. We suggest this is best expressed by the median of the "no-fire" models (17.5 ppm K^{-1} after correction for the airborne

fraction). The total range (excluding the two models mentioned above) is from 6.8 to 19.9 ppm K^{-1}.

5. We point out already that the total fire-feedback estimate from Ward et al. (2012), which is negative, depends strongly on the (highly uncertain) magnitude of the indirect aerosol effect. We should point out that their estimate of the effect of fire on the carbon cycle is also questionable. They estimated that elimination of fire would increase land carbon storage by a large amount, approximately 500 Pg C. The effect of this change on atmospheric CO_2, however, is masked in their analysis by the application of a small effective airborne fraction of 0.177 – which is based on a formula designed for application to much longer than centennial time scales. Moreover this analysis is based on a single model. Work now in progress (Lasslop et al.: *Geophysical Research Abstracts* 20, EGU2018-13445, 2018) shows that there are very substantial differences in how current models treat fire: for example, the decrease in the carbon turnover time due to fire varies between 2.5 and 10% across the FIREMIP ensemble of models. It would be much more useful and robust to address this issue using outputs from multiple models, and preferably when we have a better understanding of why models show such a large range of responses.

Response to Referee #3

Thanks Sam for this positive review, the helpful suggestions and spotting the typos.

1. Although the methodology is not complex, we agree that it might seem so given that we are including analyses of satellite-era and palaeo-data that have to be carried out somewhat differently. We think it will be worthwhile to include a paragraph at the beginning of the Methods section to spell out the underlying logic and to illustrate this with a flowchart, as you suggest. This paragraph will also clarify the logic of including the methane and methane-isotope versus charcoal comparison in this paper. We agree that establishing the good relationship between the methane record and the charcoal record is an important part of this paper since it demonstrates conclusively, and for the first time, that the assumption that charcoal can be interpreted as a record of biomass burning on palaeo-timescales is valid. We propose to emphasise this point further in the revision.

2. We agree that the description of the GFED data sets in the Readme file and in the various publications is not exactly clear. In van der Werf et al. (2017), which is the most comprehensive description of what was done to obtain the GFED4 and GFED4s data sets, it is clearly stated that the MODIS-era data starts in August 2000. The derivation of pre-MODIS era burnt area from the VIRS and ASTR active fire data is via optimization against the post-2001 MODIS data. Our decision to include 2000 was because the record included five months of MODIS data, corresponding to the southern hemisphere fire season, and increased the number of data points available in a time series that was already limited in length. However, we did test whether inclusion of these data made a difference to our results. What emerges from these tests, documented in the Supplementary, is that the year 2000 is not anomalous and if there is an overly-influential observation it is 2003. Thus, we feel comfortable with using the 2000-2014 period in our analyses. However, we will note that the records for 2000 are a mixture of pre- and post-MODIS in the methods section and justify our inclusion of this year in the analyses.

3. This is a very good suggeston. The flowchart that we propose to add at the start of the Methods section will make it clear how we go from a relationship with mean land temperature to a feedback strength related to mean global temperature.

4. Our focus in the SI was weighted towards explaining the derivation and testing of relationships in the palaeodata, because we assumed that this was less well known. However, we will expand the SI further to include parallel information about the modern data to that provided for the palaeodata.

5. The important point here is not that our central estimate of δ_m is low. The point is that it is highly uncertain (with confidence intervals wide enough to include all published values), so our analysis does not allow us to place any further constraint on the appropriate value of δ_m. Our calculations do not assume any particular value.

6. We agree that we do not explicitly state that the relationship between temperature and fire is positive, although we cite a number of references that explicitly show that it is, but we will clarify this in the text.

7. We do not claim that peatland only burns as a result of human intervention. Palaeodata show that peatlands have burnt through natural fires (see e.g. Grant et al., 2014 Journal of Quaternary Science; New et al., 2016, Mires and Peat, Volume 18, Article 26, 1–11, http://www.mires-and-peat.net/, ISSN 1819-754X). However, it is true that the major tropical peatland fires in recent years have been associated with substantial modification of the natural environment by humans, particularly through drainage for agricultural use. Nevertheless, in the case of both peatlands (see e.g. Page et al., 2009 DOI 10.1007/978-3-540-77381-8_9) and deforestation fires (see e.g. van der Werf et al., 2008, PNAS), there is also a strong climate driver with major fire years associated with ENSO variability. Thus, in making our analyses, we initially exclude agricultural fires – on the assumption that these are solely human controlled – and subsequently exclude peatland and deforestation fires because of the likelihood that they show a substantial imprint of human activities. We agree that other landscapes may be heavily human-modified, but that does not necessarily detract from the fact that climate plays a major role in the year-to-year incidence of fires.

8. We have provided a high level of detail in the Supplementary Information so that others can reproduce our analyses. And it is clear that this is useful because you are asking for further expansion of the SI. However, we agree that it would be useful to provide a reader with more guidance about which sections of the SI to refer to in the main text, and we will pay attention to enhancing its readability.

9. We have changed this to "N2O, and ozone precursors".

10. We will change this to read: Changes in biomass burning therefore need to be taken into account in estimating the 'climate-carbon cycle feedback', i.e. the longer-term positive feedback by which global warming leads to a reduction in land carbon storage, a consequent reduction in the net uptake of CO_2 so that more CO_2 remains in the atmosphere, and thus an amplification of the initial warming (Arora et al., 2013; Cox et al., 2013; Wenzel et al., 2014)'.

11. The use of a semi-colon here is grammatically correct.

12. We can add a comma here to increase readability.

13. We can remove the comma here to increase readability.

14. We will indicate that this is the GHCNMv2 dataset. This was created from more than >7000 stations worldwide and provides a historical records since 1901. The construction of the data set is described by Peterson and Vose (1997) and we will add this reference to the text.

15. We have corrected this typo.

16. Thanks. This should be c_t^*.

17. We will provide the information about the sections in the Supplementary Information where R code and/or relationships are documented. We will also add references to the appropriate sections in the SI for other material cited in the text.

18. Figure 2a shows the relationship between emission and land temperature anomalies, and we will make that clear in the text.

19. We will add a definition of climate sensitivity at the first use of this term in the text in Section 2.7 (i.e. 'the global mean surface temperature change for a doubling of CO_2 concentration').

20. We will correct this typo. Thanks for spotting it.

21. We assume that this refers to Figure 4c, which parallels Figure 2a but for the palaeodata. And again, yes, this shows the regression between anomalies. We will clarify this in the text.

22. Fig. 4: We will provide a new version of this figure with Y-axis labels on all three panels.

Response to Referee #4

The relationship between temperature and charcoal has been established in previous studies, but we agree that establishing the quantitative relationship between charcoal and the ice-core methane and methane-isotope record is an important additional piece of information. As a result of comments by Reviewer #1, we will expand the discussion of previous studies on the charcoal-temperature relationship in the Introduction, and we will also stress the importance of the quantitative relationship between charcoal and methane in the discussion.

We were at pains to point out the relationship between temperature and emissions over the satellite era is not robust, and that it becomes non-significant if deforestation and peatland fires are not taken into consideration. It is clear that other factors, including the impact of human fire suppression, have had an overwhelming impact on fire during recent decades. Our goal here however is not to investigate the regional controls on fire (the subject of a number of recent papers), whereas our emphasis on testing for a temperature-fire relationship is necessary in order to estimate the global feedback strength. We have included the satellite-era analysis here for completeness, but we hope that it is clear from the discussion in the paper that the more robust estimate of the feedback is based on the palaeodata.

It is true that including pre-2000 data in the regression produces a negative slope. We omitted these data, however, because the pre-MODIS era data are thought to be much less reliable since they are derived from VIRS and ASTR active fire counts via optimization against the

post-2001 MODIS data (see response to Sam Rabin, Referee #3). As we stress in the paper, even after eliminating these early (anomalous and less reliable) data points, the relationship we find is barely significant and becomes non-significant if peatland and deforestation fires are omitted.

We agree that an examination of the relationship between palaeodata and temperature at a regional scale could provide additional corroboration for the global relationship. Such analyses have already been done e.g. by Marlon et al (2013) for the data-rich regions of North America, Europe and southeast Asia. In all cases they showed a positive relationship between temperature and charcoal abundance. We have now performed separate analyses for broad latitudinal bands (see our response to Referee #1) and propose to add these in section 3.4 of the paper.

This referee's comments highlight a key point that should be clarified in our revised manuscript. We are *not* claiming that fire responds only to temperature. We are well aware that this is not the case (see e.g. Bistinas et al., 2014, which is cited in the text along with other analyses of the multivariate controls on fire). However, we argue that if other factors are properly taken into account, the relationship between fire (and fire emissions) and temperature is positive. We did not say, nor do we mean to imply, that the relationship between fire emissions and temperature was positive in the pre-industrial epoch, became negative in recent decades and will become positive in the future. The lack of a significant relationship between fire emissions and temperature during the post-2000 interval, and the observed decrease in fire over recent decades (e.g. Andela et al., 2017) while climate has been warming, point to the increased influence of other controls on the fire regime.

Response to Referee #5

1. We agree that our general assumption is that warming will lead to increased fire. However, we do not claim that this is the only factor influencing fire. Analyses of satellite-era data, cited in the paper (e.g. Krawchuk et al., 2009; Bistinas et al., 2014), show that other factors play a role but that that the impact of temperature, when these other factors are taken into account, is strong and positive. On palaeo-timescales, the paper by Daniau et al. (2012, also cited in the text) shows that globally the influence of temperature is positive whereas changes in moisture lead to an increase in regions where increased moisture improves fuel loads and a decrease where increased moisture creates a situation where the fuel is too wet to burn. Analyses by Marlon et al. (2013) considered the impact of climate on regional patterns, and showed that the strength of the relationship with temperature varied regionally but was always positive. We will cite this paper and will add regional analyses of the charcoal-temperature relationship to this paper (see response to Referee #1). We will also expand the Introduction to make it clear that our focus here on temperature is because we are *assessing the magnitude of the global fire feedback*, and not analysing the relative importance of the multiple controls on fire.

