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Heat transport in the Red Lake Bog, Glacial Lake Agassiz Peatlands

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Abstract:

We report the results of an investigation on the processes controlling heat transport in peat under a large bog in the Glacial Lake Agassiz Peatlands. For 2 years, starting in July 1998, we recorded temperature at 12 depth intervals from 0 to 400 cm within a vertical peat profile at the crest of the bog at sub-daily intervals. We also recorded air temperature 1 m above the peat surface. We calculate a peat thermal conductivity of 0·5 W m^{-1} $°C^{-1}$ and model vertical heat transport through the peat using the SUTRA model. The model was calibrated to the first year of data, and then evaluated against the second year of collected heat data. The model results suggest that advective pore-water flow is not necessary to transport heat within the peat profile and most of the heat is transferred by thermal conduction alone in these waterlogged soils. In the spring season, a zero-curtain effect controls the transport of heat through shallow depths of the peat. Changes in local climate and the resulting changes in thermal transport still may cause non-linear feedbacks in methane emissions related to the generation of methane deeper within the peat profile as regional temperatures increase. Copyright © 2006 John Wiley & Sons, Ltd.

KEY WORDS Glacial Lake Agassiz Peatlands; heat transfer; thermal conductivity; SUTRA

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INTRODUCTION

Peatlands, which are dominated by the accumulation of waterlogged peat soils more than 40 cm thick (National Wetland Working Group, 1997), constitute a major global carbon sink. Northern peatlands store $\sim 5 \times 10^{17}$ g of organic carbon (Gorham, 1991), equivalent to ~100 years of current fossil-fuel combustion (Moore *et al*., 1998). The interaction of peatlands with the Earth's atmosphere is dynamic and complex. Peatlands release methane, formed by anaerobic bacteria at depth, to the atmosphere. The transfer of methane from depth within the peat profile occurs either by episodic releases of large volumes of methane gas associated with the lowering of peatland water tables (Windsor *et al*., 1992; Moore and Roulet, 1993; Rosenberry *et al*., 2003) or by continuous diffusion through the peat soil. In some peatlands, the methane generated can form deep pockets of free phase, overpressured methane that causes the surface of the peatland to rise and fall due to the build-up and episodic release of the gas (Rosenberry *et al*., 2003; Glaser *et al*., 2004).

Numerous factors control the overall production of organic matter in a peatland, including the availability of nutrients, the hydrologic regime, the climatic regime, and temperature. Temperature is important because it controls the rate at which biological and chemical processes take place. Bridgham *et al*. (1999) found from empirical greenhouse studies that increased surface radiation produces significant changes in methane and carbon dioxide production rates in peat. Respiration rates and the development of anaerobic conditions depend strongly on temperature (Lewis, 1995). A decrease in soil temperature lowers the rate of decomposition and increases the rate of peat accumulation (Fox and Van Cleve, 1983). Chapman and Thurlow (1996) calculated for a bog in Scotland that an increase in surface temperature of 4·5 °C might double CO_2 emissions and increase methane emissions by 60% based on observations of two peat sites.

Rappoldt *et al*. (2003) proposed that near-surface waters (less than 35 cm depth) in the peat column freely convect. In their study, water moved primarily because of daily density differences driven by diurnal temperature changes. However, free convection only applies to the very open, upper fibric parts of peat columns where the distinction between pure surface water and water in porous media is blurred. Below this fibric zone, humified peat in peatlands can be many metres thick.

Few studies report the results of investigations on heat transport through entire peat profiles in non-permafrost regions (e.g. Wilson, 1939). Thermal investigations of peatlands generally do not cover the deeper (>3 m depth) parts of a peat profile because seasonal temperature

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oscillations are most pronounced within 1–2 m of the peat surface. Moore (1987) measured temperature from the land surface to a 200 cm depth in peat at seven depth intervals from six peatland sites in northern Quebec and found that both surface temperatures and depth of freezing were controlled by the depth of snow cover. In a Minnesota bog, Brown (1975) found the temperature of surface peat closely followed that of air temperature, but at depths of 2 m there was a multi-month lag between surface temperature and temperature at depth. The timing of the seasonal freezing cycle is important because the highest diffusive methane release from peatlands occurs from a few months to nearly 6 months after snowmelt, when soil temperatures reach their annual maximum (Boeckx and van Cleemput, 1997; Duval and Goodwin, 2000).

Surface waters on domed bogs move downward into the peat profile because of local water table mounds that develop within bogs; and in many cases, groundwater in larger scale flow systems can move vertically into or out of the base of peat profiles (Siegel *et al.*, 1995). Vertically moving peat pore water is a potentially important process controlling the distribution of solutes within the peat profile (e.g. Romanowicz *et al.*, 1993; McKenzie *et al.*, 2001). 'Flow reversals', a situation where the vertical direction of pore-water movement changes episodically or seasonally, have been well documented in many peatland types and settings (e.g. Siegel *et al*, 1995; Devito *et al.*, 1997; McKenzie *et al.*, 2001) and also in non-peatland settings (Phillips and Shedlock, 1993; Rosenberry and Winter, 1997). Horizontal flow in domed bogs is primarily radial in nature, moving away from the crest of the bog (Figure 1b).

