How Warm Was the Medieval Warm Period?

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Report

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A frequent conclusion based on study of individual records from the so-called Medieval Warm Period (~1000-1300 A.D.) is that the present warmth of the 20th century is not unusual and therefore cannot be taken as an indication of forced climate change from greenhouse gas emissions. This conclusion is not supported by published composites of Northern Hemisphere climate change, but the conclusions of such syntheses are often either ignored or challenged. In this paper, we revisit the controversy by incorporating additional time series not used in earlier hemispheric compilations. Another difference is that the present reconstruction uses records that are only 900–1000 years long, thereby, avoiding the potential problem of uncertainties introduced by using different numbers of records at different times. Despite clear evidence for Medieval warmth greater than present in some individual records, the new hemispheric composite supports the principal conclusion of earlier hemispheric reconstructions and, furthermore, indicates that maximum Medieval warmth was restricted to two-three 20–30 year intervals, with composite values during these times being only comparable to the mid-20th century warm time interval. Failure to substantiate hemispheric warmth greater than the present consistently occurs in composites because there are significant offsets in timing of warmth in different regions; ignoring these offsets can lead to serious errors concerning inferences about the magnitude of Medieval warmth and its relevance to interpretation of late 20th century warming.

INTRODUCTION

For many years it has been widely known that a ”Medieval” warm period occurred during an interval generally cited as being approximately 1000–1300 A.D. (e.g. 1–7). For example, grapes suitable for wine-making were reportedly grown in England (2), and the tree line in Scandinavia was 100–200 m higher than present (8). But were all of these changes synchronous, with hemispheric amplitudes comparable to or warmer than present? Very early in the discussion of this period a number of authors (1, 3, 4) pointed out that there were some significant phase offsets between the timing of warmth in different regions. Two recent (9, 10) Northern Hemisphere temperature reconstructions support the idea of Medieval warming being at most comparable to the mid-20th century Northern Hemisphere temperature peak (that is, about 0.3°C cooler than the decadal average of the 1990s).

Despite these compilations there are still widespread differences of opinion as to the relative warmth of the so-called Medieval Warm Period (MWP) *vis-à-vis* the present century (11). Some authors, especially greenhouse gas skeptics (e.g. 12), continue to extrapolate evidence from individual sites and small regions to infer that the present 20th century warmth is not unusual and is therefore evidence against a major effect of greenhouse gas changes on global climate. Because of the continued debate on this topic, it is revisited in this paper, with some different choices in data, which are also analyzed in a different manner than previous studies.

METHODS

There are two principal differences between the present reconstruction and those of Jones et al. (9) and Mann et al. (10): i) whereas the earlier reconstructions used a different number of records for different time intervals (with coverage for earlier time intervals sparser), the present reconstruction has almost the same number of records used for all time periods—there are a few instances of data cutoff problems at the ends of records but the number of records is still more time-invariant than previous studies; ii) a number of records (ice core, pollen, marine, historical climate records) were chosen that were not included in either of the previous reconstructions; the justification for these inclusions is that these records have often been cited as evidence for Medieval warmth and it is important to test robustness of conclusions with respect to relative levels of Medieval warmth.

Fifteen records were included in the summary (Fig. 1), with an attempt to obtain a balanced spread of sites from among the relatively small number of records that extend back approximately 1000 years. Four records are from the western two-thirds of North America—the White Mountain tree ring record from the lee of the Sierra Nevadas (13); tree ring records from central Colorado (14); and Jasper National Park, Alberta, Canada (15), and a pollen record from central Michigan (16). An oxygen isotope record from the western Sargasso Sea (17) was included, as were 6 sites from the northern North Atlantic/western European sector: the central Greenland GISP2 $\delta^{18}O$ ice core record (18); a historical sea ice/temperature record from Iceland (19); the central England temperature record (20) extended to 1000 A.D. by Lamb (1); tree ring records from northern Sweden (21); the Alps of southeastern France (22); and the Black Forest of Germany (23). The final set of 4 sites are from Asia: the Ural Mountains of western Siberia (24); a tree ring record from the Qilian Shan Mountains of western China (25); a $\delta^{18}O$ ice core record from the Dunde Ice Cap on the Tibetan Plateau (26); and a “phenological” temperature record from eastern China (3). This latter record is from the extensive historical Chi-

