



## Improved methodology to quantify the temperature sensitivity of the soil heterotrophic respiration in croplands



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### ABSTRACT

Soil heterotrophic respiration ( $R_H$ ) is usually modeled using simple temperature dependence equations where the temperature sensitivity of  $R_H$  could vary for different soils and climate conditions. The temperature sensitivity is expressed as a function of the base rate of heterotrophic respiration ( $R_{H-0}$ ) and the respiration change rate over a 10 °C temperature shift ( $Q_{10}$ ). A methodology was developed to better quantify these two parameters, and was validated using seven contrasting year-site soil respiration datasets collected in wheat fields. The data were acquired using soil respiration chambers and eddy flux towers in three mid-latitude European sites and one North American site. The first step consisted in parameterizing and initializing a semi-mechanistic process-based model then validating the prediction performance using 2/3 of the datasets. The coefficient of determinations between the predictions and the observations of daily soil respiration ( $R_s$ ) was 0.71 and was 0.73 for its heterotrophic component ( $R_H$ ). The second step consisted in using the daily semi-mechanistic model predictions of  $R_H$  for each growing season and site to calibrate a simple empirical model describing  $R_H$  response to soil temperature and water content. It was shown with the contrasting years-sites that coherent results were only obtained when a common average  $Q_{10}$  value was determined prior to fit the base rate of heterotrophic respiration coefficient. Using a common  $Q_{10}$  value of 2.2 provided more stable  $R_{H-0}$  for each site over time. It reflected the strong relationship between the  $R_{H-0}$  and the slow decomposing C in the first 30-cm soil layer. The simple empirical model, which was validated using 1/3 of the data, explained between 42% and 92% of the variability of  $R_H$  over the different sites.

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### 1. Introduction

The soil CO<sub>2</sub> efflux ( $R_S$ ) is a major component of terrestrial ecosystems CO<sub>2</sub> emissions (Ryan and Law, 2005). Because crops cover about one third of the European land surface (FAOSTAT, 2010), the exchanges between crops and atmosphere are a major driver of annual atmospheric CO<sub>2</sub> fluctuations. Accurate understanding of the mechanisms that govern  $R_S$  response to climate change is essential for forecasting future changes in the terrestrial carbon balance (Buchmann, 2000; Ryan and Law, 2005). Soil respiration is the result of the production of CO<sub>2</sub> by root respiration including the rhizosphere (autotrophic respiration,  $R_A$ ) and by microbial activity related to the decomposition of soil organic matter (heterotrophic respiration,  $R_H$ ). The mean annual contribution of  $R_H$  to  $R_S$  was estimated from 45% to 70% from long-term experiments over different crops (Moureaux et al., 2006; Moyano et al., 2007; Shi et al., 2006; Zhang et al., 2013). These estimates highlight the

importance of understanding the factors controlling the decomposition processes and the associated CO<sub>2</sub> production. Studying  $R_H$  is essential for crops, as agricultural soils have been reported to lose large amount of carbon (Janssens et al., 2003; Smith, 2004), which results in increasing atmospheric CO<sub>2</sub> concentration. With regard to climate change and to increasing temperature, CO<sub>2</sub> emissions due to heterotrophic respiration may become more important (Davidson and Janssens, 2006). Moreover, the recently introduced carbon sequestration program for agriculture, named "4 per 1000", which aims to adapt agricultural practices with the goal of storing carbon more efficiently in the soil justify the necessity to better quantify soil fluxes components and to better understand their response to climate variations and to agricultural management.

Studies on the contribution of heterotrophic source are difficult to carry out because of the uncertainty involved in separating experimentally root respiration fluxes ( $R_{Ab}$ , belowground autotrophic respiration) from  $R_H$  in the field (Mäkiranta et al., 2008; Moyano et al., 2007; Subke et al., 2006). Thus most of previous studies used models to estimate the heterotrophic contribution to  $R_s$ . Most of the models used for simulating

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soil organic matter (SOM) decomposition partition SOM into several pools according to their potential decomposition rates, reflecting the mean residence time of carbon compounds in the soil profile. Kinetics of SOM decomposition is calculated as a combination of SOM quality (i.e., inherent stability of carbon compounds) and the environmental constraints (i.e., soil texture, soil moisture and temperature). The two well-known semi-mechanistic models CENTURY (Parton et al., 1987, 1988) and Roth-C (Coleman and Jenkinson, 1995) use at least five pools. They include two compartments of plant material in the litter layer and three pools into the mineral soil (active, slow, and passive) with residence time varying from a few months to several hundreds of years. Previous studies have shown that these models successfully simulate the long-term dynamics of carbon stocks in soils of different ecosystems and also in the case of land use change (Kelly et al., 1997; Smith et al., 1997). However, these models remain difficult to implement, especially when developed for the daily time scale (Parton et al., 1998) as they require to document a lot of input parameters and to define initial conditions. Modeling soil respiration at the daily time scale is essential to understand the impact of climate variations and management practices on soil CO<sub>2</sub> fluxes.

