Soil carbon flux research in the Asian region: Review and future perspectives

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Abstract

Soil respiration \((R_s)\) is the largest flux of carbon dioxide \((\text{CO}_2)\) next to photosynthesis in terrestrial ecosystems. With the absorption of atmospheric methane \((\text{CH}_4)\), upland soils become a large \text{CO}_2 source and \text{CH}_4 sink. These soil carbon \((C)\) fluxes are key factors in the mitigation and adaption of future climate change. The Asian region spans an extensive area from the northern boreal to tropical regions in Southeast Asia. As this region is characterised by highly diverse ecosystems, it is expected to experience the strong impact of ecosystem responses to global climate change. For the past two decades, researchers in the AsiaFlux community have meaningfully contributed to improve the current understanding of soil C dynamics, response of soil C fluxes to disturbances and climate change, and regional and global estimation based on model analysis. This review focuses on five important aspects: 1) the historical methodology for soil C flux measurement; 2) responses of soil C flux components to environmental factors; 3) soil C fluxes in typical ecosystems in Asia; 4) the influence of disturbance and climate change on soil C fluxes; and 5) model analysis and the estimation of soil C fluxes in research largely focused in Asia.

**Key words:** Chamber, Climate change, Disturbance, Methane, Soil respiration

1. Introduction

Based on the fifth Intergovernmental Panel on Climate Change (IPCC) report, global temperatures are predicted to increase between 2.6 and 4.8 °C, as per the Representative Concentration Pathway 8.5 (RCP8.5) scenario by 2100 (IPCC, 2013). Climate change mitigation and adaptation requires a better understanding of the global carbon \((C)\) cycle, including proposals for necessary preventative measures to avoid further climate change. Generally, the top layer of soils contain approximately 3000 Gt C of soil organic carbon (SOC) up to a depth of 2 m (Sanderman et al., 2017). This amount is approximately 3.5 and 6.5 times the C in the atmosphere (860 Gt C, Friedlingstein et al., 2019) and the global plant biomass (450 Gt C, Erb et al., 2018), respectively. This huge amount of SOC and its dynamics are key elements in the global C cycle, and thus, have huge influence on the future of climate change. SOC is decomposed by soil microbes and is released into the atmosphere as \text{CO}_2 (heterotrophic respiration, \(R_h\)). Soil respiration \((R_s)\), the sum of \(R_h\) and plant root respiration \((R_r)\), is the largest \text{CO}_2 flux next to photosynthesis in terrestrial ecosystems (Raich and Schlesinger, 1992). However, upland soils are a sink for atmospheric methane (\text{CH}_4) and an important component recognised in the global \text{CH}_4 budget (Ni and Groffman, 2018). As such, soils (and particularly forest soils) are a major \text{CO}_2 source and \text{CH}_4 sink. Synthetic analysis based on in-situ observations of soil C flux in various ecosystems has been proposed as a straightforward solution to estimate global soil C fluxes. There are also recent reports on the global estimation of \(R_s\) based on site-observation data (Bond-Lamberty and Thomson, 2010; Hashimoto et al., 2015). In recent years, advancements in instrumentation used for observation have also facilitated the continuous field measurement of soil \text{CH}_4 flux (Sakabe et al., 2015; Ueyama et al., 2015; Ishikura et al., 2019).

Asian terrestrial ecosystems occupy vast areas which includes tropical forests and wetlands in Southeast Asia, boreal ecosystems in Northeast Asia and the alpine ecosystems on the Tibetan Plateau. These ecosystems make significant contributions to the regional and global C budgets. Accurately quantifying the
CO$_2$/CH$_4$ balance is critical to set emissions reductions targets in these regions and identify and promote effective climate change mitigation strategies. For the past two decades, the Asian observation network and the flux community (AsiaFlux) have made substantial contributions to advancements in soil C flux research. This paper reviews the history of soil C flux research, recent advances in monitoring and modelling approaches and describes future perspectives, focusing on the Asian region.

2. Methodology

2.1 Introduction to the chamber method

In general, soil heterogeneity and impermeability creates challenges in carrying out precise observations of soil C flux using remote sensing technology (Bond-Lamberty and Thomson, 2010). Despite the publication of several studies locally estimating soil C flux using the remote sensing approach (e.g. Wu et al., 2014; Yan et al., 2020), the calibration and validation of those estimations based on in-situ observation data for soil C flux are essential. Soil C flux may be determined through application of the micro-meteorological (Missen et al., 2007) and soil CO$_2$ gradient methods using CO$_2$ probes (Liang et al., 2004; Pingintha et al., 2010). At present, the chamber method continues to be the most common method used in soil C flux research (Lankreijer et al., 2003). This section describes the origin and progress associated with this method for soil C flux research from the field measurement perspective.

The chamber method essentially consists of covering the soil surface with a sealed container (chamber) for several minutes to hours to determine soil greenhouse gas (GHG) fluxes. This determination is based on the increased ratio (slope) of GHG concentrations during the measurement period, or the difference in GHG concentrations between ambient air and the sampled air within the chamber (see Pumpanen et al., 2010). The chamber method has been classified into two methods; the static chamber and dynamic chamber methods. In the former, there is no inflow of ambient air inside the chamber. In contrast, the dynamic chamber method involves the circulation of air between the chamber and the analyser using a pump. Sections 2.2 and 2.3 provide a detailed description of these chamber methods.

2.2 Static chamber method

In the static chamber method, $R_c$ increases the CO$_2$ concentration inside the chamber; this may be measured using several approaches such as the alkali absorption method, gas chromatograph (GC), or infrared gas analyser (IRGA). Lundegårdh (1927) began field measurements of $R_c$ using the static chamber method (soil respiration bell) and the alkali absorption method. The alkali absorption method had been used for a long period until recently. In the alkali absorption method, the alkali medium (solid medium such as soda lime or liquid medium such as NaOH and KOH), inside the static chamber absorbs CO$_2$. This absorbed CO$_2$ may be measured by the increased weight of the alkali medium (soda lime) or by titration with an acid to calculate $R_c$ (Lundegårdh, 1927; Edwards, 1982).

Kirita (1971) improved CO$_2$ absorption efficiency by soaking a sponge in the liquid alkali. This was intended to increase the absorptive area of the alkali medium inside chamber. This modified method has been widely used to measure $R_c$ in the Asian region (Nakane et al., 1984; Tulaphtitak et al., 1985; Koizumi et al., 1993). One of the merits of the alkali absorption method is its low cost as this method does not need an expensive analyser (Table 1, Schiedung et al., 2016).

However, the precision of the alkali absorption method has been called into question following the application of the IRGA (Witkamp, 1969; Kucera and Kirkham, 1971; Edwards and Sollins, 1973). Edwards and Sollins (1973) tested the precision of the alkali absorption method during the summer season in a poplar forest in eastern Tennessee, finding that the $R_c$ measured with this method was approximately 63% of that measured with the IRGA (open flow method described below). The increasing number of studies determining the bias of the alkali absorption method (Cropper et al., 1985; Ewel et al., 1987; Nakadai et al., 1993; Jensen et al., 1996; Bekku et al., 1997; Yim et al., 2002) have demonstrated that this method tends to overestimate low $R_c$ and underestimate high $R_c$. Two possible factors were identified in association with this bias (Yim et al., 2002). The first factor was the limited CO$_2$ absorption efficiency of the alkali medium and the resulting CO$_2$ gradient between soil and chamber air (Jensen et al., 1996; Yim et al., 2002). The second factor was the suppression of microbial (Koizumi et al., 1991) and plant root respiration (Qi et al., 1994) under high CO$_2$ concentrations within the chamber. In studies conducted in Euro-American countries (e.g. Cropper et al., 1985; Ewel et al., 1987; Jensen et al., 1996), the conventional alkali absorption method was used to compare the measurements using IRGA. In contrast, studies conducted in the Asian region (mainly in Japan) used the protocol in Kirita (1971) to cast doubt over the precision of the alkali absorption method (Koizumi et al., 1993; Bekku et al., 1997; Yim et al., 2002). Despite the bias being unresolved using the protocol of Kirita (1971), Yim et al. (2002) suggested that the magnitude of underestimation for $R_c$ under high CO$_2$ condition should be relatively mild with this protocol of Kirita (1971) compared to other alkali absorption methods because of its increased CO$_2$ absorptive efficiency. Following the publication of the research on the bias of alkali absorption method for $R_c$ measurement, other chamber methods that applied GC or IRGA for gas analysis had become increasingly popular as opposed to the alkali absorption method.

The closed chamber method, was also considered as an effective method to measure $R_c$ instead of alkali absorption method. In this method, air within the static chamber is sampled several times using a syringe according to the time course during the measurement period; the CO$_2$ concentration of sampled air is then analysed by GC or IRGA (Bekku et al., 1995). This method was first introduced as a simple method to measure nitrous oxide (N$_2$O) flux using GC (Matthias et al., 1980; Hutchinson and Mosier, 1981). Following this, the method had become popular for use in many types of ecosystems because of its simplicity and applicability (Mariko et al., 1994; Bekku et al., 1995; 1997; Tokida et al., 2013). Whilst this method needs an IRGA or a GC to measure the CO$_2$ concentration of the sample air in the laboratory, it does not need any chemicals in the field (Table 1). Recent studies have also demonstrated the application of closed chambers equipped with a portable diffusion-based
IRGA (e.g. GMP343, Vaisala, Helsinki, Finland), battery, and data logger (Inoue et al., 2012; Noh et al., 2016a; 2016b; Wakhid et al., 2017). In such experiments, the \( R_s \) measurement and data recording (\( CO_2 \) concentration in the closed chamber was automatically recorded every 5 to 10 s), were carried out concurrently in the field.

2.3 Dynamic chamber method

In the dynamic chamber method, negative pressure within the chamber compared with outside the chamber renders an overestimate of \( R_s \) whilst a positive pressure results in the underestimation of \( R_s \) (Fang and Moncrieff, 1998; Lund et al., 1999). Therefore, techniques to avoid this pressure difference are necessary. The dynamic chamber method is divided into two categories; the open and closed dynamic chamber methods. The open flow method is one of the open dynamic chamber methods, and the application of this method using IRGA had commenced from the late 1960s (Reiners, 1968; Witkamp, 1969; Kucera and Kirkham, 1971). This method involves the constant flow of ambient air into the chamber (usually using a buffer tank), where \( CO_2 \) concentrations are measured at the inlet and outlet of the chamber (Liang et al., 2003). The \( R_s \) is determined based on the difference in the inlet and outlet \( CO_2 \) concentration and the flow rate of air passing through the chamber (inlet and outlet) (Fang et al., 2004; Suh et al., 2006). This method enables continuous measurement in a steady state (Suh et al., 2006; Yasutake et al., 2014); however, it is necessary to precisely control flow rate such that it is equal at the inlet and outlet to avoid pressure differences between the inside and outside of the chamber (Table 1).

The open-top chamber method (a type of open dynamic chamber method) was developed to address this pressure differential issue (Fang and Moncrieff, 1998). In this method, it was possible to minimise the influence of the pressure differential between the inside and outside of the chamber by opening the headspace of the chamber (Fang and Moncrieff, 1998; Liang et al., 2004). Additionally, this method was advantageous as it did not need a buffer tank. However, the downside to the open-top chamber method was that it tends to be influenced by wind (Table 1, Fang and Moncrieff, 1998; Liang et al., 2004).

The application of the closed dynamic chamber to measure \( R_s \) was a relatively recent development compared with the other chamber methods (Hall et al., 1990; Rochette et al., 1991; 1992; 1997). The popularisation of the compact IRGA had greatly influenced the applicability of the closed dynamic method (and several types of dynamic chamber methods) to measure \( R_s \) (Rochette et al., 1997). In this method, only the air sample collected in the closed chamber is led into the \( CO_2 \) analyser with no ambient air flow into the chamber (Rochette et al., 1997). The measurement is conducted in a non-steady state under increasing \( CO_2 \) concentrations inside the chamber due to \( R_s \). There are several commercial portable-type closed dynamic chamber systems such as the Li6400 (equipped with a soil respiration chamber 6000–09, Li-cor, Lincoln, NE, USA), Li8100 (Li-cor), or EGM-5 (equipped with a soil respiration chamber SRC-2, PP-Systems, Amesbury, MA, USA); many studies have used these systems (Wang et al., 2005; Ohashi et al., 2008; Adachi et al., 2009; Liu et al., 2016a; Li et al., 2019b; Qin et al., 2019). In recent research, the originally developed portable-type closed dynamic chamber systems have also been used in the field (Takada et al., 2015; Sun et al., 2017; 2020a; Gao et al., 2019; Zhao et al., 2021).