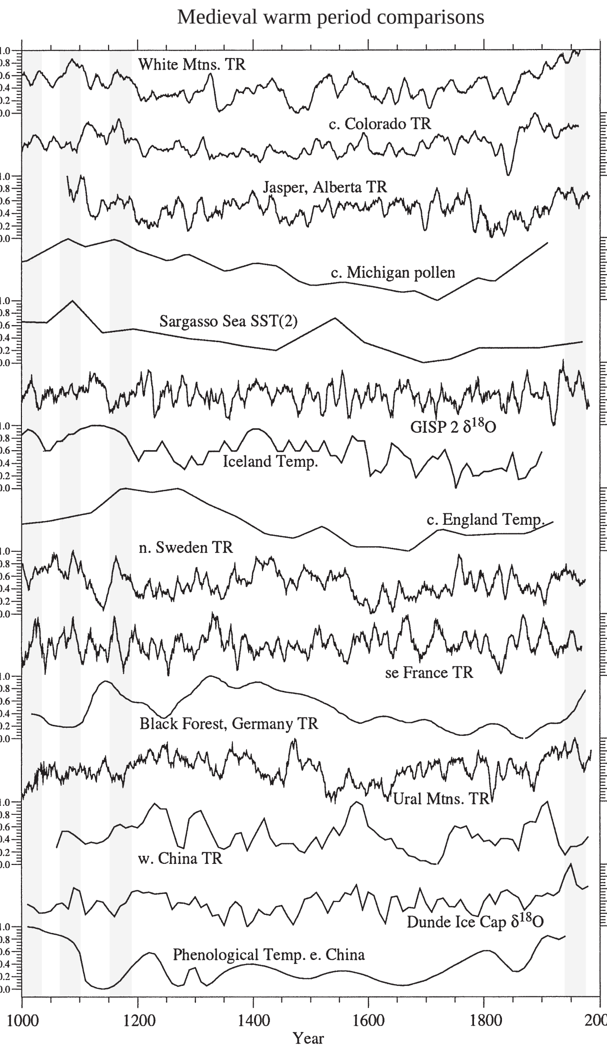


Figure 1. Scaled record of climate change based on 15 sites discussed in text. Vertical grey bars indicate intervals of maximum hemispheric warmth from Figure 2; see text for record citations.

nese data set (27) and is based on observations of changes in distribution of temperature sensitive biota and other climate indices.

Of the 15 records chosen for this investigation only 6 were included in the Jones et al. (9) and Mann et al. (10) data bases. The records chosen are also less homogeneous than the records employed by Jones et al. (9) and Mann et al. (10) in terms of both type of index and their correlation with temperature. For example, the Iceland record (18) is primarily a winter index, the central Michigan pollen record (15) is an estimate of growing season temperature, the Swedish (21), Urals (24), and west China tree ring records correlate best with summer temperature, the Black Forest (23) record is based on $\delta^{13}C$ measurements, and the central England record is an estimate of mean annual temperature (1). For reference, the Jones et al. (9) compilation is an estimate of summer temperature and the Mann et al. (10) reconstruction is an estimate of mean annual temperature.

Temporal resolution also differs from the earlier studies. While the Jones et al. and Mann et al. records have annual resolution, only 7 of the records from this study have such resolution. Five of the records have decadal-scale resolution and 3 have an average sampling resolution of about 50 years (16, 17, 20). In a sense, these inhomogeneities can be considered in a positive light as a sensitivity test to the robustness of the conclusion of Medieval warmth, with the repeat analysis justified based on the sheer frequency with which such records are used to make broad-scale generalizations about the relative magnitude of warmth in the Middle Ages.

With respect to analysis of the records they (Fig. 2) were scaled from 0 to 1, with annual resolution records first smoothed with an 11-point Stineman filter to bring out the lower frequency trends. The coarser resolution records (Michigan, Sargasso Sea, and central England) were interpolated to 1-year intervals. Due to chronology uncertainties (± 50 years) in the Sargasso Sea record (17), the peak warming was deliberately reset by 20–30 years to line up with maximum warming in the composite (see below); this was done to obtain an optimal configuration for maximum hemispheric warming so that the final conclusions would not be sensitive to chronology uncertainties. Because of the more uncertain temporal resolution of the Michigan pollen (16) and Sargasso Sea (17) records, we constructed 2 composites, the baseline without these 2 indices, and a second "full" composite with these 2 indices included.