In a solid matrix and pore-water system, heat is transferred by conduction through the water and the solid matrix, and by advection (convection) of pore water through the interconnected pore space. The one-dimensional transfer of heat by conduction alone q_h is described by Fourier's law (Carslaw and Jaeger, 1959):

$$q_h = -k_m \frac{dT}{dz} \tag{1}$$

where k_m is the thermal conductivity of the medium, T is temperature, and z is the distance in the vertical direction. Assuming a temperature range where neither ice nor gas will form, the governing equation for heat conduction and advection through isotropic, homogeneous, saturated soils is (Bredehoeft and Papadopulos, 1965)

$$\frac{\partial^2 T}{\partial x^2} + \frac{\partial^2 T}{\partial y^2} + \frac{\partial^2 T}{\partial z^2} - \frac{c_f \rho_f}{k}\left[\frac{\partial(v_x T)}{x} + \frac{\partial(v_x T)}{y} + \frac{\partial(v_z T)}{z}\right] = \frac{c_{fs}\rho_{fs}}{k}\frac{\partial T}{\partial t} \tag{2}$$

where T is temperature, x, y, z are Cartesian coordinates, c_f and ρ_f are respectively the specific heat and density of the fluid, v_x, v_y, v_z are components of fluid velocity, c_{fs} and ρ_{fs} are respectively the specific heat and density of the fluid–solid system, and k is the thermal conductivity of the solid–fluid complex.

Considering the importance of heat with respect to surface and deeper biochemical processes on the storage and release of organic carbon in peatlands, we report here the results of a study to evaluate the detailed thermal processes of a peat bog in the Glacial Lake Agassiz Peatlands (GLAP). Through the analysis of field data and the use of numerical simulations of heat transport, we addressed the following questions:

1. To what depth are seasonal land-surface temperature variations propagated into the peat profile?
2. What are the thermal processes controlling the thermal regime of the peatland?
3. What is the influence of advection and conduction on the vertical transfer of heat through the peat column?

STUDY AREA

The Red Lake Bog is situated within the GLAP (northern Minnesota and southern Manitoba), a 7000 km^2 expanse of sub-boreal patterned peatlands (Glaser *et al.*, 1981; Figure 1). The surface waters of bogs in the GLAP have total dissolved solids less than 15 mg l^{-1} and a pH less than 4·0 (Glaser *et al.*, 1981). The primary source of water to the bog is precipitation, but there is a small amount of groundwater discharge into the base of the peat profile. The Red Lake Bog has been shown to function as a local recharge cell superimposed over a regional flow system (Glaser *et al.*, 1997, 2004).

The vegetation assemblage of the raised bogs is dominated by *Picea mariana, Carex oligosperma*, and ericaceous shrubs with a continuous mat of *Sphagnum* (Heinselman, 1970; Glaser *et al.*, 1981). The average peat thickness is 2 to 3 m, reflecting ~5000 years of accumulation (Glaser *et al.*, 1981). The study site in the Red Lake Bog is situated along the bog crest in watershed II (Glaser *et al.*, 1981), where the peat depth is 4·3 m. The upper 3 m of peat consist of oligotrophic bog peat derived largely from *Sphagnum* moss and wood and the upper ~50 cm is fibric in texture, grading to dark brown humic peat below. There is no sapric peat in the peat column. Beneath the bog peat is 60 cm of fen peat, which is also humic. Within the peat column are discontinuous layers of wood and occasional ash layers (Janssens *et al.*, 1992).

Average precipitation on the GLAP ranges from 64 cm in the eastern part to 56 cm in the west (Glaser *et al.*, 2004). The peatlands straddle a north–south-trending divide where average annual evapotranspiration is approximately equal to average annual precipitation. Consequently, the peatlands are highly sensitive to the regional climate and east–west shifts of the precipitation and evapotranspiration divide caused by droughts or wet periods. The peatlands are typically covered by at least 15 cm of snow for 70 to 100 days per year (Glaser *et al.*, 2004) and subject to extreme multiyear droughts (Glaser *et al.*, 2004).

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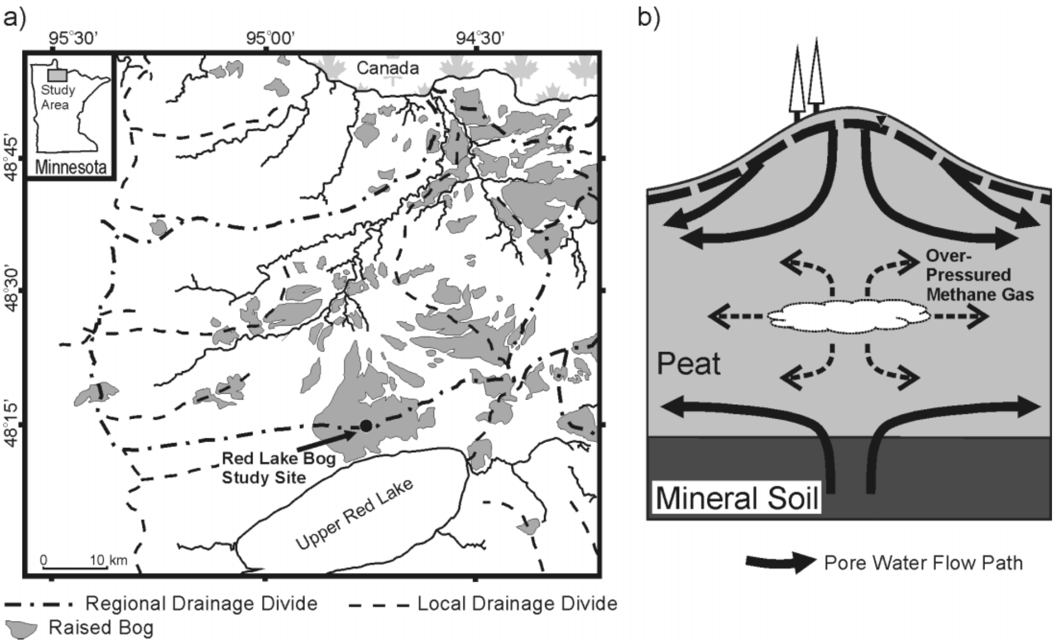

Figure 1. a) Map of the study area in the GLAP showing the location of the Red Lake Bog. b) Schematic diagram of the hypothesized pore water flow field in the Red Lake Bog. Groundwater flow is upward into the base of the peat profile and downward from the surface of the peat (Siegel *et al.*, 1995). Hydraulically over-pressured methane is episodically present at ~2 m depth (Rosenberry *et al.*, 2003). The bog is approximately 10 km wide and the peat is 4 m thick at the center

METHODS

Fieldwork

We intensively instrumented the Red Lake Bog site for long-term hydrometric analysis. Thermocouples were installed at depths of 0, 25, 50, 75, 100, 125, 150, 200, 250, 300, 350, and 400 cm at the crest of the Red Lake Bog. Hydraulic pressure was measured at three depths, and the results are reported in Rosenberry *et al.* (2003). Temperature and atmospheric pressure were measured approximately 1 m above the land surface. The data are stored with a solar-powered data logger connected to a mobile phone for remote data access. Temperature data at all depths were averaged and recorded approximately every 5 min from 18 June 1998 to 25 November 1998, daily from 25 November 1998 to 17 October 1999, and every 5 min from 17 October 1999 to 21 June 2000.