2. The analyses of the satellite-era data are inconclusive for many reasons. We discuss the limitations of the data, but we could have gone further into this aspect – for example, there is substantial disagreement between burnt area among different satellite data products, and certain trends that are apparent in GFED4 are not present in alternative data sets (e.g. cci). We agree that it is possible that the influence of temperature variability on interannual timescales might be different from its influence on decadal-to-millennial timescales, but we cannot establish this from the palaeodata because there is too little annually-resolved information and the interval for which we have satellite data is too short to be able to

investigate even decadal variability. Again we should stress the difference between apparent responses to a single variable and the underlying relationship when all factors are taken into consideration. We therefore propose to expand the discussion of the controls on fire, including the evidence from previous palaeo-studies in the Introduction (see response to Referee #1). We will also expand the discussion of the limitations of the satellite-era data, and expand on our brief mention of potential differences between inter-annual and longer-term responses in the Discussion.

3. We excluded agricultural fires on the assumption that these are set by humans during suitable short-term weather conditions, and that their incidence, timing and size are unrelated to climate or other environmental conditions. They also represent a very small contribution to total fire emissions. We will add a sentence to explain this in the text.

4. We will define the variable name N_t in the text.

5. 1750 CE marks the start of the nearly monotonic rise in atmospheric CH_4 concentration towards the present day and, with it, a trend towards less negative $\delta^{13}C$. This date also marks the beginning of the Industrial Revolution and thereafter there is increasing scope for human alteration of CH_4 sources and their isotopic signatures, e.g. through expansion of grazing and human modification of fire patterns in the first instance, and the direct input of fossil-fuel derived CH_4. See e.g. KR Lassey et al.: *Atmospheric Chemistry and Physics* 7: 2119–2139, 2007 and S Houweling et al.: *Global Biogeochemical Cycles* 22: GB1002, 2008. There is still no generally accepted account of the causes of variations in CH_4 and its isotopes from 1750 onwards. This is not surprising, given that the data record only two quantities, whereas the possible variations in sources are many.

Response to Vivek Arora's comments.

The derivation of emissions from normalized charcoal data is based on the fact that we have first established a good relationship between the charcoal normans and the methane record (which is a more direct measure of fire emissions). It is clear from several reviewers' comments that we needed to make the logic of our approach clearer and we will therefore (a) include a paragraph at the beginning of the methods section to spell out the steps involved, (b) include a flowchart as a new figure in the methods section to illustrate the methodology for both the satellite era and the palaeo-era, (c) expand the discussion of previous studies linking changes in charcoal to changes in temperature while emphasizing the importance (and novelty) of establishing a qualitative relationship via the methane record.

We have used Marlon et al. (2016) because this represents the latest version of the charcoal database (version 3). Marlon et al. (2008) used version 1 of the database. The new dataset has almost double the number of sites (736 sites versus 406 sites), including sites in regions that were relatively poorly sampled before. It therefore represents a significantly better constrained picture of changes in fire over the last millennium and the extra data will naturally improve the reliability of the charcoal indices compared to version 1 of the database. It therefore doesn't really make sense to test how much of a difference this would make to the results presented here. However, we will add a sentence in the description of this data set to make it clear that the new version is an improvement, both in terms of number of sites and spatial representivity, compared to previous versions of the database.

We chose to report the feedback in ppm/degree Celsius because this facilitates the calculation of gain. The sign is opposite because gamma refers to change in land carbon and we are focusing on change in atmospheric carbon. We converted Pg/C to ppm by first dividing by 2.12 (the simple unit conversion) and multiplying by the airborne fraction. We will spell out the logic and the conversion in the results section where we make the comparison with the results of your study.

We included the analysis of the satellite-era data because so much of the analyses of fire patterns, trends, relationships with drivers focuses solely on this period. However, we were at pains to point out in the original manuscript that (a) the results are only barely significant because the records are too short, and (b) that they become insignificant if peatland and deforestation fires are not taken into account. It is worth bearing in mind that tropical peatland and deforestation fires, while anthropogenic in origin, are strongly influenced by climate variability. Although we include the satellite-era analysis for completeness, we hope that in our revised discussion we can make it clear that (a) the similarity of the gain estimated for this period and the palaeo period is entirely fortuitous and (b) that only the palaeodata provides a robust estimate of the fire feedback.

The derivation of $\partial C/\partial T$ is described in the text, specifically "Following the convention established by Hansen et al. (1984), gain (g) is the product of the feedback strength and the climate sensitivity (i.e. the global mean surface temperature change for a doubling of CO_2 concentration) expressed in K ppm^{-1}. However, we propose to add the equation to the flowchart that we will use to illustrate the methodology, and which will then be referred to very early in the text.

Reponses to minor comments

Abstract, line 25: We gave the climate sensitivity that we actually used in calculating the gain in the abstract.

Page 3, lines 25-26: Yes, of course the emissions are derived from GFED4s. We can rephrase this to make it clear that we are describing the estimates that we used in our analysis here.

Page 4, line 6: We can clarify this as: where the c_t^* are the optimally Box-Cox transformed influx values from a particular record at time t and $\bar{c}^*$ is the mean transformed influx for that record over the interval 1–1700 CE (the transformation and normalization base period).

Page 4, lines 12-13. All we meant here was that we used the published age models and did not attempt to construct age models ourselves. We will rewrite this as: We used the published age models for each record.

There is an offset between the values obtained for the Northern Hemisphere and the Southern Hemisphere records. In order to produce a global composite, it is necessary to deal with this and we have followed Separt et al.'s recommendation for how to do this. We will clarify this in the text.

Page 5: Pseudoreplication is the process of artificially inflating the number of samples or replicates, giving a false sense of sample size, which creates problems for statistical testing. Temporal pseudoreplication occurs when there is a temporal relationship between serially adjacent samples or replicates (i.e. the samples or replicates could be measured multiple

times). This would arise if we sampled a continuous charcoal record at too close an interval. We added an in-line definitition in the text.

Page 6, line 10: As we state in the text: Equation 4 can then be resolved into the sum of three components: a constant intercept, a component proportional to M, and a component proportional to the product δM. Perhaps what you have missed here is the intercept.

Page 6, lines 26-27: We will clarify the alternative conventions used for feedback and gain, and we will explicitly add a reference to the Appendix giving the derivation of equation 6 at the point that we introduce this equation.

Equation 6: We used increase here originally because the whole sentence was framed in terms of the impact of an increase in atmospheric CO_2 concentration, but we agree that it would be better to express this generically as change throughout.

Page 7, line 8: We will clarify that these are normalized charcoal anomalies.

Page 8, line 6. We can change "variable in sign" to "are both positive and negative", to make this clearer.

Page 8, line 29-390. F is the significance level and df the degrees of freedom. We can clarify the conventional statistical terminology here, and these calculations are also described in full in the Supplementary Information.

Specific changes

1) expansion of discussion of relationships between charcoal and fire controls, including expansion of citations
2) expansion of discussions about regional relationships between temeperature and fire, including analyses of broadscale latitudinal bands (Section 3.4)
3) clarification of the fortuitous agreement between satellite era and palaeo data estimates of feedbacks and gains
4) clarification that geologic flux is not used in our calculations
5) clarification of citation of feedback estimates from Arora et al.
6) expansion of discussion of the Ward et al paper and clarification of associated uncertainties
7) increased emphasis of the novelty of the proof that charcoal is a valid records of biomass burning on palaeo-rimescales through comparison with methane records
8) increased emphasis on our analyses of the robustness of the MODIS data analysis
9) creation of flow chart to document methodology, including adding the explicit equations for each step
10) expansion of SI to include additional information about the analysis of the modern data to establish robustness
11) revision of SI so that it is easy for the use to follow, and additional references to explicit SI sections in the main text
12) provision of new version of Figure 4
13) clarification for the exclusion of agricultural fires in our analyses
14) minor corrections to references and wording throughout

The biomass burning contribution to climate-carbon cycle feedback

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Abstract. Temperature exerts strong controls on the incidence and severity of fire. All else equal, warming is expected to increase fire-related carbon emissions, and thereby atmospheric CO_2. But the magnitude of this feedback is very poorly known. We use a single-box model of the land biosphere to quantify this positive feedback from satellite-based estimates of biomass burning emissions for 2000–2014 CE, and from sedimentary charcoal records for the millennium before the industrial period. We derive an estimate of the centennial-scale feedback strength of 6.5 ± 3.4 ppm CO_2 per degree of land temperature increase, based on the satellite data. However, this estimate is poorly constrained, and is largely driven by the well-documented dependence of tropical deforestation and peat fires (primarily anthropogenic) on climate variability patterns linked to the El Niño-Southern Oscillation. Palaeodata from pre-industrial times provide the opportunity to assess the fire-related climate-carbon cycle feedback over a longer period, with less pervasive human impacts. Past biomass burning can be quantified based on variations in either the concentration and isotopic composition of methane in ice cores (with assumptions about the isotopic signatures of different methane sources) or the abundances of charcoal preserved in sediments, which reflect landscape-scale changes in burnt biomass. These two data sources are shown here to be coherent with one another. The more numerous data from sedimentary charcoal, expressed as normalized anomalies (fractional deviations from the long-term mean), are then used – together with an estimate of mean biomass burning derived from methane isotope data – to infer a feedback strength of 5.6 ± 3.2 ppm CO_2 per degree of land temperature and (for a climate sensitivity of 2.8 K) a gain of 0.09 ± 0.05. This finding indicates that the positive carbon cycle feedback from increased fire provides a substantial contribution to the overall climate-carbon cycle feedback on centennial time scales. Although the feedback estimates from palaeo and satellite-era data are in agreement, this is likely fortuitous because of the pervasive influence of human activities on fire regimes during recent decades.