RESULTS

A comparison of the individual climate records in Figure 1 and the hemispheric composites (Fig. 2) reveals some interesting patterns. The most prominent times of Medieval warmth in the composites are restricted to 3 relatively narrow intervals (1010–1040, 1070–1105, and 1155–1190). Highest MWP warmth is in the middle interval of the composite section (Fig. 2) and is found in 8 of the 15 records (Fig. 1), a percentage comparable to the 7 of 13 intervals that record the mid-20th century warm period. Subsequent to the third MWP decadal warming, temperatures decrease to a 17th century minimum. This time period (approximately1580–1850) has long been known as the coldest part of the "Little Ice Age" (LIA), with the beginning of this interval coinciding approximately with a pulse of volcanism in the late 16th century (28).

Despite the greater inhomogeneity of the data set in the present composite, the basic features of the previous composites are preserved in the present analysis (Fig. 2). Although the MWP temperature maxima in the different composites (Fig. 2) differ in relative magnitude they agree closely in timing. Correlations on the decadal band, using our baseline reconstruction with the 21-point. smoothed Jones et al. (9) and Mann et al. (10) records–justified to emphasize lower frequency variability–and accounting for the autocorrelation of the time series, yield values of 0.75 and 0.72, respectively, ($p < 0.01$, with correlations being 0.74 ($p < 0.01$) and 0.68 ($p < 0.05$) for our alternate (full) reconstruction). For reference, the correlation between the 21-point smoothed Jones et al. and Mann et al. records is 0.74 ($p < 0.01$). It is, therefore, clear that even a small, inhomogeneous data set can sometimes recover the basic features of hemispheric climate change, such as the Little Ice Age and mid-20th century warm period. This result supports the basic value of length-scale arguments concerning the relatively low number of independent samples needed to obtain reasonably reliable large-scale estimates of temperature (29, 30).

The non-synchroneity of temperature changes referred to in the introduction is evident when comparing the shaded intervals of maximum warmth in the composite (Fig. 2) with the patterns in individual records (Fig. 1). For example, none of the records between Germany and western China—about 100° of longitude –contribute significantly to the peak MWP warming from about 1070–1105. The oft-cited central England temperature record (1) contributes to the third MWP decadal warming (1155–1190) but most of the warming (1150–1290) postdates the final MWP peak in the composite. This response is shared by the Siberian and China records and is almost the inverse to the areas that were cool when a number of sites were warm between 1070–1105.

The spatial pattern for the center parts of two MWP warm intervals, the intervening cooler period (Fig. 2), and the mid-20th century warm period are compared in Figure 3. One difference between the Medieval and mid-20th century warmings involves the general restriction of peak MWP warming to the North American/Atlantic/western Europe sector, whereas the mid-20th century warming appears to be more of a land-sea difference. There are broad similarities between the proxy mid-20th century warming and the instrumental record (31), but due to lack of proxy data in the highest latitudes we cannot substantiate the maximum mid-20th century warming along what appears to be the snow/sea ice edge in the Arctic/North Atlantic sector. More proxy data would be required to test the robustness of the conclusion regarding spatial differences in warming pattern between the MWP and mid-20th century.

Although it might be tempting to attribute the MWP decadal temperature increases to changes in the North Atlantic thermohaline circulation, this temptation should be avoided. Peak Medieval warmth in central Greenland and Iceland, regions associated with a strong North Atlantic Current and more active thermohaline circulation, occurs during a *cool* interlude during