2.4 Continuous measurements using automatic opening and closing chamber systems

The manual measurement for soil C flux using the chamber method require labour and time, particularly if long-term measurements in steady-state with minimum pressure difference in chamber may cause pressure difference and unbalance between inflow and outflow of air in chamber may cause pressure difference and produce large errors for flux values. Therefore, the open-top chamber method is rarely used in recent studies for soil C flux measurement. At high flow rates in chamber, there is a potential for these errors, this method is rarely used in recent studies for soil C flux measurement.

Table 1. Chamber methods for soil C flux measurement.

| Static/Dynamic | Open/Closed | Method                  | Period          | Advantages                                      | Disadvantages                                      |
|----------------|-------------|-------------------------|-----------------|-------------------------------------------------|---------------------------------------------------|
| Static         | Closed      | Alkali absorption method | From 1920s in  | Low cost                                        | Overestimation at low \( R_s \) or underestimation (at high \( R_s \)) for \( R_s \) |
|                |             |                         | the field      | No need for electrical power source in field    | Need for chemical analysis in laboratory           |
|                | Closed      | Closed chamber method   | From 1980s     | Possible to concurrently measure several types  | Need for GC or analyser (cost)                     |
|                |             |                         |                 | of GHG fluxes \( CO_2, CH_4, \) and \( N_2O \) | Need for manual gas sampling several times in     |
|                |             |                         |                 | with GC                                          | one measurement and gas analysis in laboratory     |
|                | Closed      | Closed dynamic chamber  | From 1990s     | High versatility in many ecosystems (e.g. forests, | Need for analyser (cost) and power source         |
|                | Dynamic     | method                  |                 | Relatively simpler flow line compared with open  | typically battery in field measurements             |
|                |             |                         |                 | dynamic chamber systems                         | Pressure difference between inside and outside of  |
|                | Open        | Open flow method        | From late 1960s | Possible to concurrently measure GHG fluxes      | Typically requires a buffer tank                   |
|                |             |                         |                 | (depending on the analyser) and record data in    | Unbalance between inflow and outflow of air in     |
|                |             |                         |                 | the field                                        | chamber may cause pressure difference and          |
|                | Open        | Open top chamber method | From late 1990s | Measurement in steady-state with minimum pressure| Strong wind may produce errors for flux values     |
|                |             |                         |                 | difference between inside and outside of chamber | Due to the potential for these errors, this        |
|                |             |                         |                 | No need for buffer tank                          | method is rarely used in recent studies for soil   |
|                | Open/       | AOCC method             | From 1970s      | Suitable for long-term continuous measurements  | Expensive system installation                     |
| Open/Closed    |             |                         |                 | for several GHG fluxes (depending on the analyser)| Need for stable electrical power source (usually  |
|                |             |                         |                 | because of minimal difference in environmental   | commercial power supply) for measurement          |
|                |             |                         |                 | factors between inside and outside of chamber    |                                                   |
|                |             |                         |                 | Efficient data collection with high resolution   |                                                   |

* If the system was equipped with diffusion-based IRGA, battery and datalogger as per Noh et al. (2016a; 2016b) and Wakhid et al. 2017, this will be not necessary
continuous data are required for a specific study site; the automatic opening and closing chamber (AOCC) method addressed this issue. In the AOCC method, several automatic chambers are set at specific locations in the field. This method is effective for the sensitive detection of seasonal and inter-annual changes in soil C flux, and its response to variations in environmental factors (temperature and moisture content of the soil), with minimal changes in litterfall, rainfall, and temperature between the inside and outside of the chamber (Liang et al., 2003). The concept for the AOCC method had been introduced in the 1970s, and Edwards (1974) reported the diurnal change of \( R_\text{ch} \) using the method in a mixed deciduous forest in east Tennessee. Recently in the Asian region, continuous and high frequency measurements for soil C flux has been conducted in several sites using AOCC systems (Liang et al., 2004; 2010; Suh et al., 2006; Hirano et al., 2009; Lee, 2011; Sundari et al., 2012; Aguilos et al., 2013; Tan et al., 2013; Yu et al., 2013; 2020; Ueyama et al., 2015; Guo et al., 2016a; 2016b; Teramoto et al., 2016a; 2017; 2018; 2019a; Wu et al., 2016; Eom et al., 2018; Ishikura et al., 2018; Yan et al., 2019b; Sun et al., 2020b). These systems usually employ the open or closed dynamic chamber methods. There is also a commercial model of the AOCC system that is capable of continuous and high frequency measurements over a long period of time (Li8100A equipped with multiplexer Li8150 and long-term chambers 8100-104 or 8100-104c, Li-cor); however, the deployment of this system is relatively expensive (Table 1).

2.5 Soil \( \text{CH}_4 \) flux measurements using chamber methods

Soil \( \text{CH}_4 \) flux may be also measured using the chamber methods. Until recently, soil \( \text{CH}_4 \) flux in natural ecosystems was mainly measured using the closed chamber method via GC (Ishizuka et al., 2000; 2009; Morishita et al., 2007; Kim, 2013; Liu et al., 2016b). However, the development of a compact and commercially feasible \( \text{CH}_4 \) analyser has enabled continuous field measurements for soil \( \text{CH}_4 \) flux. At present, such analysers have employed laser-based technology (e.g., cavity ring-down spectroscopy) for use in field measurement for soil GHG flux for several natural ecosystems (Dhandapani et al., 2019). In addition, several recent studies in the Asian region have conducted continuous measurements of soil \( \text{CH}_4 \) flux using the \( \text{CH}_4 \) analyser and AOCC systems (Sakabe et al., 2015; Ueyama et al., 2015; Ishikura et al., 2019). To fully utilise the sporadic measurement data to estimate the annual cumulative flux of soil \( \text{CH}_4 \), Yang et al. (2019a) developed equations using 154 site-years monthly \( \text{CH}_4 \) flux data measured using the chamber methods in China. Such an approach will contribute to the improved estimation of the regional and global soil \( \text{CH}_4 \) flux. It is also expected that the observation network in the Asian region for soil \( \text{CO}_2/\text{CH}_4 \) flux measurements will continue to expand, facilitating a better understanding of soil C dynamics.

2.6 \( \text{CH}_4 \) flux measurements in paddy fields

Based on the latest estimation by Jackson et al. (2020), agriculture and waste constitute approximately one-third of the total global \( \text{CH}_4 \) emissions. Rice cultivation (through paddy fields), is one of the major \( \text{CH}_4 \) sources in the agricultural industry (Yang et al., 2009; Zhang et al., 2016), and more than 90% of \( \text{CH}_4 \) emissions from paddy fields has been suggested to originate from Asian monsoon regions (Yang et al., 2009). The chamber method has also been applied to measure \( \text{CH}_4 \) flux in paddy fields. The most popular method to measure \( \text{CH}_4 \) flux is the closed chamber method using a GC equipped with a flame ionisation detector (Gaihre et al., 2011; Tokida et al., 2013; Huang et al., 2019); the detailed protocol for this method is described by Minamikawa (2015). AOCC systems have also been used in paddy fields to measure GHG flux (Schütz et al., 1989; Bronson et al., 1997; Komiya et al., 2015). For example, Bronson et al. (1997) collected continuous data from December 1992 to April 1994 for \( \text{CH}_4 \) and \( \text{N}_2\text{O} \) in paddy fields located in Los Banos, the Philippines. In their study, AOCCs and GC were connected via tubing for gas analysis. In a recent study by Komiya et al. (2015), an AOCC was connected to a laser-based \( \text{CH}_4/\text{CO}_2 \) analyser (G2201-i, Picarro Inc., Santa Clara, CA, USA) to obtain two-day continuous data in tropical paddy field located in Thailand. They evaluated \( \text{CH}_4 \) emissions by bubble ebullition and diffusion separately, and indicated the controlling factors for \( \text{CH}_4 \) ebullition (i.e., atmospheric air pressure and soil surface temperature). Continuous and high frequency measurements of \( \text{CH}_4 \) concentrations (3.6 s interval for each data point) using the analyser and AOCC system contributed to their findings. In such continuous measurements, appropriate dehumidification (e.g., a cooling trap, membrane dryer or granular magnesium perchlorate), is essential to avoid condensation in the flow line because of high humidity from sampling air around flooded paddy fields (Schütz et al., 1989; Komiya et al., 2015).

3. Soil C flux and its response to environmental factors

3.1 \( R_\text{c} \) components and separation

The process of \( R_\text{c} \) is defined as the release of \( \text{CO}_2 \) from the soil to the atmosphere. \( R_\text{c} \) originates from different sources; microbial decomposition of SOC (\( R_\text{d} \)) and respiration by living roots (\( R_\text{r} \)) are the two main sources. Although soil fauna respiration, chemical oxidation, and carbonate dissolution may also contribute to the total flux (Burton and Beauchamp, 1994), it is normally too small to be considered. Field-based data have shown that the contribution of \( R_\text{r} \) to total \( R_\text{c} \) in forest ecosystems largely varied from 10% to 90% (averaging 48.6%), with the data exhibiting a normal distribution (Hanson et al., 2000). This variation may potentially be due to the seasonal change of the contribution rate. For example, Lee et al. (2003) indicated that the contribution rate of \( R_\text{r} \) to \( R_\text{c} \), in a cool-temperate deciduous forest in central Japan, seasonally varied from 27% to 71%, associating this variation to abiotic factors such as the temperature and phonology of vegetation. In addition, ecosystem and vegetation types may also contribute to the variation of the \( R_\text{r} \) contribution rate. Subke et al. (2006) suggested that contribution rate of \( R_\text{r} \) to \( R_\text{c} \) was relatively high in temperate and tropical forests compared with boreal coniferous forests. The factors, \( R_\text{r} \) and \( R_\text{d} \), responded differently to temperature increase, resulting in different Q_N values (where Q_N is the multiple of the \( R_\text{r} \) increase ratio due to a 10 °C temperature rise, representing the temperature sensitivity of \( R_\text{r} \)) (Rey et al., 2002). This potentially alters the net soil C flux, potential C sequestration, and climate change associated feedbacks. The
increase in \( R \), may reflect increased C inputs to the soil through photosynthesis (Högberg et al., 2001), while increased \( R \) may reduce the C storage potential in soil (Grace, 2004). For example, whilst tropical peatlands contain large amounts of C in the peat soil, the increase of \( R \) (peat decomposition) due to land-use changes accompanied with drainage usually results in net soil C loss in the ecosystem (Hergoualc’h and Verchot, 2011; detail discussed in Sections 4.4 and 5.3). Therefore, separating the components of \( R \) is important to predict the ecosystem response to climate change, understand the feedback between climate change and soil processes, interpret microbial community dynamics from an ecologically meaningful perspective and model climate change with the \( R \) components. However, the complete \textit{in-situ} separation of \( R \) (usually root and rhizosphere) from \( R \) is quite difficult. Many approaches focused on the separation of \( R \) components have been reported; some typical methods are briefly discussed by citing previous studies (also see Hanson et al., 2000; Baggs, 2006; Kuzyakov and Larionova, 2006).

In the component integration method, CO\(_2\) effluxes of each flux component such as \( R \), litter and SOC decomposition are measured separately in a laboratory to estimate contribution of \( R \) and \( R \) to \( R \) (Hanson et al., 2000; Baggs, 2006). For example, Noh et al. (2011) incubated intact roots of \textit{Pinus densiflora}, soil (including roots) and soil alone in root chambers in a laboratory. They estimated that \( R \) contributed 53–58% of \( R \) by measuring the CO\(_2\) efflux in each component. Although this method is simple and low-cost, the \textit{in-situ} evaluation of each flux component is difficult (Table 2). In particular, the root-microbial interaction and substrate supply to the microbial community is likely to differ from field conditions (Trumbore, 2006).

The root regression method estimates the contribution of \( R \) to \( R \), based on the linear regression between the root biomass (x-axis) and \( R \) (y-axis), where the intercept of the y-axis is be estimated as \( R \) (Kucerca and Kirkham, 1971; Behera et al., 1990; Tomotsune et al., 2013). This method is simple and applicable to many study sites without soil and rhizosphere structure disturbance prior to the measurement of \( R \). The disadvantage is that the uncertainty may be large when the coefficient of the regression is relatively small (Table 2, Wang et al., 2008).