The instrumentation was installed at the hydrologic divide on the crest of the Red Lake Bog, where hydraulic head measurements and the hydrogeologic setting indicate that pore water mostly moves in the vertical plane (Rosenberry *et al.*, 2003). Here, because of negligible horizontal flow, moving heat and water can be simulated as vertical, one-dimensional flow (Anderson and Woessner, 1992).

Thermal diffusivity and conductivity

We determined the thermal diffusivity of the peat from the temperature data. Fourier transforms calculated the yearly amplitude of the annual temperature fluctuations at each measured depth and the decrease in amplitude with depth was used to calculate the thermal diffusivity k (Carslaw and Jaeger, 1959; Smith, 1989):

$$k = \frac{\omega(\Delta y)^2}{2[\ln(A_z/A_{z+\Delta z})]^2} \tag{3}$$

where ω is the angular frequency of the fluctuation in radians/unit time ($\omega = 2\pi/f$, where $f = 1/\text{period} = 1\ \text{year}^{-1}$), A_z and $A_{z+\Delta z}$ are the amplitudes of the annual temperature changes at two depths and Δz is the distance between these depths. The thermal diffusivity k was calculated for each measured depth relative to the peat surface, and we used an average of values from all depths in our modelling experiments. The range in thermal diffusivities we measured was ~35 cm^2 day^{-1}, and so we feel using the average value was appropriate. Amplitudes that were less than 1 °C were not included to eliminate potential thermocouple error.

Modelling

Heat transport was simulated with the saturated–unsaturated transport model SUTRA (Voss, 1984). The Argus ONE open numerical modelling environment and the USGS graphical user interface (Voss *et al.*, 1997) were used as pre- and post-processors for the SUTRA model. The model is a hybridization of finite-element and finite-difference methods; finite-element methods are used to simulate fluxes of fluid mass, and finite-difference methods are used for all other non-flux terms, such as solute mass and energy (Voss, 1984). Groundwater flow, if present, is simulated with a numerical solution of the fluid mass balance equation for saturated flow and variable fluid density (Voss, 1984). Energy transport is simulated through numerical solutions of an energy balance equation that assumes the solid matrix and pore

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water are locally at equal temperature, and that fluid density and viscosity are variable (Voss, 1984).

The Red Lake Bog model simulates vertical one-dimensional heat transport, appropriate for the conditions at the bog crest. We used a constant-head boundary for the top boundary, reflecting the water table, and no-flow boundaries on the lateral sides of the model. The elevation of the groundwater table was held constant at the land surface because the elevation of the water table only fluctuated a small amount (<10 cm) over the period of record. The hydraulic portion of the model was set at no flow and the heat flux model was run transiently. To model heat transport, we applied specified temperature boundaries to the top and bottom of the model. The time step for the model was 1 h and the input temperature values for the heat boundaries were changed daily based on mean daily temperature values taken from the field measurements of air temperature at the peat surface and water temperature at the bottom of the peat column. We calibrated the model by varying the porosity and thermal properties of the soils. The model was calibrated to mean daily recorded temperatures at all depths for the first year of data, starting in July 1998. The model was evaluated by comparing model results for a subsequent year (starting in July 1999) with measured data.

Sensitivity analysis was done to assess the relative effect of individual parameters on the model results. The sensitivity analysis measured the relative effect of small changes in the values of specific heat capacity of the solid grains, the thermal conductivity of the solid grains, and the total porosity on the model results. The calculated root-mean-squared statistic compared measured versus modelled values for heat data for all depths, and from these data we calculated a normalized sensitivity coefficient (Zheng and Bennett, 2002).

RESULTS

Temperature measurements

Heat data in the peat from the raised bog shows both seasonal and daily cycles, both of which are predominately controlled by temperatures at the surface of the peat profile (Figure 2). The temperature at the base of the peat profile is nearly constant at $\sim$4·8 °C. The mean air temperature for the 5 years prior to the experiment and the 2 years during the experiment was 3·4 °C (from the NCDC-NOAA data for the Waskish, Minnesota, weather station). Commonly, the average temperature of deeper soils is assumed to be approximately the average air temperature plus 1 °C (Soil Survey Division Staff, 1993), which is similar to what we observe.