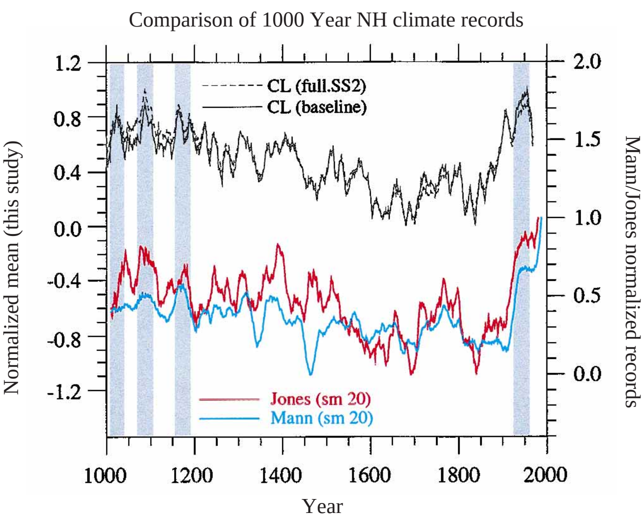


Figure 2. Comparison of hemispheric composites from this study with that of Jones et al. (9) and Mann et al. (10). Shaded intervals refer to times of peak warmth (see text). The dotted line indicates hemispheric composite values if two lower resolution records [Michigan pollen record (16) and Sargasso Sea $\delta^{18}O$ record(17)] are added to the baseline composite (see text). All records have been scaled between 0 and 1.

the MWP (Fig. 1). This and other temperature offsets may reflect displacements of the meridional flow pattern of the upper air westerlies (3–5). Again, more geographic coverage would be required to test this hypothesis, although a relatively dense network of tree rings for North America (32) supports this type of response for the cold 17th century climate fluctuation (Fig. 2).

The new composite time series were converted to mean annual temperature in the following manner. The two composites were scaled to agree with the Jones et al. (31) instrumental record for the Northern Hemisphere over the intervals 1856-1880 and 1920-1965 (too few of the proxies record information after this date). The reason for restricting the comparison to these two intervals involves the considerable deviation of the proxy time series from the instrumental record over the interval 1880–1920 (Fig. 4). The deviation occurs in 5 of our records (White Mountains, Colorado, Urals, and west and east China records), has been observed before (10, 33) and been attributed to (10) anomalous tree-ring growth due to the late 19th century rise in CO_2. Mann et al. (10) addressed this problem by removing the postulated CO_2 growth effect before estimating past temperatures. However, because this response also occurs in the Chinese phenological data set, another source of variance for high tree-ring growth rates cannot be excluded. The correlations between the present Crowley-Lowery (CL) composite and the Jones et al. instrumental record were therefore determined in two ways—one using the entire record 1856–1965 and the other using only the tie points 1856–1880 and 1920–1965, excluding the hypothesized interval of CO_2-induced tree ring growth. Correlations using the full time series are 0.55 (CL baseline) and 0.49 (CL full). Correlations using the 2 end member intervals are 0.87 (CL baseline) and 0.88 (CL full). All correlations are significant at the 1% level. Although all detrended correlations are significant at the 5% level, none explains more than 17% of the variance and are therefore of limited use from a paleoclimatic perspective.