1 Introduction

Fire is a natural, recurring event in most terrestrial ecosystems. About 4% of the global land area is burnt every year (Giglio

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et al., 2013), resulting in global CO_2 emissions of around 2 PgC per year (van der Werf et al., 2010), substantial contributions to the budgets of other direct or indirect greenhouse gases (including CH_4, CO, N_2O, and ozone precursors), and further contributions to the atmospheric aerosol loading (black carbon, organic compounds). Climate-induced interannual variability in biomass burning, particularly variability associated with the El Niño-Southern Oscillation (ENSO),

5   is an important component of the interannual variability of the atmospheric CO_2 growth rate (van der Werf et al., 2010). However, changes in biomass burning also occur in response to longer-term climate variability and trends (Marlon et al., 2008; Power et al., 2008; Mooney et al., 2011; Daniau et al., 2012; Marlon et al., 2013). Changes in biomass burning therefore need to be taken into account in estimating the 'climate-carbon cycle feedback', i.e. the longer-term positive feedback by which global warming leads to a reduction in land carbon storage, a consequent reduction in the net uptake of

10   CO_2 so that more CO_2 remains in the atmosphere, and thus an amplification of the initial warming (Arora et al., 2013; Cox et al., 2013; Wenzel et al., 2014). The dominant terms in the terrestrial carbon balance are gross primary production and total ecosystem respiration. The climate-carbon cycle feedback is generally attributed to the temperature-dependent balance of these two large annual fluxes (Keenan et al., 2016; Ballantyne et al., 2017; Jung et al., 2017); but this neglects the potential contribution of biomass burning, which we consider here.

15   Although there have been attempts to quantify the contribution of deforestation fires (Bowman et al., 2009) or the aerosol-related component of biomass burning (Arneth et al., 2010), the global-scale contribution of biomass burning to the climate-carbon cycle feedback has been quantified only once (Ward et al., 2012). That study reported a variety of feedbacks based on simulations using a single Earth System Model (ESM). Ward et al. (2012) found that the simulated total climate feedback due to fire was negative, but their conclusion rested mainly on a large (and highly uncertain: Boucher et al., 2013; Carslaw et

20   al., 2013; Lee et al., 2016) indirect aerosol effect that exceeded the simulated fire feedback through the carbon cycle. In contrast, Arneth et al. (2010) estimated the aerosol feedback from biomass burning to be small and of uncertain sign.

Remotely-sensed observations of biomass burning offer a uniquely detailed global perspective on fire regimes. However, they cover only a limited period, and our ability to use these records to derive an empirical estimate of the biomass-burning contribution to the carbon-cycle feedback is further compromised by the complexity of the controls on fire. Climate

25   influences the occurrence and magnitude of fires on daily to seasonal time scales; both climate and fire affect vegetation productivity and hence the availability of fuel on yearly to decadal timescales; and human activities increase ignitions, but more importantly decrease fuel availability and fire spread (Bistinas et al., 2014: Knorr et al., 2014; Andela et al., 2017). Analyses of the independent influence of individual controls, when other factors are held constant, show that burnt area and biomass burning is extremely sensitive to, and positively correlated with, spatial and temporal variations in global

30   temperature (Krawchuk et al.. 2009; Daniau et al., 2012; Bistinas et al., 2014). Regional analyses also show positive relationships between biomass burning and temperature, although the strength of this relationship relative to other controls on fire varies between regions (see e.g. Marlon et al., 2013). The overwhelming importance of temperature for fire regimes is

unsurprising given that temperature changes influence atmospheric circulation patterns, and are closely tied to changes in precipitation and surface climates (Held and Soden, 2006; Li et al., 2013). The positive relationship between temperature and fire at global and regional scales suggests that the contribution of fire to the climate-carbon cycle feedback is likely to be positive. Yet burnt area has declined over the last decade. This decline has been attributed to the effects of fire suppression

5   and landscape fragmentation outweighing the influence of climate-induced changes in biomass burning (Andela et al., 2017).

The use of palaeoclimate records obviates the problem of limited record length and avoids those various human influences that have been so large as to dominate the fire record over at least the past 150 years (Marlon et al., 2008). Ice cores provide direct evidence for past changes in atmospheric composition, and the concentration and stable carbon-isotope composition of

10   methane (CH_4) in ice cores have been used together to reconstruct changes in biomass burning during the last millennium: see Rubino et al. (2016) for a review. CH_4 is released during the smouldering phase of fires, roughly in proportion to total CO_2 emission (Andreae and Merlet, 2001). Although this process is a relatively minor contributor to total atmospheric CH_4, it disproportionately influences the ^{13}C content of CH_4 because pyrogenic CH_4 carries the isotopic signature of photosynthesis. This is much less negative than that of the dominant (microbial) sources of CH_4 (Barker and Fritz, 1981).

15   But measurements of the ^{13}C content of CH_4 in ice cores are currently available with limited temporal resolution, and are subject to large uncertainties in the isotopic fractionation factors for different CH_4 sources. The abundance of sedimentary charcoal provides an alternative and more direct measure of relative changes in biomass burning (Power et al., 2008; Harrison et al., 2010), and has been shown to mirror changes in biomass-burning CH_4 (Wang et al., 2010). Sedimentary charcoal data are far more numerous than ice-core isotopic records for the last millennium. If it is possible to establish a

20   quantitative relationship between charcoal abundance and biomass-burning CH_4, it should then be worthwhile to exploit the greater number and temporal resolution of these records to quantify the fire contribution to the carbon-climate feedback. This is the approach we adopt in this paper. We use a single-box model of the land biosphere to derive an estimate of the contemporary biomass burning contribution to the climate-carbon cycle feedback using remote sensing-based estimates of biomass burning carbon emissions for the interval 2000–2014 CE. We then demonstrate that the charcoal and methane

25   records of biomass burning during the pre-industrial Common Era (1–1700 CE) are in good agreement. Finally, we exploit a good correlation of normalized anomalies of global charcoal abundance with global land temperatures during the last millennium to derive an alternative estimate of the strength of the climate-carbon cycle feedback.

2 Data and methods

We use a single-box model of the land biosphere to quantify the fire feedback, making separate calculations of the feedback

30   strength and gain for the satellite-era and the pre-industrial period (Fig. 1). Feedback strength is measured as the increase in atmospheric CO_2 concentration (ppm) per degree increase in temperature (K). In the satellite era, we use the relationship between the satellite-derived fire emissions and land temperature to estimate feedback strength, with a correction for the fact

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that land temperature variations are stronger than global mean temperature variations. We then convert feedback strength to gain assuming well-founded values for the total biomass, airborne fraction, climate sensitivity and atmospheric CO_2 concentration (Fig. 1a). We follow the same approach for the pre-industrial era (Fig. 1b), but using sedimentary charcoal data to estimate variations in fire emissions. Use of the sedimentary charcoal data in this way is predicated on our demonstration here that there is a strong statistical relationship, conforming to an expected mathematical form, between the charcoal data and the ice-core record of atmospheric methane and its stable carbon-isotope composition.

2.1 Remotely sensed burned area and carbon emissions

Burnt area and carbon emissions for 2000 to 2014 were derived from the GFED4s database (Randerson et al., 2015, http://www.geo.vu.nl/~gwerf/GFED/GFED4/). GFED4s provides monthly burnt area estimates on a 0.5° spatial grid from 1997 through 2014, but prior to August 2000 these observations were derived by calibrating ATSR and VIRS active fire data against MODIS-derived burnt area (van der Werf et al., 2017). We therefore exclude the pre-MODIS period 1997 to 1999 because of the large uncertainties in the burnt-area and emission estimates (Giglio et al., 2013). We also test whether the retention of the mixed-source estimates for 2000 (with only 5 months from MODIS) has an impact on the results (Supplementary Information, Section 8). Carbon emissions in GFED4s are divided into source sectors: savannah, grassland and shrubland fires; boreal forest fires; temperate forest fires; deforestation fires; peatland fires, agricultural fires. The estimates we use for total fire emissions include all of these sectors except agricultural fires. We exclude agricultural fires on the assumption that these are only set by people and therefore that the incidence, timing and size of these fires are unrelated to climate or other environmental factors. We also use an estimate of the total emissions from natural sources, that is, also excluding deforestation and peatland fires. Global mean land temperatures for this period, for comparison with the fire data, were taken from the NOAA Merged Land Ocean Global Surface Temperature Analysis (NOAAGlobalTemp v4.0.1, doi:10.7289/V5FN144H; Vose et al. 2012):
(https://www.ncdc.noaa.gov/data-access/marineocean-data/noaa-global-surface-temperature-noaaglobaltemp, with specific data found at http://www1.ncdc.noaa.gov/pub/data/noaaglobaltemp/operational/).

2.2 Charcoal data

The sedimentary charcoal data were obtained from version GCDv3 of the Global Charcoal Database (Marlon et al., 2016). This new version of the database contains considerably more individual sites records than previous versions, and provides better spatial coverage. Charcoal data were read directly from the database file GCDv03_Marlon_et_al_2015.mdb. The data were processed using the protocol described in Power et al. (2010) and Blarquez et al. (2014) except that the transformed charcoal influx values (or their equivalents) were expressed as normalized anomalies (normans, N_t at time t) or

$$N_t = (c_t^* - \bar{c}^*)/\bar{c}^* \tag{1}$$

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where the c_t^* are the optimally Box-Cox transformed influx values from a particular record <u>at time t</u> and $\bar{c}^*$ is the mean transformed influx for that record over the interval 1–1700 CE (the transformation and normalization base period). A 10-yr interval was used for pre-binning the 633 records used for the creation of the composite curve.