The temperature at the peat surface fluctuated from -7·9 to $+22$·2 °C over the 2-year data-collection period,

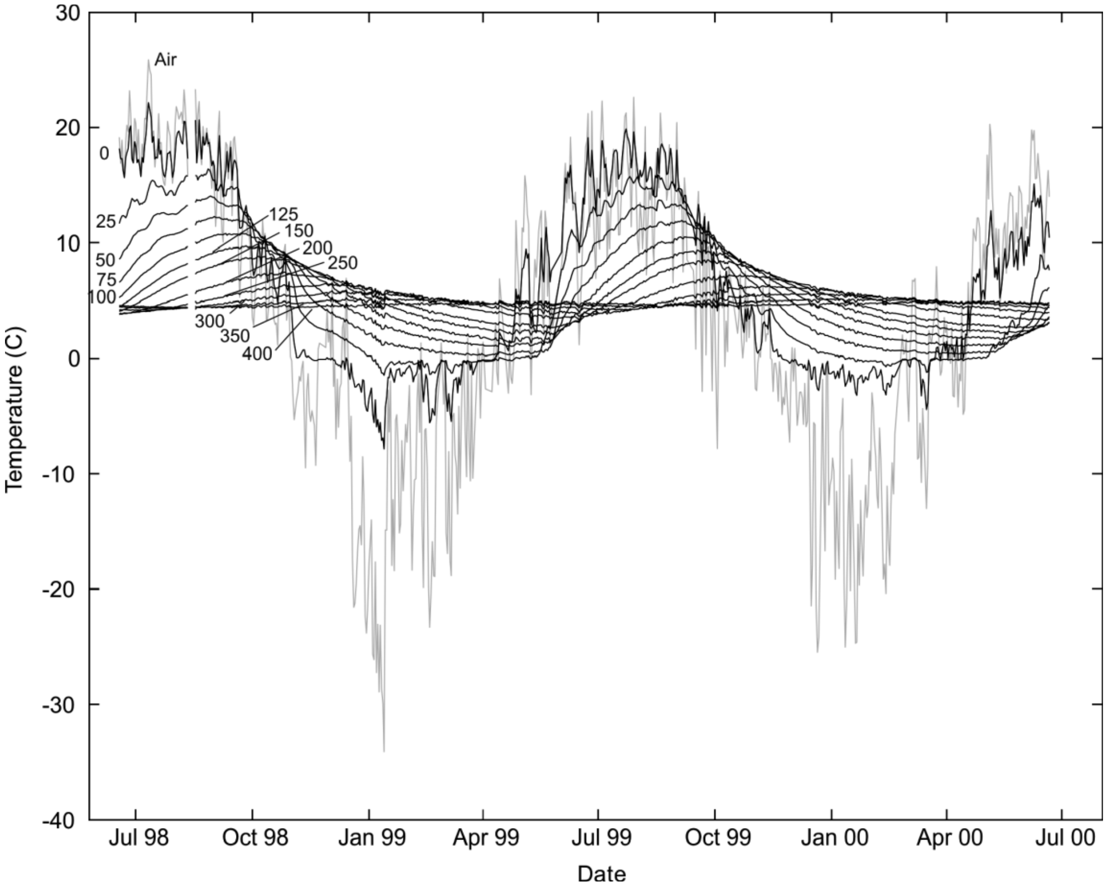

Figure 2. Daily average of heat data collected at all depths. Numbers refer to depth within peat profile in centimetres. The air temperature is measure $\sim$1 m above the land surface directly above the location where the heat data are collected. The data gap was the result of a field instrumentation failure

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and the pattern of fluctuations is a subdued replica of the local air temperature. The greatest deviation between the temperature of the peat and the local air temperature (measured ~1 m above the peat surface) occurred during the winter. For example, in January and February 1998 the average daily air temperature repeatedly dropped below $-15\,^{\circ}$C, but the temperature at the peat surface ranged between 0 and $-5\,^{\circ}$C (with the exception of 12 January 1999, when the average daily surface temperature was $-34{\cdot}1\,^{\circ}$C and the average daily temperature of the peat surface was $-7{\cdot}9\,^{\circ}$C).

The temperatures in the shallow peat fluctuated daily (Figure 3). Seasonally, a similar pattern of fluctuation of temperature was observed, except that the surface temperatures were transmitted to a substantially deeper depth because of increased amplitude of surface temperatures. Seasonally, the vertical thermal gradient reverses direction. During the summer, the surface of the peat profile is warmest, and the coolest temperatures are at the base of the peat profile. During the winter and early spring, the warmest temperatures are at the base of the peat profile and the coolest temperatures are at the surface of the peat profile (Figure 2).

From summer through fall the temperatures at near-surface depths ($\leq$100 cm) closely mimic the periodicity of temperature at the surface (0 cm) and air temperatures. During spring, there is a lag of 35 days between the rise in surface air temperature and the near-surface temperatures ($\leq$100 cm) within the bog, followed by a rapid rise in temperature at shallow depths within the bog.

Figure 4 shows our calculated Fourier transform amplitude spectrum diagrams for each depth that had calculated amplitudes greater than $1\,^{\circ}$C. The annual temperature amplitudes decrease with depth, from a $10{\cdot}4\,^{\circ}$C amplitude at the peat surface to $1{\cdot}1\,^{\circ}$C amplitude at the 250 cm depth. Below 250 cm, the annual temperature amplitude is less than $1\,^{\circ}$C and is difficult to distinguish from background fluctuations. The average thermal diffusivity of the peat is $\sim$110 cm^2 day^{-1}. Assuming an average humified peat density of 0·85 g cm^{-3} (e.g. Clymo, 1984) and a peat heat capacity of 4576 J kg^{-1} $^{\circ}$C^{-1} (Moore, 1987), the thermal conductivity of the peat–water system is $\sim$0·5 W m^{-1} $^{\circ}$C^{-1}.

Calibration results

The parameters used in the calibrated model are listed in Table I. The calibration was based on the average temperatures for each measured depth. The mean absolute residual between measured and modelled temperatures, the root-mean-squared residual error, and the linear correlation coefficient for the whole model are 0·34, 0·59 and 0·99 respectively (for 3450 data points). Figure 5 shows the results comparison between measured and modelled temperatures graphically and Figure 6 shows a calibration plot for temperature data at all depths. For individual depths, the greatest error occurred at the 25 cm depth (root-mean-squared residual error of 1·31) and decreased with depth.