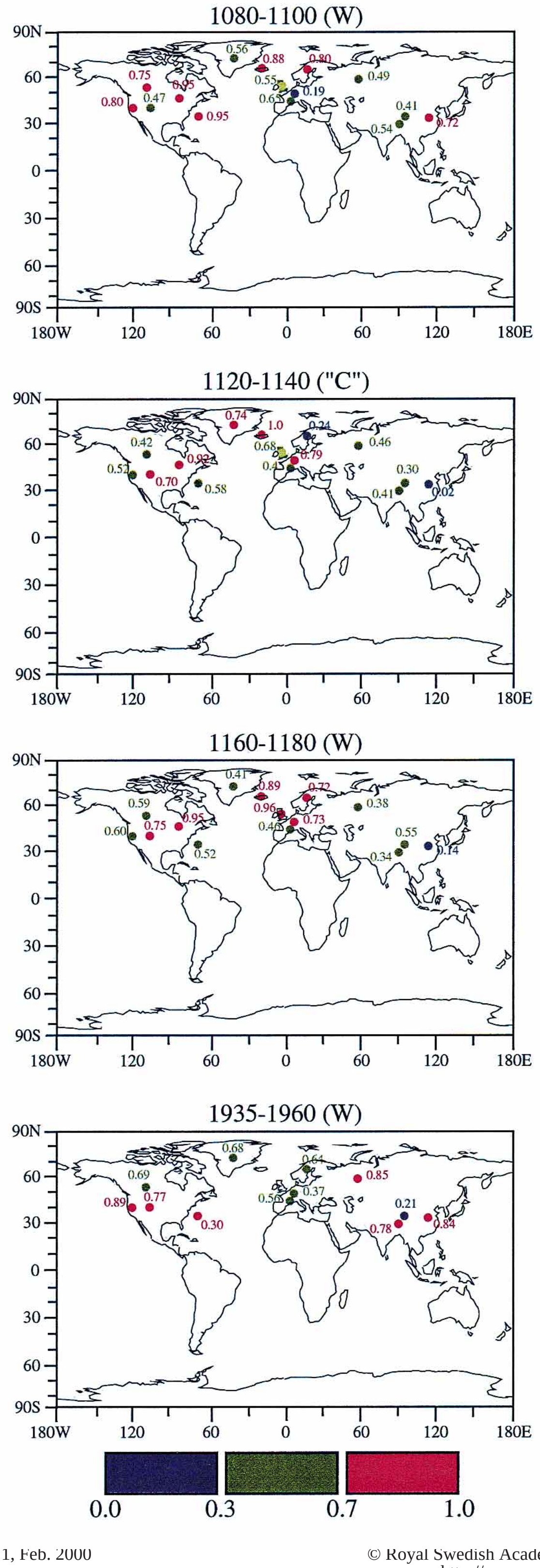


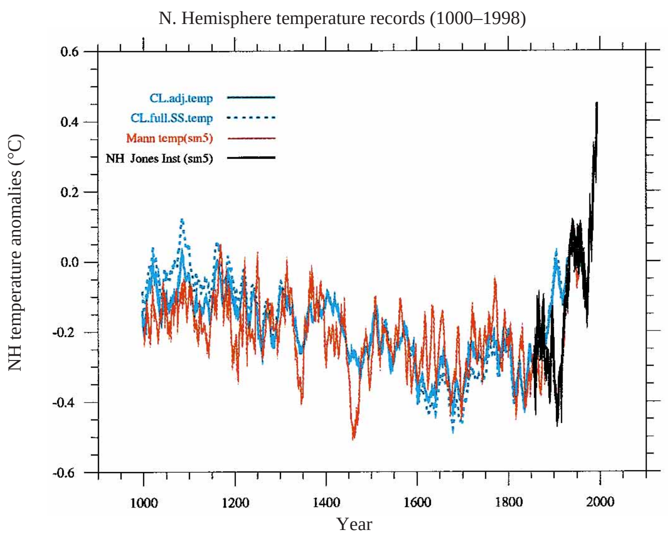
Figure 3. Comparison of the spatial pattern for the central parts of two MWP warm peaks, the intervening cooler period (Fig. 2), and the mid-20th century warm period. Ages of the intervals are listed in panel captions; values for individual sites are from Figure 1 data. Results are presented in terciles of relative warmth for the entire ~1000-year interval for each site.

Scaling the CL composites to the Jones et al. instrumental record (31) yields minimum LIA temperatures ~0.45–0.50°C less than the mid-20th century – a result similar to the Mann et al. (34) estimate of ~0.40°C, but less than the ~0.7°C estimate determined from borehole temperature estimates (35). Peak Medieval warming in our composites is with ± 0.05°C of the mid-20th century warm period. Average MWP temperatures (1000-1200) are only about 0.20°C warmer than the LIA interval of maximum cooling from about 1580–1840. If these numbers are substantiated by further investigations they would provide an important constraint on mechanisms for low-frequency climate variability. At this stage they can only be considered as estimates awaiting further clarifications of the reasons for the late 19th century divergence of the proxy records from the instrumental record and the disagreements between borehole and surface proxy records.

DISCUSSION AND CONCLUSIONS

To conclude, a new compilation of evidence for Medieval warmth indicates 3 relatively short-lived warming intervals (1010–1040, 1070–1105, and 1155–1190) that are comparable

Figure 4. Comparison of mean annual temperature records (Fig. 2) from this study with 5-pt. Smoothed Mann et al. (10) reconstruction and the Jones et al. (31) Northern Hemispheric instrumental temperature record. CL.adj.temp refers to the baseline composite adjusted to the Jones et al. record (see text); CL.full.SS.temp refers to all time series in the CL composite, with the Sargasso Sea (SS) record adjusted slightly in chronology to agree better with maximum warming in the hemispheric composite (again see text for details).