2.3 Methane concentration and stable carbon isotope data

Methane concentration data were taken from the composite Law Dome records (Etheridge et al., 2010). We used a composite data set of $\delta^{13}C$ of CH_4 from Ferretti et al. (2005), Mischler et al. (2009) and Sapart et al. (2012). We <u>used the published</u> age models for each record. We then applied the 0.51‰ correction described by Sapart et al. (2012) to the Northern Hemisphere data<u>, in order to create the global composite</u>.

2.4 Global palaeotemperature data

We calculated annual area-weighted averages of mean annual temperature anomalies for land grid points, using the 5º gridded data set of Mann et al. (2009), which covers the interval from 500 through 2006 CE. We used a base period of 1961– 1990 CE to calculate anomalies. We did not use the global average of the PAGES 2k Consortium (2013) because this reconstruction is dominated by records from the Arctic and Antarctic, where there are few or no fires, prior to 800 CE. Although there are many last-millennium temperature reconstructions for the northern hemisphere, global data sets are few and the rest cover shorter time intervals than Mann et al. (2009).

2.5 Composite curves of charcoal, $\delta^{13}C$ of CH_4, CH_4 and palaeo-temperature data

The individual charcoal records have a median sampling interval of 16.75 years over the interval 1–100 CE (with 250 sites contributing data), and 16.90 years over the interval 1601-1700 CE (350 sites), for a typical sample density of over 1000 per century. The $\delta^{13}C$ of CH_4 and CH_4 records average 2.5 and 3.0 samples per century over the interval 1–500 CE, increasing to 10 per century over the interval 1601-1700 CE. The temperature data have annual resolution. Consequently, for the regression analyses we developed composite (across sites, in the case of charcoal) or smoothed curves (for the other variables) with a common sampling interval, and an appropriate smoothing-window for each series. We used the R package locfit (R Core Team, 2016; Loader, 2013) to fit these curves.

Data smoothing can induce spurious cross-correlations between series (Loader, 1999; Granger and Newbold, 1986), while using an overly high-resolution sampling interval can create temporal pseudoreplication<u>, whereby sequential observations do not provide independent information </u>(Hurlbert, 1984). Both could inflate the apparent significance of relationships among series. We chose the sampling interval and smoothing window by examining diagnostic checks of the regression analyses of charcoal (as the response variable) with temperature, or $\delta^{13}C$ of CH_4 and CH_4 (as predictors), attempting to minimize the autocorrelation of the residuals as a guard against pseudoreplication. This process led to the selection of a 50-year time step for evaluation of the smoothed curves. For the charcoal and temperature data, we selected a 50-year (half-width) fixed

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smoothing window, which suppresses inter-annual to decadal-scale variability in those series, while preserving longer-term variations. The $\delta^{13}C$ of CH_4 and CH_4 data are too sparse in the first part of the record to use a fixed-width smoothing window, and so we used the variable window-width or "span" approach with the span parameter equal to 0.1. This strategy led to some interpolation in the sparser parts of these records. We obtained bootstrap confidence intervals for the smoothed

5 curves. For charcoal, we used the "bootstrap-by-site" approach described by Blarquez et al. (2014), which allows the impact of the variations in the spatial distribution of the charcoal records to be assessed, and the standard approach for the other series. The R code used to produce the composite/smoothed curves is included in the Supplementary Information (Sections 2-5).

2.6 Comparison of charcoal and methane records

10 The isotopic composition of atmospheric CH_4 depends on the magnitudes and isotopic discrimination factors of different contributors to the global CH_4 budget. Thus, although variations in biomass burning emission of CH_4 are expected to influence its isotopic composition, there is not a direct correspondence between isotopic composition and the biomass burning flux. The isotopic composition of CH_4 can also be influenced by changes in the magnitude and/or isotopic discrimination of other methane fluxes, of which the microbial source (methanogenesis in wetlands and wet soils, and in

15 other anoxic environments including ruminant stomachs) dominates. Moreover, isotopic discrimination by methanogenesis shows large geographic variations, and cannot be assumed to be the same now (with widespread agricultural grazing, and draining of natural wetlands) as it was in pre-industrial times. We therefore chose to compare the CH_4 isotopic record with the charcoal record by treating the isotopic discrimination factors as unknown and using a regression approach (Fig. 1), respecting the isotopic mass balance, to test whether the two types of record are systematically related to one another. After

20 1700 CE, the relationships between charcoal and temperature, and between charcoal and $\delta^{13}C$ [CH_4] and [CH_4] become significantly distorted. Regressions were therefore fitted using composite/smoothed curve data only up to and including 1700 CE.

The mass balance equation for the principal (non-fossil fuel) annual CH_4 fluxes to the atmosphere is:

$$F \; = \; F_m + F_g + F_b \tag{2}$$

25 where F is the total flux, F_m is the microbial flux, F_g is the geological flux (natural seepage from underground gas reservoirs), and F_b is the biomass burning flux. The isotopic mass balance equation is:

$$\delta \; = \; \delta_m(F_m/F) + \delta_g(F_g/F) + \delta_b(F_b/F) - \varepsilon \tag{3}$$

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where δ is the isotopic signature ($\delta^{13}C$) of global atmospheric CH_4, δ_m, δ_g and δ_b are the isotopic signatures of the microbial, geological and biomass burning sources respectively and ε is the isotopic discrimination of CH_4 oxidation in the atmosphere and soils. Re-arrangement of equations (2) and (3) yields:

$$F_b = F(\delta - \delta_m + \varepsilon)/(\delta_b - \delta_m) - F_g(\delta_g - \delta_m)/(\delta_b - \delta_m). \tag{4}$$

5   The total flux F is related to the global CH_4 concentration M in steady state by $F = fM/\tau$ where f is the conversion factor between atmospheric concentration and mass and τ is the atmospheric lifetime of CH_4, which we assume to be constant. The geological flux can also be assumed constant, although its magnitude is disputed (Schwietzke et al., 2016; Petrenko et al., 2017). The steady-state assumption is appropriate because we are considering variations over periods longer than the atmospheric lifetime of CH_4, approximately 9 years (Schwietzke et al., 2016). Equation 4 can then be resolved into the sum

10   of three components: a constant intercept, a component proportional to M, and a component proportional to the product δM. Equation (4) also holds, with appropriate adjustment of units, if the F_b are expressed in normans; then all of the fluxes are relative to the mean value of F_b. We used ordinary linear regression of charcoal normans with M and δM as predictors to quantify the relationship between the charcoal data and CH_4 isotopic composition. The inclusion of CH_4 concentration in this analysis is essential, because variations in δ could be brought about irrespective of biomass burning by variations in F_m,

15   which is generally much larger than F_b.

2.7 Calculation of feedback strengths and gain

The global relationship between biomass burning CO_2 emissions and temperature provides an estimate of the strength of the feedback. We define feedback strength as the equilibrium sensitivity of atmospheric CO_2 to global land temperature in ppm K^{-1}. This can be further converted to gain (Lashof et al., 1997). Following the convention established by Hansen et al.

20   (1984), gain (g) is the product of the feedback strength and the climate sensitivity (i.e. the global mean surface temperature change for a doubling of CO_2 concentration) expressed in K ppm^{-1}. Then the temperature amplification $\Delta T/\Delta T_0$, where ΔT is the actual temperature change and ΔT_0 is the reference temperature change without the feedback, is:

$$\Delta T/\Delta T_0 = 1/(1 - g) \tag{5}$$

Note that this convention (Hansen et al., 1984) is widely applied in the literature on terrestrial biogeochemical feedbacks.

25   However, an alternative convention exists in which the quantity defined in equation (5) is called the gain, while the quantity we call gain is called the feedback factor (see e.g. Roe, 2009).

The equilibrium sensitivity of atmospheric CO_2 concentration to a change in the biomass burning flux was estimated using a box model, with parameters derived from either present-day or palaeo-relationships. The principle is that an increased rate of

removal of land carbon due to fire results in a reduced steady-state carbon storage and a correspondingly increased atmospheric CO_2 content. The change in atmospheric CO_2 concentration is given to a good approximation by:

$$\Delta C \approx (W/\text{NPP}) \, \Delta F_b \, \text{AF}/2.12 \qquad (6)$$

where ΔC is the change in atmospheric CO_2 concentration (ppm), W is total land ecosystem carbon storage (Pg C), NPP is total land net primary production (Pg C a^{-1}), ΔF_b is the change in biomass burning carbon flux (Pg C a^{-1}), AF is the airborne fraction (the fraction of emitted CO_2 remaining in the atmosphere), and the factor 2.12 converts Pg C to ppm (http://cdiac.ornl.gov/pns/convert.html; Ciais et al., 2014). (The full derivation of equation 6 is given in the Appendix.) For the satellite era, we related ΔF_b (Pg C a^{-1}) statistically to temperature data. For the pre-industrial era, we related normalized charcoal anomalies (dimensionless) statistically to temperature data and multiplied by an estimate of the long-term mean F_b for the period up to 1600 CE (3.87 Pg C a^{-1}). This estimate was based on the calibration of the methane isotope record by Sapart et al. (2012), as follows: we multiplied the contemporary flux of 2.02 Pg C a^{-1} (the average of five satellite-based estimates from Shi et al., 2015) by the ratio of the global biomass-burning CH_4 flux inferred for 1–1600 CE (27.4 Tg CH_4 a^{-1}) to the same flux inferred from GFED4s (14.3 Tg CH_4 a^{-1}). Since feedback strength is related to timescale (Roe, 2009), we assumed an AF appropriate to the centennial time scale (Joos et al., 2013), and standard values for global net primary production and total carbon storage in vegetation, litter and non-permafrost soils. The derivation of equation (6), and details of calculations including the uncertainty propagation, are provided in the Appendix.