The greatest deviation from the measured and modelled temperatures occurred during April and May. The

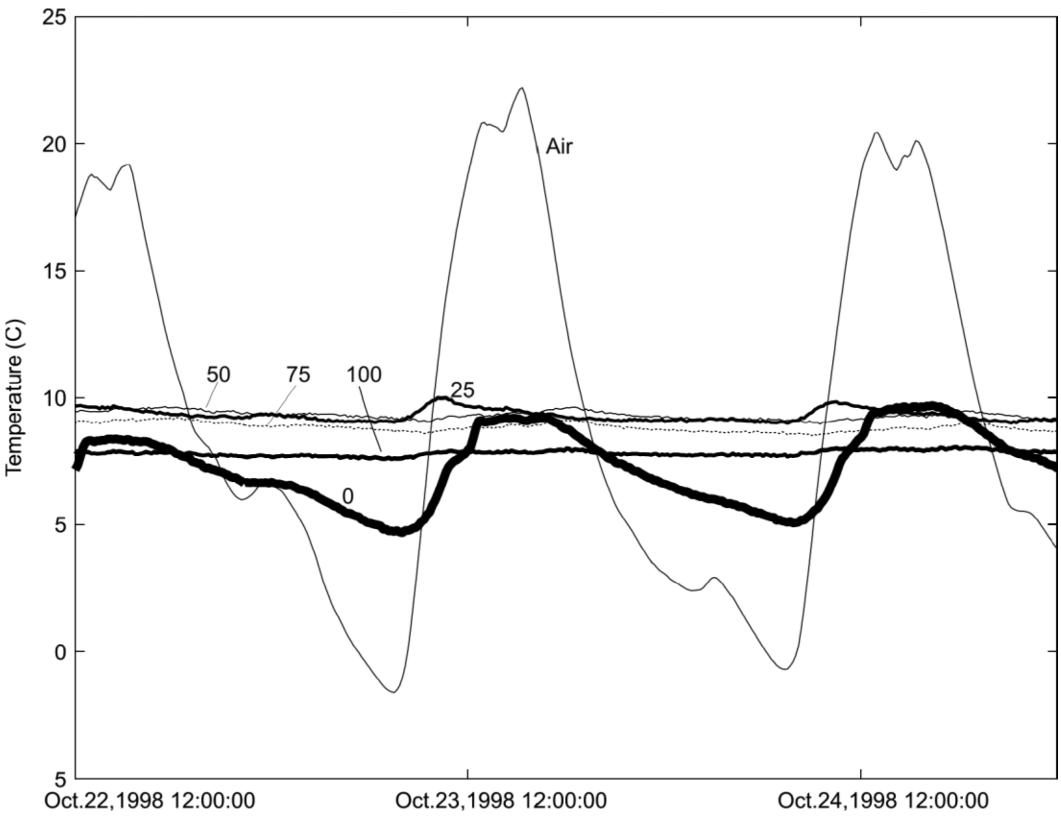

Figure 3. A 2-day focused view of the measured heat data. The numbers on the figure refer to depths at which the measurements were made, in centimetres

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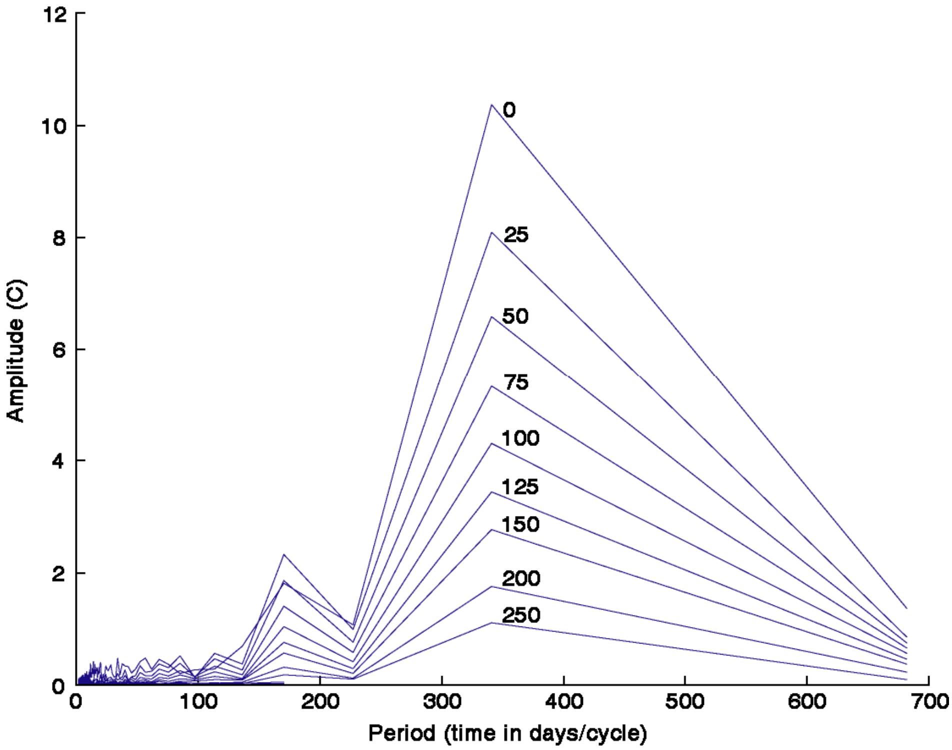

Figure 4. Results of Fourier transform analysis of daily average temperature values over 2-year period. The largest peak for each depth corresponds to 365 days, as would be expected. The numbers beside each line correspond to the depth of measurement in centimetres. The 0 cm depth represents the temperature at the peat surface

Table I. Parameter values used in calibrated model

Parameter	Value	
Fluid specific heat (J kg^{-1} °C^{-1})	4182	Fixed
Thermal conductivity of fluid (J s^{-1} m^{-1} °C^{-1})	0·6	Fixed
Specific heat capacity of peat (J kg^{-1} °C^{-1})	3500	Calibrated
Thermal conductivity of solid grains (J s^{-1} m^{-1} °C^{-1})	0·4	Calibrated
Porosity	0·3–0·5	Calibrated
Density of peat (kg m^{-3})	850	Fixed

deviation may be due to a zero-curtain effect, defined as a 'persistence of a nearly constant temperature, very close to the freezing point, during annual freezing (and occasionally during thawing) of the active layer' (van Everdingen, 2005). At the time of this research, SUTRA was unable to simulate freezing (McKenzie *et al.*, 2006). Excluding April and May, the mean absolute residual error and the root-mean-squared error of the SUTRA model improve to 0·29 and 0·38 respectively.