to the mid-20th century warm period. Scaling of the hemispheric composite to the Northern Hemisphere temperature records suggests that Little Ice Age temperatures were about 0.45–0.50°C colder than the mid-20th century warm period and that mean temperatures between 1000–1200 were only about 0.20°C warmer than the Little Ice Age. These results provide useful constraints on mechanisms of climate change on decadal-centennial time scales. For example, forced variations from CO_2, volcanism, and solar forcing have been implicated as contributing to the mid-20th century temperature increase after the Little Ice Age (28, 36–38), and changes in the latter may also have influenced the MWP (e.g. 39).

Because of uncertainties in the proxy-instrumental temperature calibration, it is still difficult to unequivocably assert that the late 20th century warming is significantly greater than the peak warmth of the Medieval Warm Period. But there is even less justification to assert the opposite–it is not possible to make a robust statement that the Medieval Warm Period was warmer than the last two decades. Similar conclusions can be derived from the sparser Southern Hemisphere data set of climate change over the last millennium (7, 9).

In an earlier study, Bradley and Jones (40) questioned the utility of the term "Little Ice Age" in light of their findings concerning significant decadal-centennial scale variability and regional climate trends sometimes of opposite signs. Given the findings of this study a similar concern could be raised about the utility of the term "Medieval Warm Period". Because mean temperatures during this interval were warmer than the subsequent Little Ice Age, we believe that the term Medieval Warm Period still has value, as long as it is restricted to the northern hemisphere (there is insufficient documentation as to its existence in the Southern Hemisphere) and as long as the user is careful to interpret regional trends within the context of hemispheric-scale variations.

The results from this study re-emphasize the hazards of using single or small-area records to make inferences about hemispheric warmth, particularly when the "evidence" is used to conclude that late 20th century warmth in not unusual in the context of the historical record of climate change. The results also indicate that the primary error associated with earlier conclusions is not statistical or climatological but rather stratigraphic – that is, the assumption that a climato- [or litho- (rock)] stratigraphic unit is a time stratigraphic unit. For more than 30 years geologists have recognized that this assumption is not valid in classical stratigraphic applications, but the error still frequently occurs when applied to interpretations of climate change. The error occurs despite the fact that the time resolution of one to a few years for records spanning the last millennium is vastly superior to any other time in the geologic record. Other examples of discordant decadal-centennial-scale trends involves peak warmth over the last 1400 years about 950 A.D. in the Greenland GISP2 $\delta^{18}O$ and borehole records (18, 41) almost exactly coincident with ice advances in Fennoscandia, the Alps, the Colorado Rockies (42, 43). The " Early Medieval Glacial Advances" (44) also occur at the same time as warmth in China (3). The widespread occurrences of such discordances underscores the need for extreme caution in extrapolating local climatic trends to larger-scale inferences.

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Comment

Temperature dependence of old soil organic matter

Comments on the paper by Liski, J., Ilvesniemi, H., Mäkelä, A. and Westman, C.J. Ambio Vol. XXVIII, No. 2, 1999.

Because of the large quantities of carbon contained in soil organic matter, its response to a changing climate is of paramount interest. In a recent paper, Liski et al. (1) argue that old soil organic matter is less sensitive to temperature change than young litter. Their result is derived from a model of soil organic matter turnover coupled with a model of net primary production (NPP) response to temperature. Virtually no experimental data exist to directly support or refute their result. Bunnell et al. (2) found in one study that the sensitivity to temperature of microbial respiration increased from the litter layer to the fermentation layer, and to the humus layer in an Alaskan tundra, i.e. temperature sensitivity increases with age. Similarly, the microbial respiration from 2-year-old standing dead *Eriophorum* was more temperature sensitive than 1-year-old standing dead but the opposite was true for *Carex*. Bosatta and Ågren (3) argued that low quality – or equivalently old – organic material is more temperature sensitive, on the basis of thermodynamics of enzyme kinetics. Fundamental theoretical arguments together with scanty empirical evidence seem therefore to contradict the assertion by Liski et al., and it becomes important to understand the features of their model that lead to the diverging results.