3 Results

3.1 Relationship between biomass burning flux and global average land temperature during the satellite era

The sensitivity of the MODIS-era biomass burning flux to temperature (Fig. 2) was obtained by regression of GFED4s annual fluxes against global (annual average) land temperature data, yielding a slope of 0.71 Pg C K^{-1} with a standard error of $\pm$ 0.34 Pg C K^{-1} (Fig. 3). Although approaching statistical significance, this relationship was weak ($R^2 = 0.25$, $p = 0.058$). The slope of the relationship however was shown to be insensitive to individual extreme years (see Supplementary Information, Section 8).

3.2 Estimation of feedback strength during the satellite era

The fitted relationship of annual biomass burning flux to temperature provides an estimate of the feedback strength of 6.5 $\pm$ 3.4 ppm K^{-1} with respect to global land temperature. We took account of the greater variability of land versus global mean temperatures by means of a regression of land versus global mean temperature anomalies for 2000–2014 (Fig. 3a), yielding a slope of 1.364 $\pm$ 0.098 K K^{-1}. Correcting the estimated land-based feedback strength with this slope yielded a corrected feedback strength of 8.9 $\pm$ 4.7 ppm K^{-1}. Assuming a value of $S = 2.8$ K, the central value for climate sensitivity recently

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obtained by a novel emergent-constraint method (Cox et al., 2018), led to $\partial T/\partial C = S/(C \ln 2) = 0.0106$ K ppm^{-1} (evaluated at $C = 380$ ppm) and an estimated gain of 0.09 ± 0.05. (The uncertainty of the gain estimate does not include the uncertainty in S, which affects all estimates of gain but does not affect comparisons of gain made with the same value of S.)

However, if deforestation and peat fires (which account for 18-28% of emissions) were excluded from the calculations (Fig. 3b), no significant relationship of biomass burning emissions to temperature remained ($p = 0.476$). Interannual variability in tropical deforestation and peatland fires is well known to be correlated with ENSO (van der Werf et al., 2010), whereas ENSO-related changes in temperature and precipitation are both positive and negative across extratropical regions – resulting in compensatory impacts on total non-anthropogenic fire emissions, which show no clear general relationship to temperature during the satellite era (Prentice et al., 2011).

3.3 Relationship between methane and charcoal records of biomass burning

The fitted regression equation relating charcoal normans (dimensionless) to the concentration of CH$_4$ (M_t at time t, ppb) and the product of the δ^{13}C of CH$_4$ (δ_t at time t, ‰) with M_t ($\delta_t M_t$, ‰ ppb) is:

$$N_t = 0.0659 + 0.001\,18\,M_t + 0.000\,046\,79\,\delta_t M_t \qquad (7)$$

($R^2 = 0.771$, $F = 54.04$ with 1 and 32 df, $p < 0.0001$). The standard errors of the fitted regression coefficients in equation (7) are as follows: ± 0.0147 for the intercept, $\pm 0.000\,70$ ppb^{-1} for the coefficient of M_t, and $\pm 0.000\,012\,37$ ‰$^{-1}$ ppb^{-1} for the coefficient of $\delta_t M_t$ (see Supplementary Information, Section 7 for more details). The Ljung-Box statistic (Ljung and Box, 1978) is 16.9 with 12 df and $p = 0.15$, i.e. not significant, indicating that pseudoreplication and the possibility of spurious correlation are absent.

This analysis shows, for the first time, that the charcoal and methane data sources (Fig. 4) are in good agreement (Fig. 5b). It is therefore appropriate to use charcoal normans (based on a global compilation, albeit with some unnevenness is sampling) as an indicator for normalized anomalies of global biomass burnt.

The ratio r of the coefficient of M_t to the coefficient of $\delta_t M_t$ could in principle provide an independent estimate of the microbial discrimination factor, as $\delta_m = \varepsilon - r$ by re-arrangement of equation (4). However, in practice this calculation does not provide a strong constraint on δ_m. Assuming $\varepsilon = -6.3$‰ (Schwietzke et al., 2016) and with $r = 25.2 \pm 16.4$‰ from equation (7), δ_m is estimated as -31.5 ± 16.4 ‰. The central estimate is small in magnitude compared to typical values around -60% (e.g. Sapart et al., 2012), but its standard error is large.

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3.4 Relationship between charcoal records and global average land temperature

The fractional sensitivity of the millennium-scale biomass burning flux to temperature was obtained by regression of charcoal normans against global land temperature. The fitted regression equation relating anomalies of charcoal normans and temperature (Fig. 5c) is:

$$N_t = -0.0205 + 0.158 \, T_t \tag{8}$$

where the N_t are charcoal normans (dimensionless) and T_t are the area-weighted average temperatures (˚C; $R^2 = 0.646$, $F = 41.98$ with 1 and 23 df, $p < 0.0001$). The standard errors of the fitted regression coefficients in equation (8) are ± 0.005 for the intercept, and ± 0.024 K^{-1} for the coefficient of T_t. The Ljung-Box statistic is 16.2 with 12 df, and $p = 0.184$, i.e. non-significant (see Supplementary Information, section 6).

Regional analyses show that the observed strongly positive global-scale relationship between temperature and normalized charcoal anomalies is mirrored in the northern extratropics, northern tropics and southern tropics (Fig. 6), but not in the southern extratropics. However the Mann et al. (2009) data set contains relatively few observations from the southern extratropics, and shows an anomalously large temperature decline from 500 to 1500 CE compared to other reconstructions (e.g. Neukom et al., 2014; Gergis et al., 2016; Supplementary Information Section 11). We reserve judgment as to whether this regional difference in the relationship is meaningful. In any case, the land area represented by the southern extratropics is small.

3.5 Estimation of feedback strength during the pre-industrial era

Applying an estimated long-term mean value $F_b = 3.87 \pm 1.94$ Pg C a^{-1} yielded $\Delta F_b = 0.61 \pm 0.32$ Pg C a^{-1} K^{-1}. The resulting estimate of feedback strength is 5.6 ± 3.2 ppm K^{-1} with respect to land temperature. A regression of land versus global mean temperatures based on the 500–1700 CE data in Mann et al. (2009) yielded a slope of 1.146 ± 0.0018 K K^{-1} (Fig. 3a). Correcting the estimated land-based feedback strength with this slope, and assuming $S = 2.8$ K as before, led to $\partial T/\partial C = S/(C \ln 2) = 0.0144$ K ppm^{-1} (evaluated at $C = 280$ ppm) and an estimated gain of 0.09 ± 0.05. The uncertainty in this value is dominated by the large uncertainty assigned to the mean pre-industrial biomass burning flux.

4 Discussion and Conclusions

Our analyses of data from the pre-industrial era yielded an estimate of the feedback strength of 5.6 ± 3.2 ppm K^{-1} for land temperature, and a gain of 0.09 ± 0.05. Our analyses for the satellite era yielded 6.5 ± 3.4 ppm K^{-1} for land temperature, and also a gain of 0.09 ± 0.05. The agreement between the two gain estimates is fortuitous, however. The pre-industrial estimate is founded on a strong relationship between charcoal data and reconstructed temperatures. Its uncertainty is largely due to

uncertainty about the absolute magnitude of average biomass burning emissions in pre-industrial time. In contrast, the uncertainty of the satellite-era estimate is largely due to the weakness of the relationship between emissions and observed temperatures. Moreover this relationship is dominated by the well-known correlation of anthropogenic burning in the tropics with the ENSO cycle. The period for which reliable satellite-based estimates of biomass burning emissions are available is too short to have allowed the effects of longer-term climate variability to emerge, especially given the uncertainties associated with the large differences between different satellite products (Hantson et al., 2016).

It is unclear whether the magnitude of the fire feedback estimated on the basis of interannual variability should be different from the estimate obtained based on decadal to centennial variability. The palaeo-record does not provide a test of this because there are too few annually resolved charcoal records, while the satellite-era records cover too short a period to be able to examine longer-term sensitivity. However, even if the satellite-era data provided a strong constraint on fire feedback, the estimate of gain based on pre-industrial, centennial-scale climate variability would likely still be more relevant to long-term climate projections.

Many of the influences on fire have changed dramatically between pre-industrial and recent times. The geographic pattern of fire frequency shows an unambiguous decline with human population density, a relationship that holds across more than four orders of magnitude of population density (Bistinas et al., 2014; Knorr et al., 2014). Moreover, global biomass burning has declined precipitously since its peak in the mid-nineteenth century, as shown by both charcoal data (Marlon et al., 2008; Marlon et al., 2016) and carbon monoxide isotopes in ice and contemporary air (Wang et al., 2010). On the other hand, tropical deforestation and burning of peat substrates yield intense, localized pyrogenic sources of CO_2 that closely covary with interannual variation in the duration and intensity of the dry season (van der Werf et al., 2010). Our estimate of gain based on pre-industrial, centennial-scale climate variability is likely more relevant to long-term climate projections, but any realistic estimation of future fire risks and feedbacks must consider the pervasive effects of human settlement and land use (Knorr et al., 2014). It is also possible that the influence of temperature variability on interannual timescales might generally differ from its influence on decadal-to-millennial timescales, but we cannot establish this from currently available palaeodata because there is too little annually-resolved information, while the interval for which we have satellite data is too short even to resolve decadal variability.