Sensitivity analysis

Sensitivity analysis ranked the relative impact of model parameters on the model output. The model was most sensitive to the thermal conductivity of the peat, followed by the porosity, the specific heat capacity of the peat, and the density of the peat. The total porosity is an important calibration factor because porosity controls the ratio of liquid water volume to solid peat volume.

Evaluation

The model was evaluated by running the calibrated model for the next year of data, from 1 July 1999 to 20 June 2000. The model evaluation (Figure 7) had a mean absolute error of 0·9 °C, a variance of residual errors of 1·3 °C, and a root-mean-squared error of 1·2 °C for the measured data at all depths versus the modelled data. The evaluation model deviates from the measured heat data during the spring melt, similar to what is observed in 1999.

DISCUSSION AND CONCLUSIONS

We feel that the calibrated model was a good analogue for one-dimensional vertical heat transport in the Red Lake Bog. The calibrated model and the spectral analysis methods determined values for saturated peat thermal conductivity of 0·4 W m^{-1} °C^{-1} and 0·5 W m^{-1} °C^{-1} respectively, similar to values calculated by other researchers (e.g. Moore, 1987; Williams and Smith, 1989).

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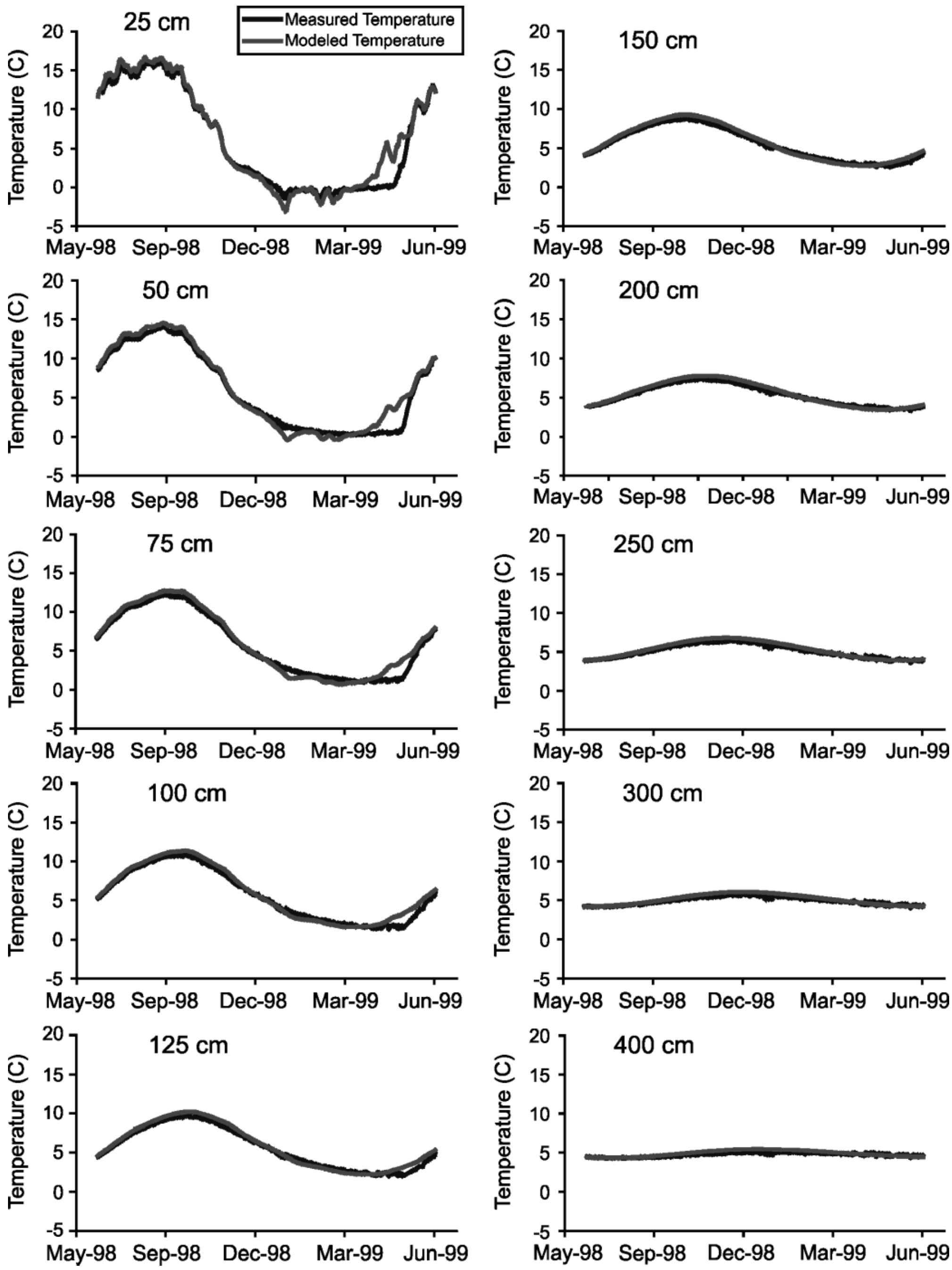

Figure 5. Comparison of the measured and modelled temperature results versus time for the calibrated SUTRA model (1998–1999) at 12 observation depths