The soil organic matter turnover model by Liski et al. is a conventional compartment model where matter cascades from one compartment to the next or is lost as respiration. Each compartment in the sequence has a lower turnover rate than the previous one. In the Liski et al. model residence times in the compartments, except that for the young litter, are constant, but respiration rates increase with temperature. Consequently as temperature increases, smaller and smaller fractions will be transferred between compartments. For example, with the temperature sensitivity of young material, increasing the temperature from –1°C to +4°C means that approximately only half as much material is transferred between compartments. In practice this means that this factor alone halves the steady-state content in a compartment. In addition, the content of a compartment decreases with temperature because of the increased respiration while the matter resides in the compartment. The assumption of fixed residence times is an additional, implicit, assumption of a temperature dependence. Liski et al. counter this by letting the turnover rates of the old soil organic matter become less temperature sensitive. In most models, e.g. the Century model (4), the rate of transfer between the compartments increases with temperature in the same way as respiration and increasing temperatures mean shorter residence times. Another way of stating this is that a transfer from one compartment to the next, with a lower turnover rate, occurs when a certain fraction of the material has been lost; this is what Liski et al. do with young litter. Under such conditions the steady state soil organic matter stores, *Css*, are proportional to $NPP(T)/k(T)$. When then we apply the same NPP temperature response as Liski et al., $NPP(T)$, and their temperature response for young litter, $k(T)$, *Css* is virtually independent of temperature in agreement with their observations.

The critical, and distinguishing, feature in the Liski et al. model is, therefore, the assumption of a set of fixed residence times. However, soil organic matter is a continuum of qualities (5) and the use of discrete, compartment-type models can be a practical tool for approximating this continuum. The compartment-type models are normally not derived from the continuous distribution but transition rules are assumed on an *ad hoc* basis. The simplest models that can be derived from a continuous distribution (5) are such that transitions occur when a certain fraction of the material in a compartment has been lost. This seems also most natural. Transfers between compartments should take place when the quality of the material in the compartment has changed to a certain degree, and this is strictly coupled to the fraction of material lost (5) and not to the time the material has spent in a certain compartment.

Liski et al. also look at ^{14}C ages of their soil organic matter and find that they agree better with the model where old soil organic matter is less temperature sensitive. Dating of organic material that consists of a mixture of ages is, however, difficult because even small changes in the mixture of ages can have drastic effects on the average age (6).

Although my arguments have been to contest the validity of the conclusions by Liski et al., namely that old soil organic matter is less temperature sensitive than young material, we have so far very little information from which to draw definite conclusions. Additional complicating factors that need to be considered when using geographical temperature gradients is that the relative composition of the litter with respect to decomposability may not be constant over the gradient, and that the temperature signal with depth (and hence age) also varies over the gradient. Finally, a climate change might cause unforeseen changes in the decomposer community. It has, for example, been shown that respiration from forest soil organic matter and agricultural soil organic matter have completely different temperature responses (7), which could be a result of a shift in the relative importance of fungi and bacteria as decomposers. Changes of that kind are not included in any models of climate change effects on decomposition.

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Temperature Dependence of Old Soil Organic Matter

Reply to the comments by Göran Ågren on a paper by Liski et al.
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In our recent paper (1), we concluded that the decomposition of old soil organic matter is less temperature sensitive than the decomposition of young litter. Consequently, we judged that studies, which assume that the decomposition of all soil organic matter is as sensitive to temperature as the decomposition of young litter, overestimate the release of carbon from soil in response to climatic warming. We based our conclusion on measurements and model calculations of the amount and age of soil carbon on temperature gradients.

Ågren contests our conclusion. He argues that it was a result of a feature of our soil carbon model, namely the constant residence times of carbon in the compartments.

In many other soil carbon models, the residence times vary with decomposition rates, as a certain fraction of carbon decomposed in one compartment is transferred to the next. In such models, the steady state amount of soil carbon at temperature T, $Css(T)$, is derivable from the corresponding carbon input to soil, $I(T)$, and the specific decomposition rates of the n compartments, $k_i(T)$, as follows:

$$Css(T) = I(T) \sum_{i=1}^{n} (b'_i/k_i(T)) \qquad \text{Eq. 1}$$

where the coefficients b_i are constants. If the temperature response of decomposition is similar in each compartment, the equation reduces to

$$Css(T) = I(T)/k_1(T) \sum_{i=1}^{n} b'_i \qquad \text{Eq. 2}$$

where $k_1(T)$ is the specific decomposition rate of the youngest compartment and any level differences between the compartments have been embedded in the constants b'_i.