The root exclusion method estimates the contribution of \( R \) and \( R \) to \( R \) by comparing soil CO\(_2\) efflux between root contained and root removed plots. Hanson et al. (2000) categorised the root exclusion procedure into three methods; root removal, trenching, and gap analysis. Among these methods, it is essential to account for the fact that root exclusion may potentially increase soil moisture due to the absence of soil water absorption by plant roots; this altered soil moisture may influence SOC decomposition (Hanson et al., 2000; Baggs, 2006; Kuzyakov and Larionova, 2006). The root removal method provides an opportunity to exclude the influence of roots in measurement plots without any dead roots; however, this method disturbs the soil structure (Hanson et al., 2000; Baggs, 2006; Kuzyakov and Larionova, 2006). Trenching, the most popular method used in field studies, will be discussed in further detail in the later part of this Section 3.1. In gap analysis, the \( R \) to \( R \) contribution is estimated by measuring soil CO\(_2\) efflux in gaps (natural or artificial) and undisturbed areas of the forest. Whilst this method is very simple and does not require additional labour following gap establishment, the gap size requires careful examination. Although large gaps will sufficiently exclude the influence of roots on \( R \), it may changes environmental factors such as the air and soil temperatures (Nuruddin and Tokiman, 2005). Gaps that are too small may experience difficulty in excluding the influence of roots to a sufficient extent, despite minimal changes in the environmental factors. In regard to this point, Ohashi et al. (2000) suggested that a gap of 2.5 × 2.5 m in an artificial Japanese cedar forest enabled the estimation of an \( R \), contribution of 49–57% to \( R \), without significant changes in soil temperature and moisture (Table 2).

The girdling method has been suggested by Högberg et al.

| Table 2. Methods to separate the \( R \) components. |
|---------------------------------|---------------------------------|---------------------------------|---------------------------------|
| **Category**        | **Method**        | **Advantages**                                                    | **Disadvantages**                                                    |
| Component integration | Component integration | Simplicity and low cost.                                        | Difficulty in evaluating each flux component \textit{in-situ}.        |
|                     |                   | Simplicity and applicability to many study sites without disturbing the soil and rhizosphere structure prior to measuring \( R \). | Difficulty in evaluating the root-microbial interaction and substrate supply to the microbial community in field conditions. |
| Root regression      | Root regression   | Exclusion of \( R \) without any dead roots in measurement plot. | Uncertainty may be large when coefficient of regression is small. |
| Root removal         | Trenching         | Convenient and practical method in field with minimum disturbance. | Initial carbon input due to dead roots and the decomposition is inevitable, whereas root litter input is eliminated after trenching. Possible increase of soil moisture. |
| Root exclusion       | Gap analysis      | Less labour cost and simplicity. No need for additional labour after the establishment of gaps. | Difficulty in determining ideal gap size. Gap must be large enough that roots from surrounding vegetation are not in the measurement area, whilst not too large so as to change the physical soil environment. Potential to increase soil moisture. |
| Girdling             | Girdling          | Applicability to field study without disturbing the soil and rhizosphere. | Uncertainty as to the extent to which recently fixed carbon by trees contributes to \( R \). |
| Isotopic approaches  | Natural \(^{13}\)C abundance | Applicability to field study without disturbing the soil and rhizosphere. Possible to trace C dynamics and allocation in field quantitatively for each flux component. | Complexity of techniques for analysis and experimental cost, Tracers and analysing devices especially AMS for \(^{13}\)C analysis is expensive. Use of \(^{14}\)C in the field is strictly limited due to safety concerns. |

[^13C]: Carbon-13
[^14C]: Carbon-14
(2001) as an indirect method based on paring the phloem of a tree trunk to suppress translocation of photosynthate from shoot to root and rhizosphere. Theoretically, this method is able to minimise \( R_s \) without disturbing the soil and rhizosphere; however, there is uncertainty as to the extent to which recently fixed C from trees contributes to \( R_s \) (Table 2, Trumbore, 2006).

Isotopic approaches are also effective methods to separate the components of \( R_s \). The advantage of isotopic methods is that it is able to evaluate \( R_s \) and \( R_t \) separately in a field study without disturbing the soil and rhizosphere. The disadvantage of isotopic method is the complex techniques required for analysis and its high experimental cost (Table 2, Baggs, 2006). For \( ^{13}C \) analysis, the contribution of \( R_s \) to \( R_t \) is calculated based on the difference of \( ^{13}C \) (\( \% \)) in each flux component (\( R_t \), litter and SOC decomposition) (Rochette et al., 1999; Albanito et al., 2012; Oggle and Pendall, 2015). Natural or labelled \( ^{13}C \) may be analysed using this method. Rochette et al. (1999) measured the natural abundance of \( ^{13}C \) in \( R_s \) and showed that rhizosphere respiration accounted for 45\% of \( R_t \) in cropland (maize, the C4 plant, was planted on the cropland). For \( ^{13}C \) labelling (photosynthetic labelling with \( ^{13}CO_2 \) to vegetation), \( ^{13}C \)-enriched (Subke et al., 2009; Biasi et al., 2012) or \( ^{13}C \)-depleted CO\(_2\) (Lin et al., 1999; Oggle and Pendall, 2015) was used. Analysis for \( ^{13}C \) may be conducted with isotopic ratio mass spectrometry (IRMS) or isotope ratio infrared spectroscopy (IRIS) (see Epron et al., 2012). The \( ^{13}C \) labelling presents itself as a powerful tool to examine C allocation from tree shoot to rhizosphere and soils using liquid scintillation counting (e.g., Sun et al., 2018b), autoradiography (e.g., Teramoto et al., 2016b) or accelerator mass spectrometry (AMS, e.g., Carbone and Trumbore, 2007). However, the use of \( ^{13}C \) labelling is strictly limited in field studies due to safety concerns (Epron et al., 2012). The ratio of \( ^{13}C \) (\( \Delta^{13}C \)), particularly bomb \( ^{13}C \) that is derived from nuclear bomb tests in the 1950s to 1960s, may be analysed using AMS in each \( R_t \) flux component (e.g., Carbone et al., 2008; Chiti et al., 2011). Using this methodology, Atarashi-Andoh et al. (2012) conducted a study in a cool-temperate deciduous broad-leaved forest in north-eastern Japan, demonstrating that the contribution of root derived C, litter and SOC decomposition to \( R_s \) was 31–39\%, 35–39\% and 23–33\%, respectively.

The most widely used method for component separation is the trenching method (Epron, 2010); we briefly focus on the trenching method and its influence on soil CO\(_2\) efflux. In the trenching method, roots around measurement plots are cut to a depth of approximately 30–100 cm (Epron, 2010). The appropriate depth of trenches is dependent on the distribution of the roots in the study site. For example, according to the global analysis in Jackson et al. (1996), on average, 75\% of roots were in the upper 40 cm in soil. However, they found that the root biomass distribution in the upper 30 cm soil largely varied from 52\% (temperate coniferous forest) to 93\% (tundra) in several ecosystems. Dead root litter input to the soil in trenched plots is inevitable at the beginning of trench treatment. The influence of dead root decomposition (overestimation of \( R_s \)) has been discussed for some time (Bowden et al., 1993; Epron et al., 1999; Ngao et al., 2007; Diaz-Pines et al., 2010; Savage et al., 2018). Bowden et al. (1993) suggested that dead root decomposition did not strongly influence soil CO\(_2\) efflux following nine months of trench treatment. On the contrary, Diaz-Pines et al. (2010) demonstrated that the decomposition rate of dead roots in the trenched plot was only 30\% following 15 months of trench treatment. They suggested that an appropriate correction was required to properly evaluate \( R_t \) and \( R_s \) separately using the trenching method. These contradicting results may potentially be attributed to the difference in the decomposability of dead root. For example, Ishikura et al. (2019) suggested the sustained influence of dead root decomposition in trenched plots even after one year of trench treatment because of the high groundwater level and relatively restricted dead root decomposition in a peat swamp forest in Sarawak, Malaysia. Root biomass analysis and root bag experiments were usually conducted in the same study site undergoing trench treatment and soil C flux measurements to correct for the influence of dead root decomposition. This was done to estimate the initial input of dead root C into the trenched plot and the rate of decomposition (Epron et al., 1999; Lee et al., 2003; Savage et al., 2018; Sun et al., 2020a). To minimise the influence of trenching on the soil C flux, it has been recommended that soil C flux measurements be conducted several months after trench treatment (Lee et al., 2003; Noh et al., 2016a). There is also an alternative approach to implement the trenching method; Sayer and Tanner (2010) estimated the contribution of root-rhizosphere respiration to \( R_s \) in a lowland tropical forest in Panama, by measuring soil CO\(_2\) efflux immediately before and after trenching. According to their report, the contribution of root-rhizosphere respiration to \( R_s \) estimated by comparing immediately before and one day after trenching was 38\%. This value was very similar to the estimation based on the comparison between control and trenched plots one year after trenching (39\%). This alternative may be effective if there was no significant change in the soil temperature and moisture (i.e., no rainfall) in the time immediately before and after trenching (Sayer and Tanner, 2010). In addition to the initial input of dead root litter into soil in trenched areas, the trenching method also affects C dynamics in the soil by cutting off C deposition from living roots (i.e., rhizodeposition). The rhizodeposition of living roots activates microbes in the soil and enhances \( R_t \) (Cheng, 2009; Dijkstra et al., 2013; Finzi et al., 2015; Savage et al., 2018). As such, the exclusion of the rhizosphere using the trench treatment also contributes to the long-term underestimation of \( R_s \) (Cheng, 2009; Finzi et al., 2015; Savage et al., 2018). Whilst the trenching method is a convenient and practical means to separate \( R_t \) and \( R_s \) in the field, it must be applied with caution to accommodate for the influence of disturbance. Savage et al. (2018) quantified the influence of artefacts accompanied with trenching such as dead root decomposition and changes in soil moisture and soil temperature, using a Bayesian modelling approach based on in-situ observation of soil CO\(_2\) efflux, in a mixed hardwood forest in Massachusetts. Such an approach may aid in minimising the influence of disturbance on soil CO\(_2\) efflux using the trenching method. In addition, comparing the estimated contribution rate of \( R_t \) to \( R_s \) between the trenching method and other separation methods at the same research site is likely to decrease uncertainty (e.g. Wang et al., 2008; Tomotsune et al., 2013).
3.2 Temperature and moisture responses of $R_s$

Fundamentally, $R_s$ is a biotic process affected by temperature (Kirschbaum, 1995). The $R_s$ is correlated either linearly or exponentially with temperature, or there is no relationship between them. The $R_s$ is usually estimated by Equation (1):

$$R_s = \alpha e^{\beta t}$$

where $R_s$ is $R_s$; $\alpha$ and $\beta$ are fitted parameters; $e$ is the natural logarithm; and $t$ is the soil temperature. The clear exponential relationship between $t$ and $R_s$ represented by Equation (1) has been commonly observed in humid Asian monsoon forest ecosystems (Zheng et al., 2009; Tan et al., 2013; Teramoto et al., 2016a; 2019; Wu et al., 2016; Noh et al., 2017). The exceptions to this are the tropical regions where seasonal temperature variation is relatively narrow (Hashimoto et al., 2004; Adachi et al., 2006). The temperature and moisture responses of soil C flux in tropical forests and peatlands are discussed in Section 4.4. The temperature sensitivity of $R_s$ is usually represented by $Q_{10}$ which is calculated using Equations (2) or (3):

$$Q_{10} = \frac{R_t(10^\circ C)}{R_t}$$

$$Q_{10} = e^{\beta t}$$

Typical $Q_{10}$ values of $R_s$ are roughly in the range of 2.0–3.0, based on previous observational studies that have largely taken place in relatively humid Asian regions covering sub-tropical forests, and temperate forests (Tan et al., 2013; Noh et al., 2017; Sun et al., 2017; Teramoto et al., 2017). In the meta-analysis presented in Wang et al. (2010), worldwide observations mainly from forests showed that the arithmetic mean of the $Q_{10}$ value was 2.67, despite a large variation in the $Q_{10}$ values from 0.98 to 6.27. These observed $Q_{10}$ values tend to be higher compared to global estimates of $Q_{10}$. For example, the global $Q_{10}$ in Bond-Lamberty and Thomson (2010) based on 1434 observations was 1.5. Recently, Hashimoto et al. (2015) synthesised a global $Q_{10}$ map using an empirical model; they found the averaged global $Q_{10}$ to be 1.40. Although the difference between the observed $Q_{10}$ (site-level) and the globally estimated $Q_{10}$ has been a crucial topic of concern, the underlying reasons for this difference has not yet been identified. Resolving this difference is key to better estimate for the feedback of regional and global scale soil C flux with climate change.