Charcoal abundances have generally been interpreted as a measure of 'fire activity' or relative changes in the quantity of burned biomass (e.g. Power et al., 2008; Harrison et al., 2010; Daniau et al., 2012; Marlon et al., 2016). There have been some attempts to quantify the relationship between charcoal abundance and burnt area or total biomass consumed at a local scale (see e.g. Peters and Higuera, 2007; Duffin et al., 2016; Leys et al., 2017). These analyses, however, show a strong dependency on vegetation type and fire regime and the need to apply calibrations accounting for charcoal source area in the same way as for the interpretation of pollen abundances (Prentice, 1985: Sugita, 1994). Such calibrations have been made for Europe (Adolf et al., 2017) but not for other regions. Our analyses establish for the first time that there is a good relationship

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($R^2 = 0.77$) between global charcoal abundance, expressed as normalized anomalies, and the methane and methane-isotopic record from ice cores. Since emissions reflect the amount of biomass consumed by fire, which in turn is influenced by area burnt and fire intensity, these analyses support the idea that the sedimentary charcoal record – when synthesized at continental to global scales – can provide quantitative evidence for changes in the global biomass-burning carbon flux. Establishing the quantitative relationship between charcoal abundance and fire emissions is key to be able to exploit the continued expansion of the spatial and temporal coverage of charcoal records (Marlon et al., 2016) to examine regional changes in fire regimes on multiple time scales.

The strength of the global land climate-carbon cycle feedback has been assessed by Arora et al. (2013) on the basis of nine CMIP5 Earth System models. Five models that do not explicitly represent fire yield feedback strengths (after converting Pg C to ppm, and multiplying by the airborne fraction) in the range 6.8 to 19.9 ppm K^{-1} with a median of 17.5 ppm K^{-1}. Of four models that do represent fire, two yield values in the same range; the other two (sharing the same land model) yield lower values but have been shown to greatly underestimate the feedback based on the observed relationship between tropical land temperatures and the rate of increase in atmospheric CO_2 concentration (Wenzel et al., 2014). Our global estimate of the biomass burning contribution as 5.6 ± 3.2 ppm K^{-1}, based on the pre-industrial period, suggests that the contribution of fire emissions to the climate-carbon cycle feedback is substantial. Our estimate may even be conservative. Sapart et al. (2012) estimated the intertemporal coefficient of variation in the biomass burning CH$_4$ flux to be 7.3% for the period 1–1600 CE, compared to only 2.9% in the charcoal anomalies.

Although some of the models in the assessment by Arora et al. (2013) included fire as an interactive process, none considered deforestation or peat fires. A substantial component of the total contemporary land climate-carbon cycle feedback appears to be attributable to anthropogenic fires in the tropics, and their spatially coherent association with ENSO variability. This is in contrast with extratropical fire regimes, which show regionally asynchronous responses to climate variability (Prentice et al., 2011); and the response of net ecosystem exchange to warming, which is asymmetrical between low and high latitudes (Wenzel et al., 2014). The importance of deforestation and peatland fires in driving fire feedback in the recent decades suggests that measures to protect tropical forests and peatlands could appreciably reduce the magnitude of the climate-carbon cycle feedback.

The climate-carbon cycle feedback is an important benchmark for ESMs. Despite growing interest in the environmental and human drivers and impacts of fire (Bowman et al., 2009; Harrison et al., 2010; Bowman et al., 2011; Fischer et al., 2016), the global-scale contribution of biomass burning to the climate-carbon cycle feedback has been poorly quantified. Our analyses provide an independent estimate of this feedback, illustrating the use of the palaeo-record to estimate Earth System quantities that may be difficult or impossible to derive from contemporary observations.

Appendix: The box model, parameter estimates and their uncertainties

In steady state, carbon inputs to biomass and subsequently (via litter production) to soil organic matter, corresponding to net primary production (NPP), must be balanced by outputs: heterotrophic respiration, R_H and biomass burning, F_b. Here we designate rates of carbon transfer by heterotrophic respiration and biomass burning respectively as k_r and k_b, such that $k_b = F_b/W$; $k_b* = F_b*/W*$ (where the asterisk denotes new steady-state values after a change in the burning rate); then $k_r = k_r* = R_H/W = (NPP - F_b)/W = (NPP - F_b*)/W*$, assuming the impact of an altered fire frequency on NPP is small compared to its effect on W (Martin Calvo and Prentice 2015). Hence, $W*/W = (NPP - F_b*)/(NPP - F_b)$ and upon re-arrangement:

$$\Delta W = -W.\Delta F_b/(NPP - F_b) \tag{A1}$$

where $\Delta W = W* - W$ and $\Delta F_b = F_b* - F_b$, or to a close approximation (given $F_b \ll NPP$),

$$\Delta W \approx -W.\Delta F_b/NPP. \tag{A2}$$

This calculation is insensitive to CO_2 effects on NPP, as an increase in NPP in steady state implies a proportionate increase in W.

Global terrestrial biosphere C is given by Ciais et al. (2014) as the sum of 450–650 Pg C (vegetation C) and 1500–2400 (soil C), i.e. 550 ± 100 Pg C and 1950 ± 450 Pg C respectively – yielding a combined uncertainty of ± 461 Pg C (18.4%) For global NPP, the two bottom-up estimates given by Prentice et al. (2001) are 59.9 and 62.6 Pg C a^{-1}, yielding a mean of 61.25 and a standard error ($n = 2$) of ± 1.35 Pg C a^{-1} (2.2%). We therefore assigned values of $W = 2500 \pm 461$ Pg C and NPP = 61.25 ± 1.35 Pg C a^{-1}.

For contemporary biomass burning C emissions (Shi et al., 2015; Table 3), five satellite-derived estimates together provide a global mean of 7391.7 Tg CO_2 a^{-1} (2.02 Pg C a^{-1}) with a standard deviation ($n = 5$) of ± 1291.2 Tg CO_2 a^{-1}, corresponding to a standard error of ± 0.157 Pg C a^{-1} (7.8%). We therefore assigned $F_b = 2.02 \pm 0.157$ Pg C a^{-1} for the satellite era. For the pre-industrial era, we estimated the long-term mean biomass burning C flux as the product of the contemporary flux of 2.02 Pg C a^{-1} (Shi et al., 2015) with the ratio of the global biomass-burning CH_4 flux inferred from methane isotope data for the period 1–1600 CE (27.4 Tg CH_4 a^{-1}) to the same flux inferred from GFED4s (14.3 Tg CH_4 a^{-1}) by Sapart et al. (2012), yielding $F_b = 3.87$ Pg C a^{-1}. However, while Sapart et al. (2012) assigned an uncertainty of only ± 2.8 Tg CH_4 a^{-1} (10%) to their estimate of global biomass-burning CH_4 flux, we inflated the uncertainty of our estimate of F_b to ± 1.94 Pg C a^{-1} (50%) in order to include additional potential sources of error, which include variability of the isotopic fractionation factors and of the emission factor for CH_4 with respect to CO_2.

For the centennial-scale airborne fraction (AF in equation 6) we adopted the estimate of 0.476 ± 0.057 (12.0%) obtained by Joos et al. (2013). This estimate was derived from multiple models performing identical pulse-response experiments. The

mean value here is the multi-model mean (converted from units of years to fractions by dividing by the time scale), and the uncertainties are one standard deviation of the variation among models. The mean value is close to the empirical estimate of 0.44 given by Ciais et al. (2014).

5 Conversion of the feedback strength ($\partial C/\partial T$) into a gain requires a further assumption about the climate sensitivity (S), defined as the equilibrium change in global mean temperature for a doubling of atmospheric CO_2. We have used $S = 2.8$ K, the central estimate provided by Cox et al. (2018).

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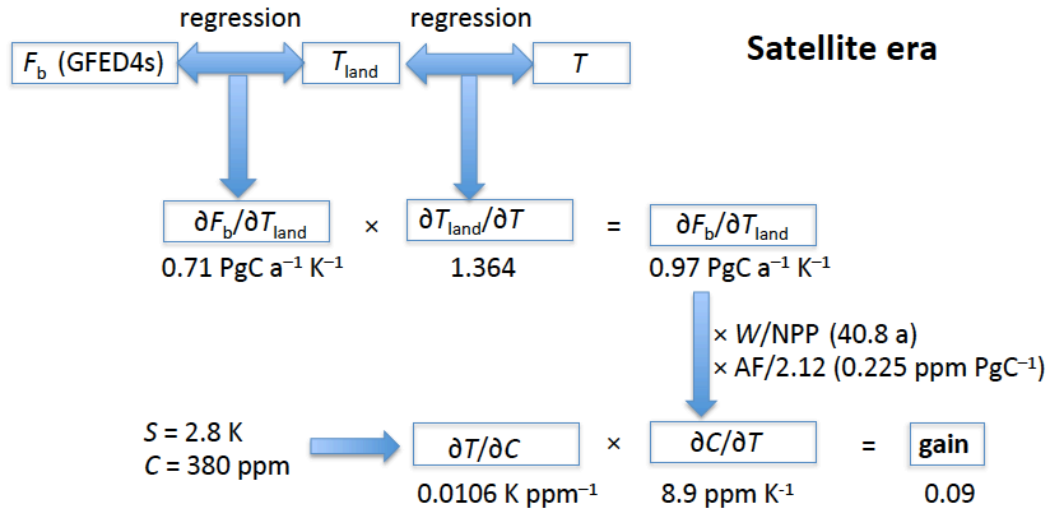
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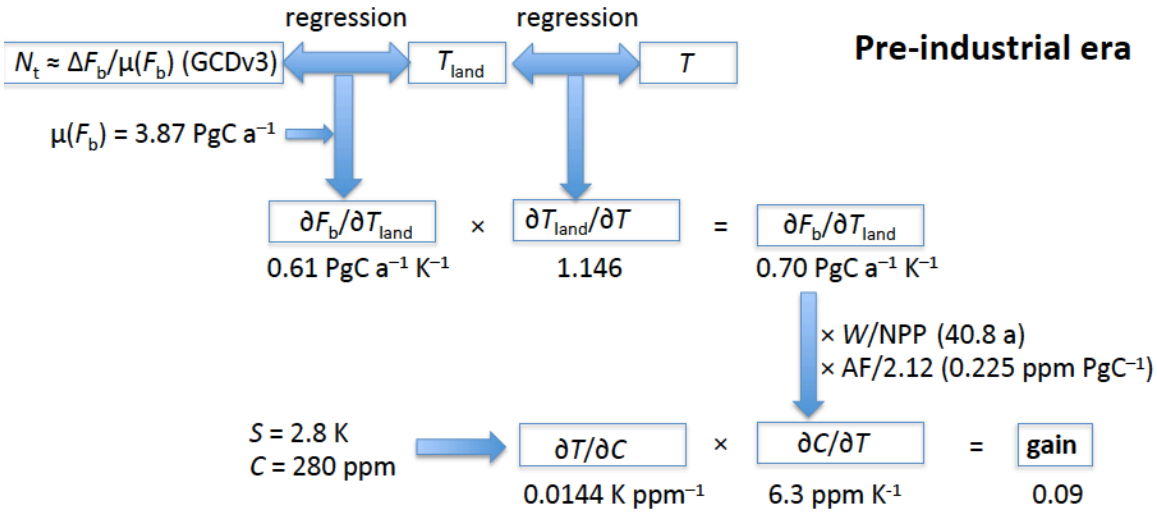
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a