In the Red Lake Bog model all of the heat transport is assumed to be by conduction. At the study site, there are small amounts of groundwater flow upward into the base of the peat profile and downward from the surface of the peat (Figure 1b). Pockets of methane at depth within the peat profile also create transient, hydraulically overpressured zones (Rosenberry *et al*., 2003; Glaser *et al*., 2004) that may decrease hydraulic conductivity. We feel that the vertical pore-water flow rates are too small for advection to be important. During our period of data acquisition, hydraulic head also was measured at 1, 2 and 3 m depth and the results were reported as part of larger study looking at methane ebullition at depth in the peat profile (Rosenberry *et al*., 2003). The hydraulic gradient between the 2 m piezometer and the water table was $\sim$0·2 upwards and the gradient between the 3 m piezometer and the water table was 0·02 downwards. The average pore-water velocity in the upper 2 m of peat is $\sim$4 mm day^{-1}, assuming a hydraulic conductivity of $5{\cdot}0 \times 10^{-6}$ cm s^{-1} and an effective porosity of 0·2. Slow average velocity rates in GLAP were also found by Siegel *et al*. (1995) based on conservative solute profiles.

The Peclet number for heat transport *Pe* can be used (Domenico and Palciauskas, 1973) to test the hypothesis that conduction is the primary control on vertical heat

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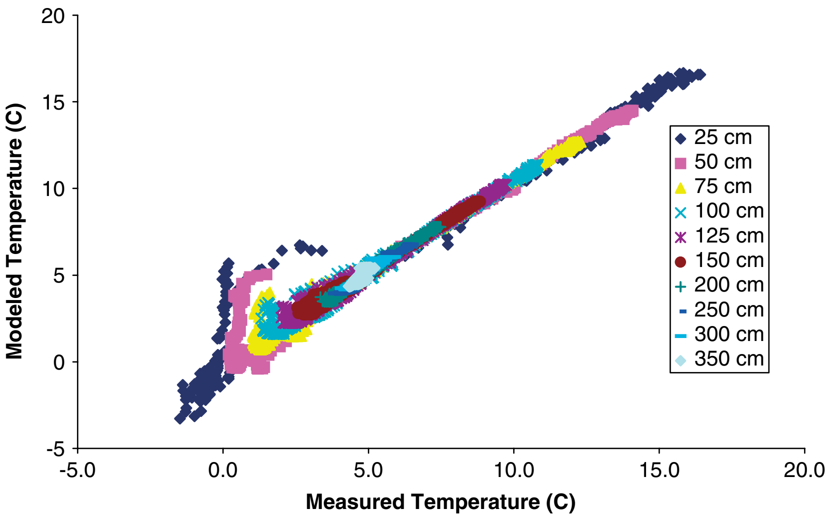

Figure 6. Comparison of the measured versus modelled temperature results for the calibrated SUTRA model for the first year of collected data (1998–1999)

transport in the Red Lake Bog:

$$Pe = \frac{\ell \overline{v}}{\alpha} \tag{4}$$

where ℓ is the characteristic length (0·2 m; taken as one-tenth of the flow regime distance), $\overline{v}$ is the vertical linear velocity of pore water (4 mm day^{-1}), and α is the bulk thermal diffusivity of the solid–fluid system ($1{\cdot}6 \times 10^{-7}$ m^2 s^{-1}) for a 90% water system. The thermal diffusivity is calculated from the bulk thermal conductivity of the peat–water system (0·58 W m^{-1} °C^{-1}) divided by the product of the bulk density (985 kg m^{-3}) and the bulk specific heat (3568 J kg^{-1} °C^{-1}). If the Peclet number is much less than one, then the system is conduction dominated; if the Peclet number is much greater than one, then the system is convection dominated. For our system the calculated Peclet number for the peat is ~0·06, indicating heat transport is primarily by conduction. In a two or three dimensional system heat transport by advection may be more important.

The most significant problem with our heat transport simulations was the lag in the modelled temperatures versus the measured temperatures during the early spring. The lag is most likely the result of a zero-curtain effect, where energy that would normally be transported downward into the bog is consumed by the latent heat of fusion of the melting ice. Evidence for this energy consumption is recorded in the heat data at the 25 cm depth. During winter, the temperature at 25 cm drops below freezing and, given the very low total dissolved solids of the pore water; it can be assumed that ice forms to at least this depth. In northern Minnesota, frost typically extends to a depth of about 50 cm in mineral soils (Minnesota Department of Transportation, Office of Materials, 2004). In the spring, the surface temperature (0 cm depth) and the air temperature rose, but the subsurface temperatures at 25 cm and below are constant and do not mimic the air temperature profile. Only after the temperature at 25 cm began to rise did the temperature at deeper depths begin to rise.

We tested the hypothesis that the heat of fusion associated with melting ice in the shallow peat caused the temperature deviation by estimating how much heat was required to melt the ice and the resulting lag in the timing of the peat temperature rise (Hall *et al.*, 2003). Assuming a unit area of ice, 1 m^2, and an ice depth of 25 cm (based on the measured heat data), a peat total porosity of 90%, an ice density of 916 g cm^{-3}, and a latent heat of 333 700 J kg^{-1}, it would require ~69 MJ to melt all of the ice. We can estimate the amount of time it would take to melt all of the ice to a 25 cm depth, assuming that the thermal conductivity of the solid peat is insignificant. The total heat conduction H is

$$H = \frac{kA(T_2 - T_1)}{L} \tag{5}$$

where the thermal conductivity $k = 1{\cdot}225$ W m^{-1} °C^{-1}, the average of the thermal conductivities of an ice and water mixture $T_1 = 0\,^\circ$C, $T_2 = 4{\cdot}3\,^\circ$C is the temperature at 0 cm depth (the average temperature at 0 cm during the lag periods of 1999 and 2000), A is the cross-sectional area, 1 m^2, and $L = 30$ cm is taken as the total depth for heat transport. Using Equation (5), the total heat conduction is calculated as 1·8 MJ day^{-1}, indicating it would take 38 days to melt 25 cm of ice. The measured field data indicate that in ~35 days the same amount of ice melted. This similarity confirms that the lag in temperature rise at the 25 cm depth is most likely caused by melting ice. McKenzie *et al.* (2006) modify the SUTRA model to include freeze-thaw processes, with applications to heat transport in peat.