Soil moisture may affect many soil biological processes in soils; as such, $R_s$ is usually reduced when soil is either very dry or very wet (Harper et al., 2005). Under very dry conditions, the soil microbial activity may be limited by the diffusion rate of extracellular enzymes and soluble C substrates. Typically, there is a reduction in gas exchange and soil oxygen concentrations in very wet soils, limiting microbial and root activity (Luo and Zhou, 2006). The temperature, rate of photosynthesis, litterfall, precipitation, root and microbial activity can affect $R_s$. When seasonal variation of temperature is relatively small, soil moisture emerges as the main factor driving $R_s$ (Singh et al., 2011). In general, a concave down-shaped relationship exists between soil moisture and $R_s$ under constant temperature (e.g. Sha et al., 2005). However, in ecosystems experiencing large seasonal variations of soil temperature, the relationships between soil moisture and $R_s$ are not clearly observed at times because the strong influence of soil temperature on $R_s$ masks these relationships. Several studies have shown a relationship between soil moisture and $R_s$ by using temperature-normalised $R_s$ values for correlation analysis (Tan et al., 2013; Wang et al., 2014a; Teramoto et al., 2018). Temperature-normalised $R_s$ values are residual values calculated as the difference between observed $R_s$ and predicted $R_s$, estimated from the temperature response of $R_s$ (Tan et al., 2013; Teramoto et al., 2018), or the ratio of observed $R_s$ to predicted $R_s$ (Wang et al., 2014a). In an arid desert shrub ecosystem in China, Wang et al. (2014a) had confirmed a clear positive relationship between soil moisture and temperature-normalised $R_s$. The concave down-shaped relationship between soil moisture and $R_s$ was also observed in the relatively humid environment of a sub-tropical forest and a cool-temperate forest (Tan et al., 2013; Teramoto et al., 2018). Previous studies also observed the concave down-shaped relationship between soil moisture and $Q_{10}$ values of $R_s$ in the temperate forests of Asia (Wang et al., 2006; Teramoto et al., 2016a).

3.3 Methane flux

CH$_4$ has a global warming potential 28 times that of CO$_2$ (IPCC, 2013), its concentration in the atmosphere has increased by more than 100% since the beginning of the 19th century (Forster et al., 2007). CH$_4$ in soil is produced by methanogenesis under anaerobic conditions and consumed by methanotrophic microorganisms that use CH$_4$ and O$_2$ for metabolic activity under aerobic conditions. The dominant natural sources of CH$_4$ are natural wetlands, whilst those from anthropogenic sources include paddy fields and biomass burning (Saunois et al., 2016). Whilst forest soils normally serve as a CH$_4$ sink, during the rainy season or in anaerobic conditions they become a CH$_4$ source (Yan et al., 2008). According to the global model analysis by Tian et al. (2015), the estimated CH$_4$ emissions from global terrestrial ecosystems for 1981–2010 was 144.39 ± 12.90 Tg C yr$^{-1}$ with an annual increasing trend (0.43 ± 0.06 Tg C yr$^{-1}$) associated with a rapid increase in CH$_4$ emissions in natural wetlands and paddy fields because of an increase in the rice cultivation area and global warming. They found that the increase in CH$_4$ emissions was considerable in tropical regions and Asia (Tian et al., 2015). In Asia, China and India accounted for 30% to 50% of global CH$_4$ emissions (Saunois et al., 2016).

In contrast, upland soils are a major CH$_4$ sink (−16.13 to −17.76 Tg C yr$^{-1}$, Tian et al., 2015). Previous studies have indicated that soil water condition is one of the primary factors controlling the soil CH$_4$ uptake rate (Kim, 2013; Zhao et al., 2019b). This means that soil CH$_4$ uptake usually decreases with an increase in soil moisture. The influence of climate change on this CH$_4$ sink is key to the outcomes of the future CH$_4$ budget. Ni and Groffman (2018) reported on soil CH$_4$ flux from the late 1990s to the mid 2010s in temperate forests in north-eastern United States, finding that soil CH$_4$ uptake rate had significantly decreased over this period. They suggested that the cause for this decreased uptake rate was due to the increase in precipitation and hydrological flux (Ni and Groffman, 2018). There are very limited long-term datasets on soil CH$_4$ flux in the Asian region, although continuous measurements over
several years have already begun in some research sites (Sakabe et al., 2015; Ueyama et al., 2015; Ishikura et al., 2019). The increase of long-term data on soil CH\textsubscript{4} flux in the Asian region will contribute to a better understanding on changes in soil CH\textsubscript{4} flux under the influence of climate change and its response mechanisms in this region.

4. Soil C flux research in different ecosystems

4.1 Temperate grasslands and arid environment

Temperate grasslands in Asia are primarily distributed in Kazakhstan, Mongolia, and northeast China, whilst tropical grasslands are distributed in west India. \( R_{\text{g}} \) has been reported in no less than 70 study sites in natural and disturbed grasslands in Asia; most of these \( R_{\text{g}} \) measurements have been carried out in the Inner Mongolian Plateau, the Loess Plateau, and the Qinghai-Tibet Plateau in China (Table 3).

For this analysis, we used open data sources (Wang and Fang, 2009; Bond-Lamberty and Thomson, 2018; Feng et al., 2018; Steele and Jian, 2018; Mei et al., 2019; Yang et al., 2019a). The \( R_{\text{g}} \) in Asian temperate grasslands averaged 516 ± 436 g C m\textsuperscript{-2} yr\textsuperscript{-1}, based on observation records from the natural environment (i.e., without manipulation, for example, mowing, warming, precipitation control, and fertilising; Table 3). Although the mean \( R_{\text{g}} \) of tropical grasslands (620 ± 191 g C m\textsuperscript{-2} yr\textsuperscript{-1}), was higher than that of temperate grasslands, the CO\textsubscript{2} efflux of temperate grasslands was as high as 2407 g C m\textsuperscript{-2} yr\textsuperscript{-1} (Zhai et al., 2017). Wang and Fang (2009) reported a global \( R_{\text{g}} \) from temperate and tropical natural grasslands of 390 ± 46 and 601 ± 46 g C m\textsuperscript{-2} yr\textsuperscript{-1} (mean ± S.E.), respectively. The C emissions from tropical grasslands in Asia are comparable to those determined from a global study, although the emissions of temperate grasslands in Asia were 1.32 times higher; this was primarily from the relatively high \( R_{\text{g}} \) of the Loess Plateau (803 ± 528 g C m\textsuperscript{-2} yr\textsuperscript{-1}) and the Qinghai-Tibet Plateau (605 ± 321 g C m\textsuperscript{-2} yr\textsuperscript{-1}).

The \( R_{\text{g}} \) of deserts or sandy soil patches has seldom been investigated, with long-term observations lacking for these environments. The existing research findings (Bond-Lamberty and Thomson, 2018; Steele and Jian, 2018; Wang et al., 2019a) suggest low annual soil C emissions from Asian deserts of 158 ± 94 g C m\textsuperscript{-2} yr\textsuperscript{-1}. In arid and semi-arid environments, biological soil crusts significantly influence ecological processes, including soil C flux. The CO\textsubscript{2} flux of a sandy ground with a lichen and algae crust has been reported to be 307 and 168 g C m\textsuperscript{-2} yr\textsuperscript{-1}, respectively. This is higher than that of a mobile dune without a crust, the value of which is 115 g C m\textsuperscript{-2} yr\textsuperscript{-1} (Wang et al., 2019a). Aside from biotic respiration, abiotic processes are also considered to be important for CO\textsubscript{2} flux. A study in the Mu Us Desert in China found that natural desert soils abiotically absorb CO\textsubscript{2} at a mean rate of 0.08 g C m\textsuperscript{-2} d\textsuperscript{-1}, 73% of which is stored in the solid phase of the soil (Liu et al., 2015a).

There has been much less reporting on the CH\textsubscript{4} flux than CO\textsubscript{2} flux in arid and semi-arid environments. The few studies on this topic (Wang et al., 2014c; Mei et al., 2019; Yang et al., 2019a), have shown that grasslands consume CH\textsubscript{4} at a rate of 0.84 g C m\textsuperscript{-2} yr\textsuperscript{-1} (Table 3). A study on the Hobq Desert indicated a lower CH\textsubscript{4} uptake ability for deserts compared to grasslands, the former having an uptake rate of 0.15 g C m\textsuperscript{-2} yr\textsuperscript{-1} (Wang et al., 2019a).

4.2 Boreal forests

Boreal forest, also referred as taiga, is mainly distributed in the northern hemisphere stretching through Russia, Canada, Alaska and Scandinavia; in terms of area, it accounts for approximately one-third of global forests. The largest area of boreal forest in the world is located in Siberia (Soja et al., 2007). About 1700 Pg C is buried in soils of boreal forests (Deluca and Boisvenue, 2012); this is more than two times the amount of C storage in the atmosphere (IPCC, 2007). As one of the most vulnerable ecosystems, small changes in soil C storage may have a considerable impact on atmospheric C concentrations. However, there are limited studies on soil C flux of boreal forest, with most investigations carried out during the growing season.

Limited data (11 records) shows that the average annual \( R_{\text{g}} \) of Asian boreal forests is 275 ± 169 g C m\textsuperscript{-2} yr\textsuperscript{-1} (ranging from 27 to 576 g C m\textsuperscript{-2} yr\textsuperscript{-1}, Table 4). This is apparently lower than that of temperate and tropical forests and grasslands (see Sections 4.1 to 4.4). Furthermore, the average annual \( R_{\text{g}} \) is also lower than that of global natural boreal forests, of 483 ± 242 g C m\textsuperscript{-2} yr\textsuperscript{-1} based on 178 records (Table 4). When evaluating the regional differences, the \( R_{\text{g}} \) of boreal forests in Asian Russia was the lowest (217 g C m\textsuperscript{-2} yr\textsuperscript{-1}) compared to other locations (i.e., European Russia, Canada, USA, Finland, Sweden and Switzerland). These low C emissions are speculated to partly be caused by the distinct cold and dry weather of Siberia which is located in Asian Russia (Table 4).

Table 3. \( R_{\text{g}} \) and methane uptake in grasslands and arid environments.

| Gas       | Ecosystem/Region               | Mean (g C m\textsuperscript{-2} yr\textsuperscript{-1}) | S.D.\textsuperscript{*} (g C m\textsuperscript{-2} yr\textsuperscript{-1}) | MIN (g C m\textsuperscript{-2} yr\textsuperscript{-1}) | MAX (g C m\textsuperscript{-2} yr\textsuperscript{-1}) | Sample size |
|-----------|--------------------------------|-------------------------------------------------------|-------------------------------------------------|-------------------------------------------------|-------------------------------------------------|-------------|
| \( R_{\text{g}} \) | Temperate grassland            | 516                                                   | 436                                             | 2                                               | 2407                                            | 145         |
| | ♦ Inner Mongolian Plateau       | 217                                                   | 150                                             | 46                                              | 866                                             | 57          |
| | ♦ Loess Plateau                | 803                                                   | 528                                             | 171                                             | 2051                                            | 18          |
| | ♦ Qinghai-Tibet Plateau        | 605                                                   | 321                                             | 2                                               | 1161                                            | 42          |
| | Tropical grassland             | 620                                                   | 191                                             | 310                                             | 900                                             | 6           |
| | Desert                         | 158                                                   | 94                                              | 33                                              | 307                                             | 10          |
| CH\textsubscript{4} | Temperate grassland            | 0.84                                                  | 2.77                                            | -0.17                                           | 16.00                                           | 41          |
| | Desert                         | 0.15                                                   | 0.04                                            | 0.12                                            | 0.20                                            | 3           |

\( \textsuperscript{*} \) Standard deviation
Up to 80% of boreal forests in the world are located in continuous, discontinuous, sporadic, and isolated permafrost zones (Heilig et al., 2016). Limited precipitation accumulates on top of the permafrost layer to form large areas of wetland (swamp) forest where saturated soils provide conditions suitable for CH₄ production (Glagolev and Shnyrev, 2008). Unlike temperate and tropical forests that are a CH₄ sink, high CH₄ releases have been reported (up to 140 mg CH₄ m⁻² d⁻¹) from water-logged forests in Siberia (Shingubara et al., 2019); this is particularly the case in western Siberia which features 52.4 million ha of wetland (Glagolev et al., 2011; Terentieva et al., 2016). A study in eastern Siberia indicated that CH₄ emissions were highly dependent on the water table, whereby small changes in the water table may transform the soil from a CH₄ source to a CH₄ sink (van Huissteden et al., 2008). Compared to emissions, the CH₄ uptake in the upper dry soil were much lower; for example, −17 to −13 μg C m⁻² h⁻¹ for a forest in eastern Siberia (Morishita et al., 2003) and −280 to −92 μg C m⁻² h⁻¹ for a birch forest in western Siberia (Nakano et al., 2004). At times boreal forests have been found to switch between a weak CH₄ source and sink, for example, from −6.6 to 3.1 μg C m⁻² h⁻¹ for a forest in central Siberia (Morishita et al., 2014). Regionally, annual CH₄ oxidation by upland forest soils of west Siberia was estimated to be 0.36 ± 0.32 Tg CH₄ yr⁻¹; this is 0.12 times the emission (Glagolev et al., 2011).