regression   regression

F_b (GFED4s) ⟷ T_{land} ⟷ T

Satellite era

$\partial F_b / \partial T_{land}$ × $\partial T_{land} / \partial T$ = $\partial F_b / \partial T_{land}$

0.71 PgC a^{-1} K^{-1} 1.364   0.97 PgC a^{-1} K^{-1}

× W/NPP (40.8 a)
× AF/2.12 (0.225 ppm PgC^{-1})

S = 2.8 K
C = 380 ppm → $\partial T / \partial C$ × $\partial C / \partial T$ = **gain**

0.0106 K ppm^{-1} 8.9 ppm K^{-1} 0.09

b

regression   regression

$N_t \approx \Delta F_b / \mu(F_b)$ (GCDv3) ⟷ T_{land} ⟷ T

Pre-industrial era

$\mu(F_b)$ = 3.87 PgC a^{-1} →

$\partial F_b / \partial T_{land}$ × $\partial T_{land} / \partial T$ = $\partial F_b / \partial T_{land}$

0.61 PgC a^{-1} K^{-1} 1.146   0.70 PgC a^{-1} K^{-1}

× W/NPP (40.8 a)
× AF/2.12 (0.225 ppm PgC^{-1})

S = 2.8 K
C = 280 ppm → $\partial T / \partial C$ × $\partial C / \partial T$ = **gain**

0.0144 K ppm^{-1} 6.3 ppm K^{-1} 0.09

Figure 1: Schematic of the analysis of global fire-temperature relationships for the (a) satellite and (b) pre-industrial eras. F_b, biomass burning flux; T_{land}, global mean land temperature; T, global mean temperature. W, global land carbon storage; NPP, global net primary production; W/NPP = residence time of land carbon. AF, airborne fraction; 2.12 = conversion factor from ppm to Pg C. S, climate sensivity; C, atmospheric CO$_2$ mole fraction. N_t, normalized anomalies of charcoal data; ΔF_b, biomass burning flux anomalies; $\mu(F_b)$, mean biomass burning flux. See text for sources of temperature data.
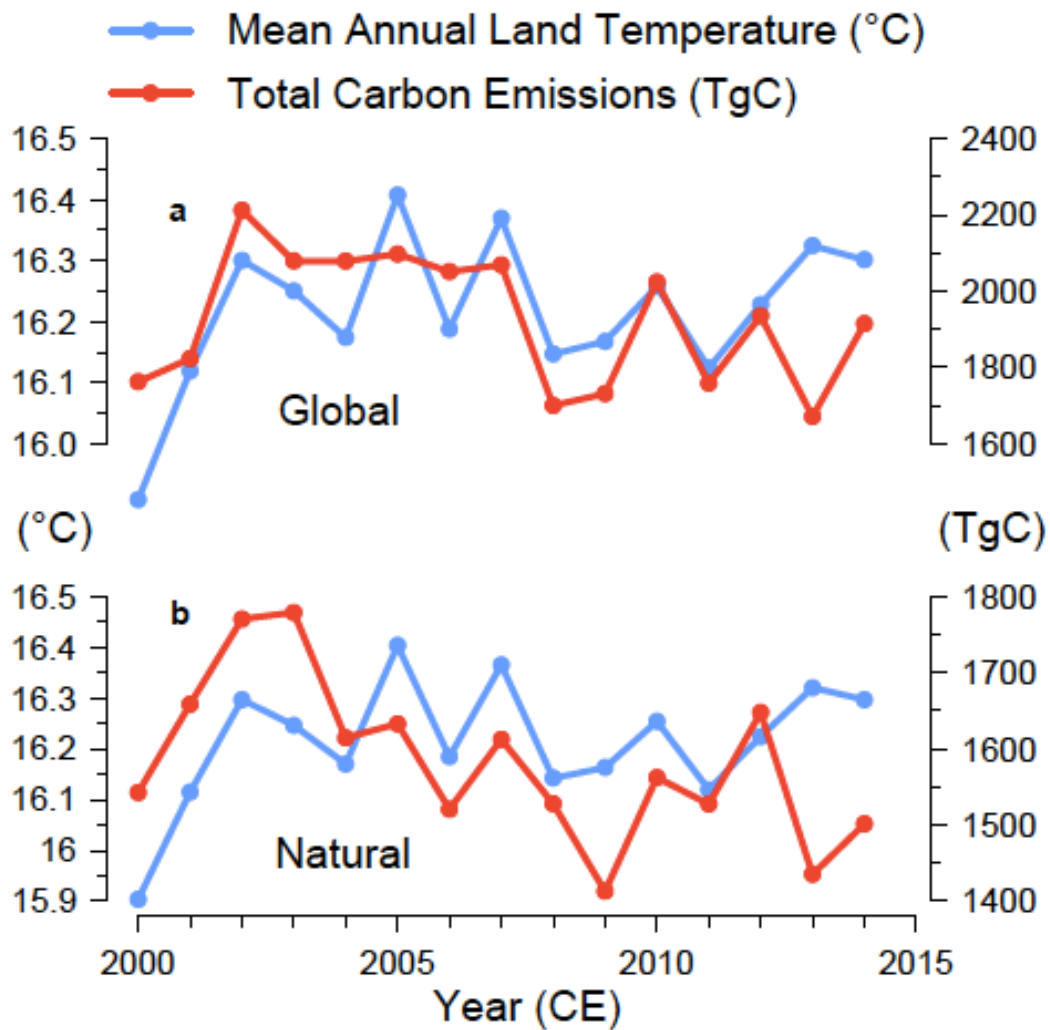
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Figure 2: Co-evolution of temperature and fire-related emissions over the period between 2000 and 2014. The temperature data are from the NOAA data set (NOAAGlobalTemp_v4.0.1; doi:10.7289/V5FN144H; Vose et al. 2012) and the emissions data are from GFED4 (Randerson et al., 2015, www.globalfiredata.org). The top panels show global (a) temperature and (b) emissions after excluding agricultural areas; the bottom panels show (c) temperature and (d) emissions from areas of natural vegetation only, excluding both deforestation fires and peatland fires.