The generation of methane within a shallow peat soils profile is primarily a function of temperature, the depth to the water table, and productivity (Walter and Heimann, 2000). The seasonal methane cycle in northern peatlands is dominated by temperature, as opposed to tropical wetlands, which are controlled by changes in the depth to the water table (Walter *et al.*, 2001). Although a common assumption is that soil temperatures below 50 cm depth

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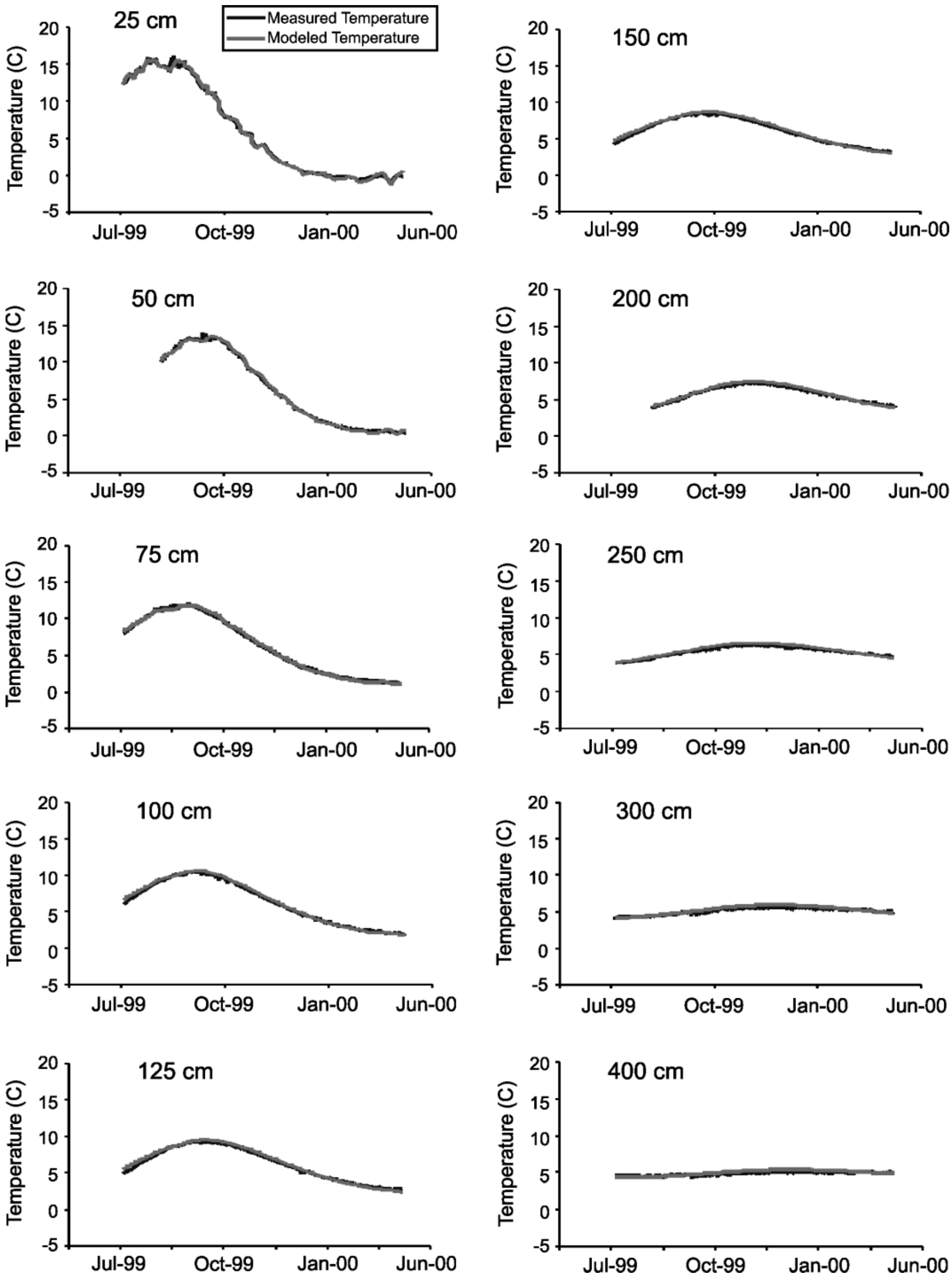

Figure 7. Comparison of the measured and modelled temperature results versus time for the evaluated SUTRA model (1999–2000) at 12 observation depths

are relatively constant in peat (e.g. Lewis, 1995), we found that temperatures fluctuated significantly at depths far greater than 50 cm, including an annual amplitude of $\sim$4 °C at a depth of 200 cm. We find large amounts of free-phase methane at this depth and deeper with GLAP (Rosenberry *et al*., 2003; Glaser *et al*., 2004). Therefore, temperature fluctuations at these deeper depths should directly affect the generation of methane (Segers, 1998).

We propose that changes in the surface heat regime, due to regional warming, would have non-linear impacts on the heat regime of the deeper zones within thicker peatlands where methane is being generated. With increased surface temperatures, the temperature of the regional groundwater that flows into the base of some peat profiles would be warmer, and in time warm soil temperatures throughout the peat profile. The observed zero-curtain effect may explain the observed delay in the onset of methanogenesis from bogs in the spring season (e.g. Boeckx and van Cleemput, 1997), because ice below the peat surface inhibits warming of deeper peat and traps gases formed in deeper peat layers. If global warming causes a decrease in the amount of ice formed in the winter in the upper peat, an increase in the rate of ice melting, and warmer temperatures within the deeper peat,

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then the result might prove to be a positive feedback mechanism because of increased methane release to the atmosphere from deep peat.

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