Fire poses a significant threat to the boreal C pool; in addition to direct C loss from burning, fire also accelerates permafrost thawing, prompting greater amounts of C to be released as CH₄ (Dean et al., 2018; Ribeiro-Kumara et al., 2020). The impact of a single event of fire may last several decades (Köster et al., 2018). In addition to fire, global warming, deforestation and outbreaks of insects also threaten boreal forests (Shvidenko and Schepaschenko, 2013). Under the influence of natural and anthropogenic disturbances, boreal forests may turn from massive C pools into a climate bomb (Treat and Frolking, 2013). Close monitoring and modelling of the C cycle in boreal forests is one of the many essential requirements to prevent the formation of positive feedback loops and maintain its ability to mitigate global climate change.

### 4.3 Temperate forests

In Asia, temperate forests are mainly located in south Russia, north Mongolia, Korea, North Korea, Japan, and northeast and southeast China (Schulze et al., 2019). Of all ecosystem types, the greatest number of observations on C flux have been made in these forests. Measurements using the chamber method from 174 sites have produced approximately 468 annual R, records (Bond-Lamberty and Thomson, 2018; Steele and Jian, 2018; Yang et al., 2019a). Although there are a lower number of CH₄ flux observations compared to their CO₂ counterparts, at least 230 annual records may be obtained from the published literature (Morishita et al., 2007; Wang et al., 2014c; Ni and Groffman, 2018; Yang et al., 2019a).

As shown in Table 5, the average R, of temperate forests is 769 ± 338 g C m⁻² yr⁻¹, lower than the global average emission of 810 g C m⁻² yr⁻¹ (derived from the SRDB dataset, version 4.0, from 2974 records). By multiplying the temperate forest area (ca. 300 million ha), derived from the area of forest in eastern, western, and middle Asia (FAO, 2010), approximately 2.3 Pg C yr⁻¹ is released into the atmosphere, accounting for 2.6% of the total global soil CO₂ release (87.9 Pg C yr⁻¹, Warner et al., 2019).

Forest soils are largely CH₄ sinks; 219 of 230 collected records showed that the CH₄ uptake was between 0.0005 and 1.88 g C m⁻² yr⁻¹. As shown in Table 5, temperate forests in Asia absorb CH₄ from the atmosphere at a rate of 0.39 ± 0.38 g C m⁻² yr⁻¹. Multiplying the 300 million ha of temperate forests in Asia shows an annual CH₄ uptake of 1.17 Tg C yr⁻¹ (equivalent to 1.56 Tg CH₄ yr⁻¹), accounting for 18% of the global soil CH₄ uptake by forests (6.43 Tg C yr⁻¹, Tian et al., 2015).

An evaluation of the temporal variation suggests that the R, in temperate forests in Asia was higher in the 1960s and 1970s and decreased to the lowest rate in the 1990s (Table 5). This decline is inconsistent with findings from simulation studies. Based on global warming, the most commonly used climate-driven models have always reported increasing R, in the past several decades (Bond-Lamberty and Thomson, 2010; Zhao et al., 2017b). Therefore, the small sample size from the 1960s and 1970s (17 in total) cannot provide a reliable conclusion on whether climate-driven models have incorrectly reported R, for

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**Table 4.** R, of boreal forests. Flux and climate data were derived from SRDB v4.0 (Bond-Lamberty and Thompson, 2018), Kurganova et al. (2003); Morishita et al. (2010).

| Regions | Mean (g C m⁻² yr⁻¹) | S.D.* (g C m⁻² yr⁻¹) | MIN (g C m⁻² yr⁻¹) | MAX (g C m⁻² yr⁻¹) | Sample size | MAT¹ (°C) | MAP² (mm) |
|---------|------------------|-------------------|------------------|------------------|-------------|-----------|----------|
| Canada  | 524              | 268               | 85               | 1462             | 72          | 0.7       | 513      |
| USA     | 403              | 160               | 144              | 739              | 23          | −3.3      | 287      |
| Finland | 548              | 201               | 210              | 1091             | 50          | 2.4       | 655      |
| Sweden  | 367              | 233               | 98               | 1036             | 13          | 2.3       | 561      |
| Switzerland | 325       | 35                | 300              | 350              | 2           | 1.4       | 1050     |
| Russia-Europe | 460      | 324               | 120              | 1001             | 7           | −0.8      | 440      |
| Russia-Asian | 217      | 122               | 27               | 445              | 9           | −6.1      | 310      |
| China   | 535              | 59                | 493              | 576              | 2           | −4.6      | 490      |
| Asia    | 275              | 169               | 27               | 576              | 11          | −4.3      | 369      |
| Global  | 483              | 242               | 27               | 1462             | 178         | −0.8      | 464      |

*Standard deviation

¹Mean annual temperature

²Mean annual precipitation
that period. More effort is required to determine whether these high emissions are representative of all temperate forests in Asia during this period. For example, retrospective analysis on forest management practices or climate events over this period to identify potential causes for the high emissions. Additionally, historical simulation using process-based C cycle models or machine learning models, including a greater number of natural and anthropogenic drivers (not limited to climate factors), will help determine whether these high emissions may occur extensively under simulated scenarios over this period.

$R_{\text{c}}$ gradually increased post-1990s, whilst the declining CH$_4$ uptake became more distinct (Table 5). Since the 1990s, the temperature for Asia has increased by approximately 0.7 °C, at a rate of 0.23 °C every 10 years (CMA, 2019). This is one of the reasons that global warming is considered to be a cause for increased $R_{\text{c}}$ (Chen et al., 2018). Increasing precipitation is speculated to be the main factor for declining CH$_4$ uptake in forests located within 0–60 °N latitude (Ni and Groffman, 2018). Similar to other regions, there are additional factors that influence the C cycles of temperate forests. Studies have demonstrated that nitrogen (N) deposition decreases soil C loss by impeding organic matter decomposition (Jassal et al., 2011). Furthermore, temperate forests are consistently surrounded by the largest populations in the world due to the favourable climate. As a result, temperate forests have consistently experienced the impact of various human-related impacts to a greater extent, including deforestation, fragmentation, restoration, and the establishment of plantations. The present dataset reveals an increasing variation in soil C loss since the 1990s; understanding how the soil C flux of temperate regions responds to these disturbances remains a significant challenge. Although there has been considerable effort to investigate how these forces change the C cycle, inconsistent findings and limited knowledge of underpinning complex interactive effects have rendered unclear conclusions (Zhang et al., 2017). A detailed review on the influence of these disturbances on soil C fluxes, including global warming, is described in Section 5.

### 4.4 Tropical forests and peatlands

Tropical forests in Asia, particularly Southeast Asia, comprise approximately 15% of the total global tropical forests (FAO, 1995). In Asia, there are a relatively small number of soil C flux studies in tropical forests compared to temperate forests. According to the recent database by Bond-Lamberty and Thomson (2018; SRDB v4.0), the average $R_{\text{c}}$ of tropical forests in the Asian region (i.e., China, India, Indonesia, Malaysia and Thailand) was 1318 ± 538 g C m$^{-2}$ yr$^{-1}$. This value is an unmanipulated value (i.e., without warming, fertilising and control for precipitation and litter) representing the mean of 29 records of $R_{\text{c}}$ in tropical forests located between 23° 26′ S to 23° 26′ N. Previous studies on Asian tropical forests have examined the response of soil C fluxes to environmental factors; these are critical to better understand of soil C fluxes under global climate change. Therefore, the discussion in this paper will mainly focus on the factors controlling the spatio-temporal variation of $R_{\text{c}}$ in tropical forests and peatlands in Asia. In addition, brief mention will also be given to soil CH$_4$ flux in these ecosystems.

The $R_{\text{c}}$ in Asian tropical forests vary largely in each study. For example, Zhou et al. (2019b) reported the annual $R_{\text{c}}$ as 925 g C m$^{-2}$ yr$^{-1}$ (the average of two-year measurements from 2017 to 2018) in tropical rainforests in southwest China, whereas Hashimoto et al. (2004) estimated the annual $R_{\text{c}}$ as 2560 g C m$^{-2}$ yr$^{-1}$ in an evergreen forest in northern Thailand. Several factors are related to the spatial variation of $R_{\text{c}}$ in tropical forests. Katayama et al. (2009) showed a positive relationship between the mean tree diameter at breast height in measurement plots and $R_{\text{c}}$ in a lowland mixed-dipterocarp forest in Malaysia. Their results were indirectly supported by Saner et al. (2009), who found decreased $R_{\text{c}}$ in gaps in logged lowland dipterocarp forests in northern Borneo, Malaysia. In another study in Malaysia, Adachi et al. (2006) suggested that the fine root biomass, soil moisture, and soil C content were the main driving forces for the spatial variation of $R_{\text{c}}$ in the dipterocarp primary forest, dipterocarp secondary forest, and oil palm plantations, respectively. Topological variations were also a factor related to the spatial variation of $R_{\text{c}}$ in tropical forests. For example, Takahashi et al. (2011) indicated that the $R_{\text{c}}$ in the lower slope position on the watershed was larger compared with the upper slope position and the ridge in a tropical seasonal forest in Thailand. They discussed the difference of $R_{\text{c}}$ in relation to the soil moisture variation and

| Gas     | Ecosystems/Periods | Mean (g C m$^{-2}$ yr$^{-1}$) | S.D.* (g C m$^{-2}$ yr$^{-1}$) | MIN (g C m$^{-2}$ yr$^{-1}$) | MAX (g C m$^{-2}$ yr$^{-1}$) | Sample size |
|---------|--------------------|-------------------------------|-------------------------------|-------------------------------|-------------------------------|--------------|
| $R_{\text{c}}$ |
| Temperate forests | 769 | 338 | 55 | 3559 | 468 |
| 1960s | 1414 | 201 | 1272 | 1556 | 2 |
| 1970s | 969 | 218 | 494 | 1220 | 15 |
| 1980s | 691 | 448 | 161 | 1299 | 17 |
| 1990s | 669 | 174 | 234 | 943 | 35 |
| 2000s | 779 | 298 | 176 | 1968 | 296 |
| 2010s | 761 | 450 | 55 | 3559 | 96 |
| CH$_4$ |
| Temperate forests | 0.39 | 0.38 | -0.29 | 1.88 | 230 |
| 1990s | 0.48 | 0.35 | 0.10 | 1.35 | 37 |
| 2000s | 0.42 | 0.40 | -0.20 | 1.88 | 129 |
| 2010s | 0.25 | 0.30 | -0.29 | 1.41 | 53 |

*Standard deviation
various $R_s$ responses between slope positions. However, to date, there is a lack of evidence as to which factor primarily determines the spatial variation of $R_s$ in tropical forests.

Previous studies have also demonstrated a clear seasonal variation of $R_s$ in tropical forests despite the relatively stable soil temperature, the strongest controlling factor for $R_s$. Hanpattanakit et al. (2015) showed that the seasonal average of $R_s$ in the wet season (May to October) was higher than that of the dry season (November to April), based on four-year continuous measurements using the AOCC system in a dry dipterocarp forest in western Thailand. They suggested that the soil moisture and soil temperature strongly influence the seasonal variation of $R_s$, especially $R_s$. Hashimoto et al. (2004) showed positive relationships between soil moisture and $R_s$, indicating that soil moisture was the primary factor for the seasonal variation of $R_s$ in the forest. This positive relationship between soil moisture and $R_s$ has also been supported in the findings from other studies on the tropical forests in Asia (Kosugi et al., 2007; Takahashi et al., 2009). Based on the findings from these studies, it is clear that soil moisture is one of strongest factors controlling the seasonal variation of $R_s$ in tropical forests. Previous studies in tropical forests have also demonstrated exponential positive relationships between soil temperature and $R_s$ (Sha et al., 2005; Wangluk et al., 2013; Zhou et al., 2019b). However, in some studies, it was difficult to confirm a clear temperature response for $R_s$, due to very narrow seasonal variations in soil temperature (Hashimoto et al., 2004; Adachi et al., 2006).