The common approach to predict soil CO<sub>2</sub> fluxes at the daily time scale consists in expressing soil respiration flux as an exponential function of soil temperature. The temperature sensitivity is calculated using a Q<sub>10</sub> temperature coefficient which represents the respiration change rate over a 10 °C temperature shift. Lloyd and Taylor (1994) concluded that the Q<sub>10</sub> model is not well-suited for fitting soil CO<sub>2</sub> fluxes over a wide range of temperature. They suggested the use of an Arrhenius-type equation because it takes into account the varying temperature sensitivity of soil respiration. Their conclusion was made using measurements from various ecosystems. However, Q<sub>10</sub> is usually determined at local scale and requires dataset from a specific experimentation or field campaign. Values found in literature vary widely among different ecosystems (Raich and Schlesinger, 1992). This variability is mainly associated to a difference in the proportion of resistant against easily degradable (labile) carbon pools. Moreover, small but significant ranges of Q<sub>10</sub> variation are reported even for the same type of land use: for wheat crops, the Q<sub>10</sub> coefficient varies from 1.9 to 2.5 (Moyano et al., 2007; Shi et al., 2006; Suleau et al., 2011). As the complex organic substrate is assumed to be similar to the plant litter, these differences among sites could be attributed to either an effect of the soil texture in the protection against the decomposition (clay content), or to an effect of soil water content and/or soil temperature range due to the climate conditions. Seasonal variations of Q<sub>10</sub> associated with the soil temperature range were previously reported and discussed by Del Grosso et al. (2005), Janssens and Pilegaard (2003), Qi and Xu (2001), and Tjoelker et al. (2001). However, it remains unclear whether the Q<sub>10</sub> variability either reflects actual difference in the temperature sensitivity of the SOM decomposition or is a mathematical effect caused by adjusting the coefficients of the equation. The determination of the statistical parameters is often performed simultaneously for both Q<sub>10</sub> and the base rate of heterotrophic respiration (R<sub>H</sub> – <sub>0</sub> defined at 0 °C) which makes the coefficients interpretation difficult. This model adjustment approach does not follow the basic assumptions found in most semi-mechanistic SOM models, where sensitivity to temperature is considered constant and identical for all carbon pools and independent of ecosystem types and climate zones.

Moreover, the conceptual difference between the temperature sensitivity of soil CO<sub>2</sub> efflux (i.e., the efflux Q<sub>10</sub>) and the temperature sensitivity of soil CO<sub>2</sub> production (i.e., the production Q<sub>10</sub>) needs to be explored. As the production sensitivity of R<sub>H</sub> corresponds to the absolute change in the decomposition rates (in SOM models) with respect to a change in soil temperature, several environmental constraints on respiration could modulate the production temperature sensitivity resulting in an efflux temperature sensitivity corresponding to the observed temperature response of soil CO<sub>2</sub> efflux (Davidson et al., 2006).

In this study, we compared the daily outputs of R<sub>S</sub> and its heterotrophic component from a semi-mechanistic model in wheat cultivated land pertaining to different soil and climate conditions to several datasets acquired on wheat in four contrasting mid-latitude regions (issued from three European sites and one North American site). Following their validation, the R<sub>H</sub> outputs of the semi-mechanistic SOM model were used as continuous datasets to calibrate the coefficients of a simple model combining temperature and water content sensitivity by either (1) allowing the two coefficients (i.e., R<sub>H</sub> – <sub>0</sub> and Q<sub>10</sub>) to vary or (2) assuming a constant temperature sensitivity. Our objectives were (1) to quantify the change of the base rate of heterotrophic respiration (R<sub>H</sub> – <sub>0</sub>) and the respiration change rate over a 10 °C temperature shift (Q<sub>10</sub>) in response to temperature and water content and to incorporate them in a simple equation describing the heterotrophic respiration (R<sub>H</sub>) and (2) to verify whether the temperature sensitivity of R<sub>H</sub> varies over different soils and climatic conditions. We also verified the magnitude of the difference between CO<sub>2</sub> production and CO<sub>2</sub> surface efflux temperature sensitivities and we investigated the change in R<sub>H</sub> – <sub>0</sub> rates between sites according to the quantity and/or quality of soil organic carbon.

## 2. Material and methods

### 2.1. Semi-mechanistic modeling approach

In the semi-mechanistic SOM model, R<sub>S</sub> was calculated as the sum of R<sub>H</sub> and R<sub>Ab</sub> (belowground autotrophic respiration).

#### 2.1.1. Heterotrophic source

The soil organic carbon sub-model (SOC) used to simulate the dynamic of heterotrophic respiration is based on the soil organic matter (SOM) sub-model of CENTURY (Parton et al., 1987). The original version developed by Parton et al. (1987) has been successfully adapted at a daily time step (i.e., DAYCENT) over different ecosystems (Del Grosso et al., 2005; Epron et al., 2001).

Soil organic carbon is divided into three major components including active, slow and passive soil carbon. The “active” pool includes live soil microbes plus microbial products; the “slow” one includes mainly resistant plant material (lignin-derived material). The passive material is very resistant to decomposition, i.e., physically and chemically stabilized SOM. The model also includes a surface microbial pool, which is associated with decomposing surface litter (mainly leaf litter). Carbon flows between these pools are controlled by decomposition rate (Table 1) and microbial respiration loss parameters (Fig. 1), both of which are a function of soil texture, soil temperature and soil water content.

In this soil model, all the fractions are located in the soil vertical profile (Fig. 1). The soil is divided into four layers according to the vertical distribution of carbon content and soil texture, and crop management (i.e., depth of plowing): a surface layer, a superficial soil layer (from surface to 15 cm depth) and two deeper soil layers (from 15 to 30 cm and from 30 to 45 cm). No carbon migration is assumed between the superficial and the deep soil layers. The “slow

**Table 1**

Maximum decomposition rate (K, day<sup>-1</sup>) of the soil organic matter for each C pools of the semi-mechanistic model (Parton et al., 1988).

Pool	K
Soil metabolic fraction	5.07 × 10 <sup>-2</sup>
Surface metabolic fraction	4.05 × 10 <sup>-2</sup>
Soil active C	2.00 × 10 <sup>-2</sup>
Surface active C	1.64 × 10 <sup>-2</sup>
Soil structural fraction	1.34 × 10 <sup>-2</sup>
Surface structural fraction	1.07 × 10 <sup>-2</sup>
Slow C	5.48 × 10 <sup>-4</sup>
Passive C	1.23 × 10 <sup>-5</sup>

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