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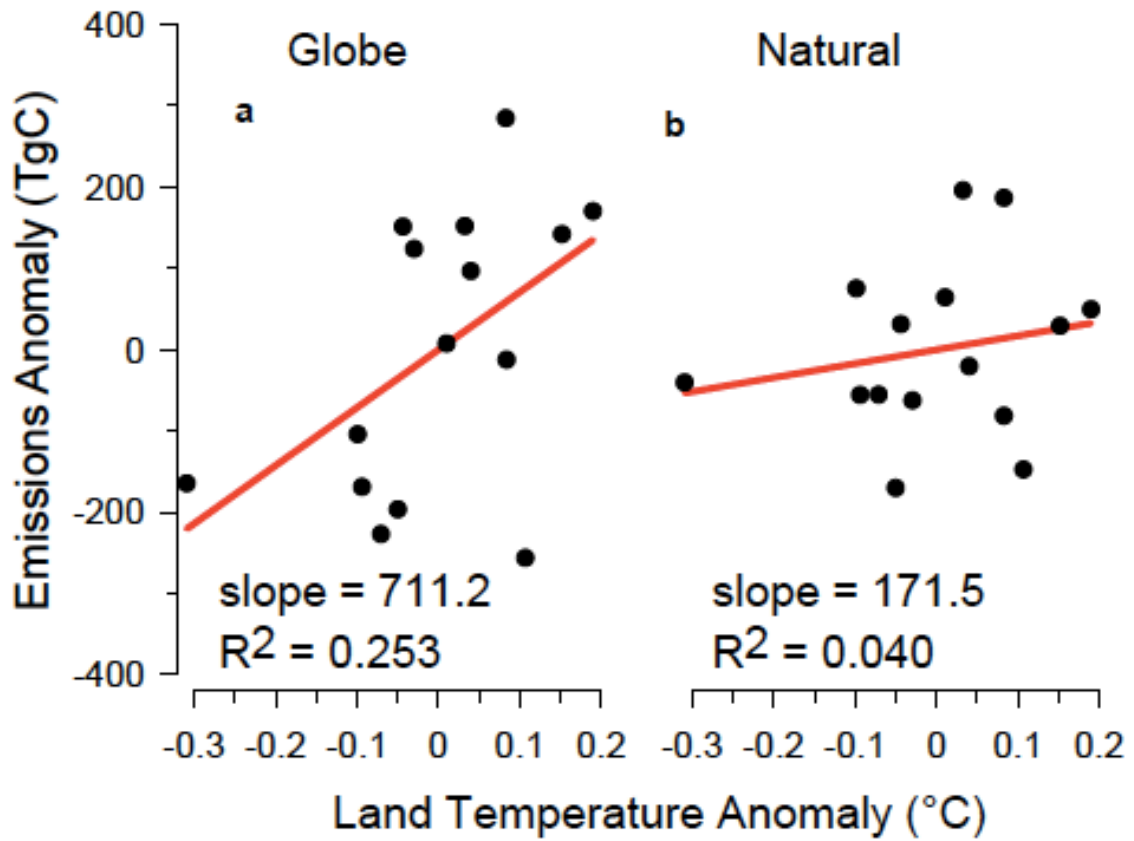
Figure 3: Relationship between global fire-related emissions and temperature over the period between 2000 and 2014. The left-hand panel shows the relationship between global temperature and emissions after excluding agricultural areas; the right-hand panel shows the relationship between temperature and emissions from areas of natural vegetation only, excluding both deforestation fires and peatland fires.
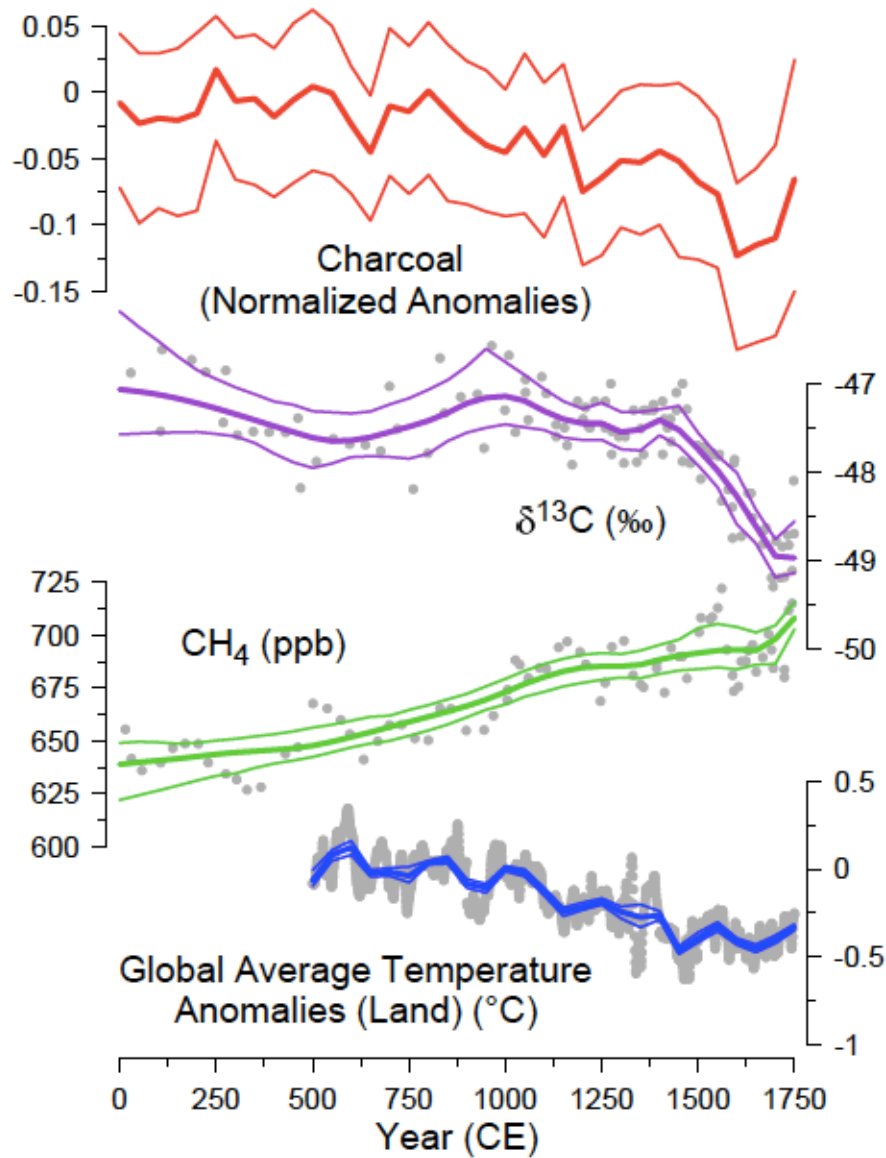
Figure 4: Indices of pre-industrial global biomass burning trends, 0–1750 CE: (a) normalised charcoal anomalies, (b) $\delta^{13}C$ of CH_4 (‰) based on a composite of the data from Ferretti et al (2005), Mischler et al. (2009) and Sapart et al. (2012), and (c) CH_4 concentration (ppb) from Etheridge et al. (2010). The bottom plot shows global average temperature anomalies over land (°C) from Mann et al. (2009). The plots show the 50-year smoothed record for each indicator, with 95% bootstrap confidence intervals; the individual data points for $\delta^{13}C$, CH_4 and land temperature are shown by grey points. There are too many individual charcoal points to be shown.
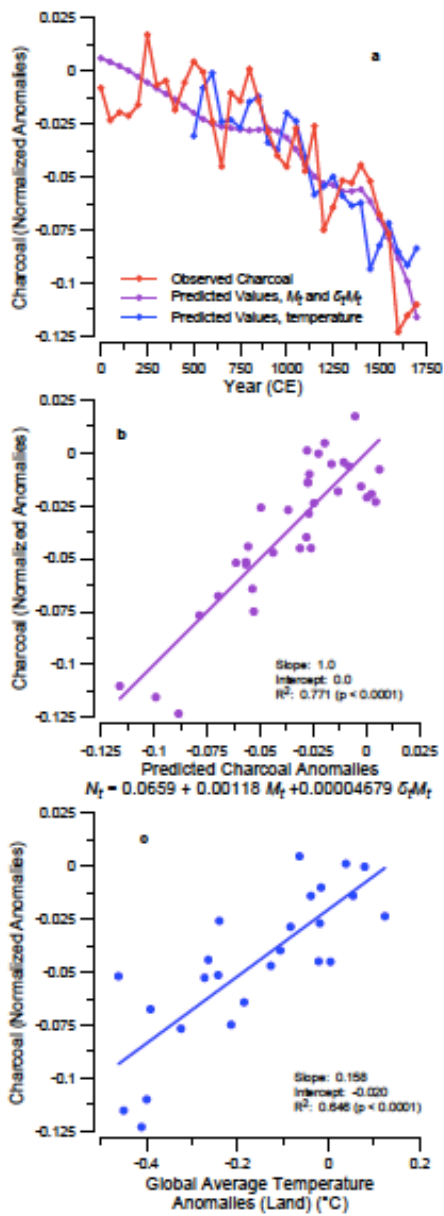
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Figure 5: Relationship between normalized charcoal anomalies and global land temperature. The data points refer to 50-year binned data. The top panel (a) shows observed charcoal normans; estimated values based on the linear regression of charcoal normans against the $\delta^{13}C$ of CH_4 and the product of this $\delta^{13}C$ value with the concentration of CH_4, as plotted in (b); and estimated values based on the linear regression of charcoal normans against temperature, as plotted in (c). Note that the slope and intercept of the relationship shown in panel (b) are necessarily 1.0 and 0.0, respectively – the key point is the goodness of fit shown between the two data sources after the charcoal data have been calibrated against the CH_4 and CH_4 isotopic records.
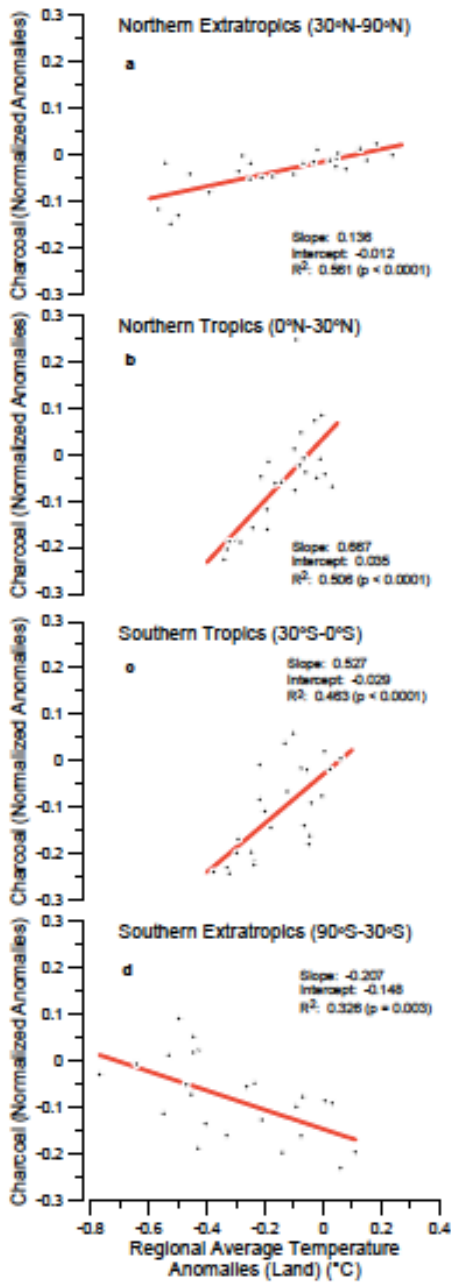
Figure 6: Relationship between normalized charcoal anomalies and land temperature for the (a) northern extratropics, (b) northern tropics, (c) southern tropics and (d) southern extratropics. The data points refer to 50-year binned data.