Tropical peatlands in Southeast Asia cover an area of 247,778 km²; this area is largely located in Indonesia (206,950 km²) followed by Malaysia (25,889 km²). The $C$ pool of tropical peat soils in Southeast Asia was estimated at 68.5 Gt C (Page et al., 2011). Therefore, soil $C$ flux in tropical peatlands is one of the most important factors influencing the terrestrial $C$ cycle in the Asian region. For the past two decades, soil $C$ flux studies have been conducted in several tropical peatland ecosystems such as peat swamp forests (Furukawa et al., 2005; Jauhiainen et al., 2005; 2008; Ali et al., 2006; Hirano et al., 2009; Sundari et al., 2012; Comeau et al., 2013; Melling et al., 2013; Hergoualc’h et al., 2017; Ishikura et al., 2019), plantations (Watanabe et al., 2009; Jauhiainen et al., 2012; Comeau et al., 2013; Dariah et al., 2014; Husnain et al., 2014; Marwanto and Agus, 2014; Hergoualc’h et al., 2017; Wakhid et al., 2017; Matysek et al., 2018; Manning et al., 2019) and disturbed (drained, burned or logged) peatlands (Ali et al., 2006; Jauhiainen et al., 2008; Hirano et al., 2009; Comeau et al., 2013; Adjji et al., 2014; Ishikura et al., 2017; Itoh et al., 2017). Recently, Prananto et al. (2020) synthesised the results of 91 studies to estimate GHG emissions in tropical peatlands. Based on the database, they estimated annual CO₂ emissions as 1265, 1173, 1091 and 927 g C m⁻² yr⁻¹ in shrublands or croplands, forests, drained and burned peatlands, and plantations, respectively. The $R_s$ in tropical peatlands were found to be largely controlled by the groundwater level (e.g. Jauhiainen et al., 2012; Hirano et al., 2014; Ishikura et al., 2018). Hirano et al. (2014) conducted a study in a burned peatland in Central Kalimantan, and identified clear negative relationships between groundwater level and soil CO₂ efflux (peat decomposition). This meant that the drainage of peatlands accompanied with land-use change may cause significant amounts of CO₂ release from these drained peatlands. Prananto et al. (2020) estimated that for plantations, a decrease of groundwater level by 10 cm will result in an increase of $R_s$ by 138 g C m⁻² yr⁻¹.

As discussed in Section 3, forest soils include tropical forests functions as a CH₄ sink (Singh et al., 1997; Yashiro et al., 2008; Zhao et al., 2019b). In contrast, tropical peatlands are a CH₄ source. Furukawa et al. (2005) found a positive relationship between groundwater level and CH₄ emissions in several land-use types for peatlands in Indonesia. Ishikura et al. (2019) showed a bell-shaped relationship between soil CH₄ efflux and groundwater level based on using two years of continuous measurements in an undrained tropical peat swamp forest in Malaysia. Other studies in tropical peatlands have also suggested that the groundwater level was the primary factor controlling CH₄ emissions; this implies that wet soil conditions enhance CH₄ emissions whilst dry conditions decrease emissions (Inubushi et al., 2003; Hirano et al., 2009; Adjji et al., 2014). However, CH₄ emissions from tropical peatlands is relatively small when compared to other wetlands including boreal and temperate peatlands (Couwenberg et al., 2010).

As there continues to be limited availability of long-term continuous measurements for soil $C$ fluxes in tropical regions in Asia, there is uncertainty in the regional and global estimation of soil $C$ fluxes and the response mechanisms to global climate change. Further assessment of $C$ dynamics accompanied with ongoing anthropogenic and natural ecosystem disturbances especially in tropical peatlands requires the use of such long-term continuous measurements (Page et al., 2009; Hergoualc’h and Verchot, 2011). In addition, there is considerable uncertainty regarding the influence of climate change on soil $C$ flux in tropical forests as warming experiments in tropical forests are overwhelmingly limited in Asia and globally (Cavaleri et al., 2015; Kimball et al., 2018). Recently, Nottingham et al. (2020) reported the results of a two-year warming experiment for a tropical forest in Panama, showing that soil CO₂ emissions from tropical forest soils had increased 55% due to a 4 °C soil warming. Increase of long-term continuous measurements and soil warming experiments in Asian tropical forests would contribute to decrease large uncertainty of soil $C$ flux.

4.5 Paddy fields

Asia, specifically South, Southeast and East Asia, is a region with the largest area of paddy fields in the world, accounting for approximately 87% of the harvest area worldwide (FAO, 2019). During the long period of flooding for cultivation, a favourable anaerobic environment is established CH₄ production (Serrano-Silva et al., 2014); as such, paddy fields are one of the most important atmospheric CH₄ contributors (IPCC, 1996; Saunois et al., 2020). There are many studies on CH₄ flux across the Asian rice cultivation zone focusing on spatio-temporal variations (Mingxing and Jing, 2002), the role of environmental and management influences, including that of temperature (Wassmann et al., 2000b), soil pH (Yan et al., 2005), water management (Yagi, 1996), the use of fertilisers (Naser et al., 2007), planting methods (Oo et al., 2018), and global warming
Based on 942 observations on CH₄ emissions, the average CH₄ emissions for the entire cultivation season vary largely from 0.10 to 103.42 mg m⁻² h⁻¹ (Table 6). CH₄ emissions also vary among countries from 4.39 ± 0.21 to 18.22 ± 12.48 mg m⁻² h⁻¹ (Table 6). Water regime, soil pH and organic matter amendments are the main factors driving these differences in CH₄ flux (Wang et al., 2018). The Food and Agriculture Organisation (FAO) reported on long-term CH₄ emissions from Asian rice paddy fields from 1961 to 2017 (FAO, 2019); emissions averaged 19.8 Tg CH₄ yr⁻¹ (ranging from 16.2 to 22.4 Tg CH₄ yr⁻¹). This accounted for 2.7–3.7% (535–737 Tg CH₄ yr⁻¹) of global total CH₄ emissions from natural and anthropogenic sources (IPCC, 1995; Fujita et al., 2020; Saunois et al., 2020).

This long-term estimate showed an apparent rise in CH₄ emissions due to the steady increase in the rice cultivation area from 106.96 million ha in 1961 to 145.54 million ha in 2017 (the emission rate was set as constant throughout those years in the FAO estimation). Field observations have suggested an approximate 3% increase in terms of CH₄ flux since the 1990s (Table 6). Global warming and enhanced fertilisation are implicated in accelerating the release of CH₄ from the soil (Yuan et al., 2014; Yvon-Durocher et al., 2014). This suggests that actual CH₄ emissions from Asian paddy field may be higher than the FAO report, as their estimation was derived from data collected prior to the 1990s. Using a top-down method, Fujita (2020) reported that the global CH₄ emissions from rice cultivation was 33 Tg CH₄ yr⁻¹, 1.5 times higher than the FAO reported value of 22 Tg CH₄ yr⁻¹ (FAO, 2019).

The population of Asia in 2019 was 4.6 billion, this accounts for 60% of the global population. Furthermore, a population increase is almost certain in the next few decades (UN, 2019). The severe stress induced from a burgeoning population will place greater demands on food production in the near future; this makes the positive correlation between rice production and CH₄ emissions (Epule et al., 2011). This issue is two-fold: the proportional increase in the cultivation area associated with meeting greater demand, and the use of substrate enrichment processes such as fertiliser application and the subsequent efficient CH₄ transport by aerenchyma systems from the soil into the atmosphere (Butterbach-Bahl et al., 1997). Many studies have proposed a means to decrease CH₄ emissions through various cultivation management practices (Zou et al., 2005; Shang et al., 2011; Hussain et al., 2015). For instance, the mitigation potential through improved water management was up to 73% (Hussain et al., 2015). However, when considering population and food security issues, the real confronting challenge is the means by which to decrease CH₄ emissions from paddy field whilst increasing rice production. Although some studies have explored this issue (Kim et al., 2017), further research is still required to establish applicable practice guidelines.

5. Climate change and disturbance

5.1 Global warming

Global warming is capable of potentially stimulating the soil C loss due to the general temperature dependency of respiration and the magnitude of soil C stocks in terrestrial ecosystems (Carey et al., 2016; Melillo et al., 2017). However, despite the robust research on the response of Rₑ to changing temperatures, there is inconsistency regarding the impacts of Rₑ responses from global warming on the global C balance; this introduces great uncertainty in predicting scenarios for future climate change and related feedback (Bond-Lamberty and Thomson, 2010). Main cause for uncertainty regarding the response of Rₑ to warming is based on the complex soil processes regulating Rₑ, associated with Rₑ and Rₛ (Schindlbacher et al., 2009; Kuzyakov and Gavrichkova, 2010), and the high spatial heterogeneity across biomes in terrestrial ecosystems (Wang et al., 2014b; Li et al., 2019a). There is also insufficient information regarding whether, when, and to what extent the partitioned respiratory components acclimate to warming (Melillo et al., 2017; Bradford et al., 2019), and contradictory results on the acclimation of Rₑ to elevated temperature (Wang et al., 2014b). These factors also contribute to the uncertainty in predicting the response of Rₑ to ongoing climate change.

| Regions/Periods | Mean | S.D.* | MIN | MAX | Sample size |
|-----------------|------|-------|-----|------|-------------|
|                 | (mg CH₄ m⁻² h⁻¹) | (mg CH₄ m⁻³ h⁻¹) | (mg CH₄ m⁻² h⁻¹) | (mg CH₄ m⁻³ h⁻¹) |             |
| Asia            | 9.61 | 10.89 | 0.10 | 103.42 | 942         |
| Bangladesh      | 4.39 | 0.21  | 4.05 | 4.60  | 5           |
| Cambodia        | 10.48| 2.34  | 8.30 | 15.06 | 8           |
| China           | 9.94 | 9.91  | 0.14 | 56.20 | 388         |
| India           | 6.15 | 14.86 | 0.39 | 103.42 | 142         |
| Indonesia       | 12.75| 9.96  | 0.79 | 39.00 | 105         |
| Japan           | 6.42 | 6.15  | 0.20 | 29.11 | 104         |
| Philippines     | 6.95 | 7.84  | 0.10 | 39.67 | 86          |
| South Korea     | 18.22| 12.48 | 2.89 | 61.52 | 70          |
| Thailand        | 11.11| 10.27 | 1.00 | 39.02 | 20          |
| Vietnam         | 7.91 | 5.35  | 1.68 | 19.10 | 14          |
| 1990s           | 9.81 | 11.80 | 0.10 | 103.42 | 550         |
| 2000s           | 9.64 | 9.79  | 0.17 | 61.52 | 181         |
| 2010s           | 10.09| 10.02 | 0.25 | 52.95 | 142         |

*Standard deviation
warming. Several biogeochemical factors combine to influence the timing (short-term and long-term), magnitude, and thermal acclimation of soil C loss; 1) depletion of microbial accessible C pools (Eliasson et al., 2005; Knorr et al., 2005), 2) reductions in microbial biomass (Wang et al., 2014b), 3) a shift in microbial C use efficiency (Tucker et al., 2013; Li et al., 2019a), 4) changes in microbial community composition (Romero-Olivares et al., 2017), and 5) the thermal acclimation of R, (Jarvi and Burton, 2018).

Over the last decade, with there are a number of studies on climate change manipulation and artificial warming experiments that have been intensively conducted in various ecosystems of Asian regions such as grasslands in the Loess and Tibetan Plateau in China, croplands in central and southeast China, temperate deserts in northwest China, cool-temperate forests in northern Japan, warm-temperate forests in central and southern Japan and central China, and sub-tropical forests in southern China (Aguillos et al., 2013; Liu et al., 2015b; 2016c; Noh et al., 2016a; 2017; Teramoto et al., 2016a; 2018; Wu et al., 2016; Fang et al., 2017; 2018; Yue et al., 2018; Hu et al., 2019; Sun et al., 2019; Wang et al., 2019c; 2020a; Yuan et al., 2019). These studies reported on the differential responses of R, and R, to warming, such as decreasing trends in the warming effect over time (Noh et al., 2016a; 2017; Fang et al., 2018). Other warming studies resulted in sustained stimulations of the warming effects on R, due to sufficient soil moisture conditions or the negligible depletion of C substrate (Aguillos et al., 2013; Teramoto et al., 2016a; 2018; Wu et al., 2016). In the Asian monsoon region, the sustained stimulatory warming effects on R, in the long-term have outcomes that contrast the decline in the initial stimulatory warming effect after several years of warming treatment in other ecosystems (Melillo et al., 2002; Eliasson et al., 2005). A five-year warming study by Teramoto et al. (2018) had demonstrated that inter-annual variation in the magnitude of the annual warming effect was positively related to the number of rainy days associated with precipitation. Despite the accumulation of this evidence from the field, there is still a critical need to conduct multi-factorial manipulation experiments associated with natural or anthropogenic disturbances (Zhu et al., 2015; Li et al., 2017; Sun et al., 2018a; 2019; Zhou et al., 2019b), and long-term experiments including extreme climate events to better integrate experiments and models to contribute toward global synthesis (Rustad, 2008).