Using this model (Eq 2), Ågren calculated relative steady–state amounts of soil carbon for the temperature range of our soil carbon measurements. He applied the same temperature response for carbon input to soil as we did, and the same temperature response for all decomposition as we applied for the decomposition of young litter. He obtained 0.98, 1.06, 1.08, 1.06, and 1.00 for the amounts at annual mean temperatures –1.0, 0.4, 1.6, 2.8 and 4.0, respectively. He stated that the values were "virtually independent of temperature in agreement with their (our) observations". This would mean that the decomposition of old soil carbon would not be less sensitive to temperature than the decomposition of young litter.

Did the above relative steady state amounts of soil carbon Ågren calculated really agree with our observations? We calculated the probability of observing the trend in soil carbon we observed in our original study (2) if the amount of soil carbon depended on temperature like in Ågren's calculation. To convert his relative amounts of soil carbon to absolute ones, we took the absolute amount of soil carbon at 4°C from the linear regression fitted to our original data and multiplied this by his relative amounts to estimate the absolute amounts at the lower temperatures. Note that the choice of the reference temperature, here 4°C, does not affect the resulting trend. We assumed that at each temperature the variance of soil carbon was equal to the variance observed in our original data and that the variation in soil carbon followed the normal distribution. We then randomly drew 4 soil carbon values for each temperature from such normal distribution and fitted a linear regression to the values. We repeated this 1000 times to mimic 1000 sets of field observations and calculated in how many sets the slope of the linear regression was equal to or greater than the slope of the regression fitted to our original data.

The probability of obtaining such a slope was 7% for the compounded organic and 0–1 m mineral soil layer of a high productivity forest type and 6% for this layer of a low productivity forest type; we studied two forest types in our original work. For the 0–1 m mineral soil layer alone, the probability was 0.5% for the high productivity forest type and 0.3% for the low productivity forest type. The probability of obtaining the slopes simultaneously in both forest types was 0.4% for the compounded organic and 0–1 m mineral soil layer and 0.002% for the 0–1 m mineral soil layer.

These probabilities indicate that if the amount of soil carbon depended on temperature like in Ågren's calculation, it would have been very improbable that we had observed the trends in soil carbon we observed in our original study on the temperature gradient. This suggests that the relative steady-state amounts of soil carbon Ågren calculated do not agree with our observations.

The agreement between our observations and the results of the model Ågren used for his comments (Eq. 1, 2) could be improved by decreasing the temperature sensitivity of decomposition. Since this sensitivity has been determined experimentally for young litter, but little is known about older soil organic matter, it would be reasonable to decrease the sensitivity of the decomposition of old soil carbon. On the other hand, making the decomposition of old soil carbon more sensitive to temperature than the decomposition of young litter, as Ågren suggests in his comments on the basis of "fundamental theoretical arguments" (3), would impair the agreement between the model results and our observations.

These analyses illustrate that our conclusion about the temperature tolerance of the decomposition of old soil carbon is not merely an artefact caused by technical assumptions in our soil carbon model, but the same conclusion is obtained using the more conventional model structure employed above. However, it is true that the differences between the scenarios of decreasing, increasing and stable temperature sensitivities become less pronounced with the model structure suggested by Ågren. The present conclusions are therefore strongly dependent on our empirical data on both carbon input to soil, the temperature sensitivity of the decomposition of young litter and the soil carbon contents on the temperature gradient. This kind of evidence is always more or less circumstantial, and further data would certainly prove enlightening.

Ågren argues that models, in which a certain fraction of carbon decomposed in one compartment is transferred to the next one, are more "natural". We agree, although they are not necessarily more accurate. Quantifying the fractions to be transferred is not easy. Individual compartments in soil carbon models or fluxes between the compartments cannot usually be measured or otherwise observed in real life. In most cases, the transfer parameters must be adjusted on the basis of the behaviour of the whole system. Sometimes several parameter combinations may give equally satisfactory calibration results but the system may behave differently in applications depending on which combination is applied. We chose our modelling approach with constant residence times to