Global warming also generally enhances CH uptake in grassland, tundra, and forest ecosystems due to the stimulated activity of methane-oxidising methanotrophs with increasing soil temperature (Luan et al., 2019; Yue et al., 2019). As the influence of increasing temperature on CH uptake is dependent on the soil water content/level, N deposition and their interaction (Yue et al., 2019; Li et al., 2020), the impacts of warming on CH uptake may be inconsistent across different ecosystems. Previous studies have reported that experimental warmings enhanced CH uptake in arid alpine grasslands in the Tibetan Plateau and in a tundra ecosystem at Changbai Mountain, China (Lin et al., 2015; Zhou et al., 2016; Chen et al., 2017; Li et al., 2020). In contrast, they had no significant effect on CH emissions from alpine meadows on the Qinghai-Tibetan Plateau, and a temperate desert in China (Zhao et al., 2017a; Yue et al., 2019; Li et al., 2020; Wu et al., 2020). This highlights the need for future research to consider the interactive impacts of N deposition, precipitation and warming manipulations on CH uptake (Zhu et al., 2015; Chen et al., 2017; Yue et al., 2019; Wu et al., 2020).

5.2 Other natural disturbances

Natural disturbances such as droughts, typhoons, forest fires, and insect infestations have profound effects on biogeochemical processes in terrestrial ecosystems (van der Molen et al., 2011; Edenburg et al., 2012; Reichstein et al., 2013). Recent climate models have predicted that drought will increase in duration and intensity, whilst the frequency of extreme precipitation events will also increase globally (Borken and Matzner, 2009; Dai, 2011). As the increased drying/wetting alterations and changes to the moisture regime affect soil biogeochemical cycles, there is growing interest in quantifying and predicting CO fluxes in terrestrial ecosystems (van der Molen et al., 2011; Edenburg et al., 2012; Reichstein et al., 2013; Khokhar and Park, 2019; Wu et al., 2020). The monsoon climatic interaction with geographical complexity in the Asian region has resulted in serious droughts or extreme precipitation events (Xu et al., 2015; Ceglar et al., 2017). The impact of drought and precipitation on R, has been investigated in several forest ecosystems in China (Deng et al., 2018; Huang et al., 2018; Liu et al., 2019). Prolonged severe drought conditions had significantly decreased R, and R, due to suppressed microbial activity and fine root metabolic capacity for a temperate forest in the Jigong Mountains National Nature Reserve, northeast China, and a sub-tropical forest in the Wuyi National Nature Reserve, southeast China (Huang et al., 2018; Liu et al., 2019). These findings were consistent with those of most previous studies (Schindlbacher et al., 2012; Deng et al., 2017). However, in a tropical forest located in the Dinghushan Biosphere Reserve, southern China, an increase in the frequency of precipitation had stimulated R, due to enhanced leached dissolved organic C production (Deng et al., 2018). In contrast, in a temperate forest located in the Changbai Mountains National Reserve, northeast China, reduced precipitation significantly increased R, due to changes in the bacterial community (Yan et al., 2019a). These contradictory findings may be attributed to differences in soil texture, moisture regime, and microbial community composition among the different ecosystems. In tropical peatlands, CH flux is generally negative or negligible at low groundwater levels. Although this is low compared with emissions from boreal temperate peatlands, precipitation regulated changes in groundwater level may lead to greater variability (Couwenberg et al., 2010).

Typhoons are also considered to be an important natural disturbance affecting the Asian monsoon regions (Hong and Kim, 2011). Typhoons occur annually accompanied by gale force winds and heavy precipitation leading to severe damage, especially in forest ecosystems. The damage includes canopy defoliation and wind-throw-related tree mortality, which influences the C budget and R, of the ecosystem (Ito et al., 2005; Lee et al., 2006; Reichstein et al., 2013). Diurnal, seasonal, and inter-annual variations in soil and understory CO fluxes have been investigated to understand the response of soil and understory CO fluxes to changes in environmental factors from typhoons in a sub-tropical mangrove wetland (Jin et al., 2013) and a temperate larch forest (Teramoto et al., 2017), respectively. However,
the impact of typhoons on $R_i$ is to be determined by complex interactions between abiotic and biotic factors such as increases in tree mortality-induced litter inputs and soil moisture, and changes in soil temperature, light intensity, and biomass of living roots.

Globally, approximately half the average annual flux of 2.0 Pg C yr$^{-1}$ has been emitted due to fires in forest ecosystems between 1997 and 2009; 20% of these fires were from deforestation and tropical forest fires, 16% from woodland fires, and 15% from mostly extra-tropical forest fires (2001–2009) (van der Werf et al., 2010). Although most fires in ecosystems are set by humans, a complex interaction between climate, ecosystem processes, and human activities determines fire-driven C emissions from burned areas (van der Werf et al., 2010). In tropical peatlands of Southeast Asia such as the Indonesian lowlands and Peninsular Malaysia, there are substantial fire-driven C losses in the global terrestrial C cycle (Herawati and Santoso, 2011; Smith et al., 2018; Kumar et al., 2020). For examples, the $R_i$ from burned peat surfaces increased 46% over the first nine months following a fire in the Kalimantan tropical peatland forest (Astiani et al., 2018), and a ground fire significantly stimulated $R_i$ by 17.5% for a mixed forest of sub-tropical-temperate transition zones in central China (Hu et al., 2020). Although the humid climate typical of the Asian monsoon region is relatively free from wildfire when compared to drier ecosystems in other arid or boreal ecosystems, the frequency, intensity, and size of wildfires have been increasing and is expected to increase rapidly in future given the increase in lightning strikes and seasonally extreme droughts associated with global warming (Hu et al., 2017; Chen et al., 2019; Song et al., 2019). Sun et al. (2014) investigated the effect of fire disturbance on $R_i$ in birch and larch forests in China, and Hu et al. (2017) examined the effects of fire severity on $R_i$ in larch forests to better estimate $R_i$ in boreal ecosystems; they found that wildfires decreased $R_i$ mainly due to decreased autotrophic respiration. Experimental burning studies in the temperate forests of northern Japan and a Tibetan meadow grassland in China, have demonstrated that the artificial fire significantly decreases $R_i$ due to reductions in root activity, litter layer, microbial biomass and soil moisture (Kim, 2013; Chen et al., 2019). Song et al. (2019) found that wildfires in a montane coniferous forest decreased $R_i$ largely from the reduction in $R_i$ leading to an increase in the $R_{soil}$ to $R_i$ ratio. Wildfires may have positive and negative effects on $R_{soil}$ as fire-induced heat suppresses microbes, whilst post-fire ash supplies a greater volume of substrate for microbes and reduces rhizospheric respiration by inhibiting photosynthesis and the development of roots and mycorrhizae (Irving et al., 2007; Kuzyakov and Gavrichkova, 2010). Based on the large spatial heterogeneity of vegetation cover, fire regime, and increasing deforestation/degradation fires, there is a need for a greater number of studies on the impact of fire on $R_i$, and its components across diverse ecosystems, to obtain a better understanding on future C cycles (Smith et al., 2018; Hu et al., 2020). At present, there are very few studies on the fire impacts on CH$_4$ uptake in the Asian region to reliably assess post-fire impacts on CH$_4$ flux compared to northern peatlands (Davidson et al., 2019; Kumar et al., 2020). Kim (2013) reported that in a white birch forest located at the Teshio Experimental Forest, northern Japan, a low-intensity experimental surface fire did not affect CH$_4$ uptake, finding no significant alterations in soil properties such as soil moisture, pH, and inorganic N concentrations. However, severe fires increases soil CH$_4$ uptake through an increase in gas diffusivity induced by the removal of the organic layer, thereby enhancing CH$_4$ oxidation (Burke et al., 1997; Jaatinen et al., 2004). A recent study in Southeast Asia by Lupascu et al. (2020) found a post-fire increase in CH$_4$ eflux for a tropical peat swamp forest in Brunei Darussalam due to a prolonged higher groundwater level and more optimal methanogenesis conditions in burned areas compared to intact areas. As drainage and clearing on peatlands over recent years has resulted in an unprecedented increase in severe peat fires in Indonesia, Singapore, Malaysia, and Brunei (Kumar et al., 2020), further research is required to investigate the impacts of fire on CH$_4$ flux in these regions.

Due to recent changing climates in North America, insect and disease outbreaks such as pine beetles and emerald ash borer have been more intense than for any previous periods (Edenburg et al., 2012; Hicke et al., 2012; Fratteggi et al., 2018; Kolka et al., 2018; Van Grinsven et al., 2018). In many Asian countries including Japan, China, Taiwan, Hong Kong, and Korea (Kwon et al., 2011; Mabuhay and Nakagoshi, 2012), pine ecosystems have been radically threatened by pine wilt disease (PWD) caused by the pine wood nematode (Bursaphelenchus xylophilus (Steiner et Buhrer) Nickle). Despite the widespread PWD infestation in the Asian region, there are few available studies attempting to understand the impacts of insects and pathogens on soil CO$_2$ and CH$_4$ flux. Jeong et al. (2013) reported the effect of PWD on $R_i$ and C storage in Pinus densiflora stands in Korea. This study indicates that severe PWD infestation had significantly decreased $R_i$, due to the reduction in living root biomass and $R_i$. Interestingly, the slightly damaged stands led to more favourable environmental conditions, such as higher soil temperature, increased litterfall inputs, and an increase in the photosynthetic rate for the remaining trees, with partial canopy removal for microbial and root growth activity compared with other damaged stands (Jeong et al., 2013). Ecosystem modelling does not often adequately represent complex changes in stand structure following insect infestations. As such, a greater number of field studies on these dynamic processes associated with pre and post-disturbance stand structure are required to better predict effects of disturbance on bio-geophysical and biogeochemical cycling, including $R_i$, its source components, and CH$_4$ flux (Borkhuan et al., 2015).

5.3 Ecosystem management and land-use change

Forest management practices such as thinning (Son et al., 2004; Tian et al., 2009; Pang et al., 2013; Teramoto et al., 2019; Wang et al., 2019b; Zhao et al., 2019a), clear-cutting (Kim, 2008; Bai et al., 2020), and N addition (Liu et al., 2016a; Wang et al., 2019d; Yan et al., 2019a; Wang et al., 2020b) may affect C sequestration, storage, and flux over time (Peng et al., 2008; Bai et al., 2020). Thinning, defined as the partial removal of trees from a plantation, is the most common silvicultural practice for sustainable forest management. Thinning changes the micro-climatic conditions and soil properties, such as soil temperature and moisture, organic matter and nutrient content, pH, conductivity, and microbial activity, ultimately impacting on $R_i$ (Tian et al., 2009; Teramoto et al., 2019). $R_i$ has been found to decrease in Japanese larch and Chinese fir stands subjected to high intensity thinning by
over 33% due to the root death-induced decrease in \( R_h \) in Korea and China (Son et al., 2004; Tian et al., 2009). In contrast, in Pinus tabulaeformis stands subjected to low-intensity thinning of 15–20%, resulted in an increase in \( R_h \) due to enhanced \( R_h \) from the increased productivity and root growth. \( R_h \) was stimulated by increased soil temperature in thinned stands located in the temperate and sub-tropical forests of China (Cheng et al., 2014; 2015; Lei et al., 2018). Such contradictory results may be attributed to a combination of the varying responses of \( R_h \) and \( R_s \) to the occurrence of thinning and the thinning intensity (Fang et al., 2013; Cheng et al., 2015; Zhao et al., 2019a). Clear-cutting also affects \( R_s \) by changing the micro-climatic conditions influencing \( R_s \) and decreasing \( R_h \). The existing inconsistent results from different studies may be attributed to variations in the time point at which measurements were taken following clear-cutting (Kim, 2008; Bai et al., 2020). Thus, long-term studies to partition \( R_s \) into \( R_h \) and \( R_s \) are required to accurately understand the influence of thinning intensity and clear-cutting on the components of \( R_s \) and C cycling in forest ecosystems.

Additionally, in Asian paddy fields that represent an exceptionally large source of \( CH_4 \) management practices such as fallow incorporation and mulching of rice straw, may reduce \( CH_4 \) emissions (Wassmann et al., 2000c). Proper irrigation through effective water management may also reduce \( CH_4 \) emissions due to low \( CH_4 \) production and high \( CH_4 \) oxidation. This needs to be implemented alongside considerations to the global warming potential of \( N_2O \) emissions and rice yields (Wassmann et al., 2000a; Wang et al., 2012; Tirol-Padre et al., 2018).

External N deposition, originating mainly from fossil-fuel burning and artificial fertiliser applications, have been widely reported to enhance plant and soil microbial respiration by directly altering soil N availability and indirectly altering physiochemical properties (Zhou et al., 2014; Peng et al., 2017; Yan et al., 2019a). In many areas in the Asian region, N deposition is expected to continue to increase (Zhao et al., 2009). Artificial N addition was found to increase \( R_h \) in a temperate forest (Yan et al., 2019a) and meadow grasslands (Zhang et al., 2014; Zhu et al., 2015; Fang et al., 2017; Wang et al., 2020b). Moreover, a 12-year long-term N enrichment treatment decreased \( R_s \), and was also observed to produce soil acidification in grasslands (Chen et al., 2016). Fang et al. (2018) found that N addition had no significant effect on \( R_h \) in a semi-arid grassland ecosystem in which N addition significantly increased \( R_s \) and inhibited \( R_h \). There are very complex responses by the soil respiratory components as a result of N addition, varying substantially with fertilising time/seasons and periods. The soil texture and N supply-demand conditions also regulate root and microbial composition and activity (Zhou et al., 2014). In terms of \( CH_4 \) flux, N deposition affects methanotrophs and methanogen activity (Yue et al., 2016). In alpine grassland ecosystems in China, N addition increased \( CH_4 \) uptake (Li et al., 2012; Zhu et al., 2015; Chen et al., 2017), and decreased \( CH_4 \) uptake with increased precipitation (Wu et al., 2020), although no significant effects were observed (Yue et al., 2016; Zhao et al., 2017a). In a northern temperate forest ecosystem in Japan, experimental N deposition inhibited \( CH_4 \) uptake from well-drained soils in response to increased N inputs (Kim, 2013). At a N-limited temperate desert in China, Yue et al. (2019) showed that low N addition enhanced \( CH_4 \) uptake, whilst N addition did not affect \( CH_4 \) uptake in another study by Zhou et al. (2019c). These conflicting results may be a result of variations in soil moisture, N level, and soil water-filled pore space (Fang et al., 2014; Huang et al., 2015; Chen et al., 2017). Recently, Xia et al. (2020) conducted a meta-analysis on the effects of N addition on soil \( CH_4 \) uptake across global forest biomes including 17 sites in the Asian region. The study confirmed that N deposition decreased forest soil \( CH_4 \) uptake. It also highlighted that the effects of N addition on \( CH_4 \) uptake were biome-specific and dose dependent. In Asian paddy fields, many studies have reported that N addition through the application of fertiliser affects \( CH_4 \) emissions, although the resulting \( CH_4 \) emissions were water level and dose dependent (e.g., Wang et al., 2012; 2018; Zhang et al., 2016). They also correlated with N-induced crop yield, most likely due to the increase in C substrates for methanotrophs (Banger et al., 2012).

Grazing, as a result of grassland management practices, is a major factor influencing \( R_s \) in grassland/pasture ecosystems (Wang et al., 2019a; 2020b). A recent global meta-analysis on grassland ecosystems by Zhou et al. (2019a) found that grazing decreases soil moisture, N availability, and soil microbial diversity; these activities inhibit plant growth and soil microbial decomposition, decreasing \( R_s \). There are a few studies on the effects of grazing on soil \( CH_4 \) uptake. Studies on grassland ecosystems located in Haihe and Inner Mongolia, China, have indicated that grazing reduced \( CH_4 \) uptake due to a decrease in air permeability (Liu et al., 2007; Chen et al., 2011), whilst grazing did not significantly affect \( CH_4 \) flux with no changes to the bacterial community (Lin et al., 2015; Zhu et al., 2015; Mei et al., 2019). In the Asian region, few studies have explored how grassland management may be able to regulate the combined and interactive effects of grazing and other global change factors on \( R_s \) components and \( CH_4 \) flux (Zhu et al., 2015; Wang et al., 2020b).

The absence of research on the impact of land-use changes such as deforestation or reforestation on \( R_s \) components and \( CH_4 \) emissions had created greater uncertainty in predicting future soil C sequestration in Asian regions (Hergoualc'h and Verchot, 2011; Jauhiainen et al., 2012; Sun et al., 2018a; Wu and Mu, 2019; Zhang et al., 2020). Further research are still needed in vulnerable ecosystems such as peatlands (Coveneberg et al., 2010; Jauhiainen et al., 2012; Sundari et al., 2012; Ishikura et al., 2019), sub-tropical wetlands (Philipp et al., 2017), and deserts (Wang et al., 2014a; Fa et al., 2018) to better estimate future C cycles. Recently McDaniels et al. (2019) conducted a global meta-analysis on the effect of land-use change on soil \( CH_4 \) fluxes using 62 studies. They found that that land-use changes in wetter ecosystems resulted in greater \( CH_4 \) flux increases. Wu and Mu (2019) investigated the \( CH_4 \) and CO\(_2\) emissions from the conversion of a mature forest to a Korean pine plantation, demonstrating that secondary forests increased \( CH_4 \) uptake and \( R_s \) compared with mature forests and plantations in northeast China. A recent study using isotope fractionation by Lang et al. (2020) reported land-use transformations from forest to lubber plantations in southwest China decreased \( CH_4 \) uptake, confirming a weakened \( CH_4 \) sink capacity for soils involved in the rubber monoculture. In tropical peatlands in Southeast Asia, representing an immense reservoir of C in Asia, drainage is a pre-requisite as part of the large scale degradation for agriculture, Acacia and oil
palm developments (Kaupper et al., 2019). Wong et al. (2020) reported that the conversion of a peat swamp forest to a drained oil palm plantation decreased CH$_4$ emissions from tropical peatlands in Malaysia. Yang et al. (2019b) also concluded that conversion from grassland to cropland (e.g., paddy field) increased CH$_4$ uptake in an agro-pastoral ecotone of Inner Mongolia, China. However, the lowered water table level in wet ecosystems may trigger the CH$_4$ oxidation process (McDaniel et al., 2019). As such, there is an urgent need to quantify C losses, changes of soil C fluxes under anaerobic and aerobic conditions associated with changes in water level following natural and anthropogenic disturbance (Couwenberg et al., 2010; Hergoualc'h and Verchot, 2011).

6. Model synthesis for soil C flux

6.1 Contributions of observations in the Asian region to global $R_s$ database
In the last decade, there have been extensive observations of $R_s$ in the Asian region. As shown in Fig. 1, there were 1247 records of annual $R_s$ in Asia, contributing to 29% of the global soil respiration database (SRDB v4.0; Bond-Lamberty and Thompson, 2018). The mapped geographical distribution shows that the locations of these observations were particularly dense in China and Japan, while measurements were sparse to the west and north of the Asian region (Fig. 1). The average annual $R_s$ rate in the Asian region was 817 g C m$^{-2}$ yr$^{-1}$, ranging from 1.7 to 4140 g C m$^{-2}$ yr$^{-1}$. Based on a data-oriented modelling study (Hashimoto et al., 2015), using the SRDB and a semi-empirical model, the $R_s$ emitted from the Asian region was 22 Pg C yr$^{-1}$; this constitutes approximately 24% of the global $R_s$. The spatial distribution of annual $R_s$ shows that $R_s$ was high in the Southeast Asian islands, India and the Malay Peninsula. $R_s$ was low in the north Asian region and dry areas to the west of China (Fig. 2). These spatial variations in the estimates from the model were a result of the climatic diversity in this region.

6.2 Importance of non-CO$_2$ GHG fluxes
The non-CO$_2$ GHG (e.g., CH$_4$ and N$_2$O) fluxes are also important in terms of the overall GHG budget in the Asian region. Ito et al. (2018) estimated the spatio-temporal distributions of N$_2$O flux in East Asia, determining that the high annual total N$_2$O emission was 2.03 Tg N$_2$O yr$^{-1}$; this was mainly due to the intensive use of fertiliser (Ito et al., 2018). The study suggested

Fig. 1. Geographical distribution of all $R_s$ records in the database (SRDB v4.0; Bond-Lamberty and Thomson, 2018) (left), and histogram of annual $R_s$ records around the world, and the contributions of data records from the Asian region (right). Blue rectangle (left) denotes the Asian region in Section 6.

Fig. 2. Modelled global $R_s$ (after Hashimoto et al., 2015) at a spatial resolution of 0.5°. The blue rectangle denotes the Asian region in Section 6.
that the soil N$_2$O emissions were likely to offset 40–74% of the mitigation from CO$_2$ sequestration by terrestrial ecosystems in this region. The importance of CH$_4$ is also similar to N$_2$O in the calculating the GHG budget in this region (Akiyama et al., 2005; Katayanagi et al., 2017). Ito et al. (2019) had evaluated the CH$_4$ budget of East Asia using inventory data and a process-based model; the estimated net CH$_4$ emissions from East Asia was $67.3 \pm 14.0$ Tg CH$_4$ yr$^{-1}$ for 2000–2012, equivalent to $525$ Tg CO$_2$-C yr$^{-1}$ and accounting for 13% of the total global CH$_4$ emissions. In the budget, soil oxidation and wetland emissions were $-2.35$ and $9.43$ Tg CH$_4$ yr$^{-1}$, respectively, and the emission from agriculture was $15.84$ Tg CH$_4$ yr$^{-1}$. This highlights the importance of soil CH$_4$ fluxes in the GHG budget in this region. Modelling CH$_4$ emissions from paddy fields is an important issue in East Asia (Akiyama et al., 2005; Katayanagi et al., 2017). These Asian-scale evaluations of CH$_4$ and N$_2$O fluxes underline the important contributions from non-CO$_2$ GHG fluxes of the Asian region to the global GHG budget.

6.3 Future perspectives

For accurate evaluation of the GHG budget, understanding key processes and incorporating these processes in modelling is essential. For example, explicit modelling of microbes and/or new conceptual categorisation of soil functional types is likely to improve models (Wieder et al., 2013; Bond-Lamberty et al., 2016). One key approach is the use of machine learning, such as Random Forests and Support Vector Machine (Ichii et al., 2017); these rely on a large amount of observational data, providing data-oriented estimates. In terms of $R_w$, Warner et al. (2019) applied a Random Forest machine learning algorithm and estimated global $R_w$ at a 1 km resolution. They used 2657 observations as input for the global annual $R_w$ based on the global $R_w$ database. There has recently been an increase in the spatial resolution associated with global scale studies, whereby location-identified observation data have been intensively used in machine learning approaches via the combination with other variables from other databases (e.g., climate). To promote such global, high-resolution estimates of soil GHG fluxes (i.e., CO$_2$, CH$_4$, and N$_2$O), there is a need for a greater volume of observational data in various locations, timing and ecosystems. As for timing, the Asian region can be characterised by the monsoon climate. Although it is not easy to identify the required number of observations, the lack of data and biased location issues are more serious for CH$_4$ and N$_2$O than CO$_2$. Whilst it may be unrealistic to establish sufficient observation networks for the end, there is a need to obtain new data, particularly in tropical, arid, and boreal regions, together with an effort to improve upscaling techniques are necessary (Schimel et al., 2015; Bond-Lamberty, 2018).

7. Conclusions

By the end of 2019, there were more than 100 flux sites in AsiaFlux (AsiaFlux, 2020), spanning forest, shrub, savanna, grassland, farmland, wetland, steppe, tundra, desert, lake, and urban ecosystems. In the last two decades, soil C flux studies in Asia has improved methods such as continuous measurement with automated chamber systems and compact GHG analysers. The number of measurements for soil C flux has drastically increased in the Asian region (as an example, see Table 5). These works had contributed to a better understanding of the mechanisms related to soil C dynamics, factors controlling the spatio-temporal variation of soil C fluxes (water, temperature, soil nutrients, C content), and improved model development, regional and global scale estimation, evaluation of the influence of disturbance and climate change on soil C fluxes, including suggestions for policy-makers. However, future research needs to focus on three major aspects: (1) The reduction of spatial bias in terms of the location of observations: although the number of soil C flux observations has increased in Asia, spatial biases still exist in the distribution of these observations. There are more observation data in temperate forests than in arid, boreal and tropical regions despite the broader area covered by those regions. Reducing these spatial bias observations would greatly improve regional and global soil C flux estimations; (2) An increase in the number of long-term continuous datasets and climate manipulation experiments: these are critical to understand ecosystem progress and dynamics, especially mechanisms controlling soil C fluxes under disturbance (natural and artificial) and climate change. Therefore, long-term climate manipulation experiments such as soil warming experiments are also needed to precisely estimate the feedback of soil C flux to climate change (Aguilos et al., 2013; Teramoto et al., 2016a; Wu et al., 2016; Noh et al., 2017; Huang et al., 2018; Wu et al., 2020); and (3) A greater number of synthetic studies integrating influential geographical, biological factors and climate change: although responses from soil C flux to individual regulatory drivers have been extensively investigated, such integrative synthetic studies are still very limited. As such, there is a need for a multi-factorial approach and related modelling. These includes process-based models, empirical models, and machine learning to disentangle the complex responses of soil C flux to disturbances and climate change scenarios and formulate strategies to mitigate and adapt to climate change.

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