

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 (CO_2) next to photosynthesis in terrestrial ecosystems. With the absorption of atmospheric methane (CH_4), upland soils become a large CO_2 source and CH_4 sink. These soil carbon (C) fluxes are key factors in the mitigation and adaptation 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 preventive 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 CO_2 (heterotrophic respiration, R_h). Soil respiration (R_s), the sum of R_h and plant root respiration (R_r), is the largest CO_2 flux next to photosynthesis in terrestrial ecosystems (Raich and Schlesinger, 1992). However, upland soils are a sink for atmospheric methane (CH_4) and an important component recognised in the global CH_4 budget (Ni and Groffman, 2018). As such, soils (and particularly forest soils) are a major CO_2 source and 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 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

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CO₂/CH₄ 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 (Misson *et al.*, 2007) and soil CO₂ gradient methods using CO₂ probes (Liang *et al.*, 2004; Pinging *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_s increases the CO₂ 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_s 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 CO₂. This absorbed CO₂ may be measured by the increased weight of the alkali medium (soda lime) or by titration with an acid to calculate R_s (Lundegårdh, 1927; Edwards, 1982). Kirita (1971) improved CO₂ 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_s in the Asian region (Nakane *et al.*, 1984; Tulaphitak *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_s 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_s and underestimate high R_s . Two possible factors were identified in association with this bias (Yim *et al.*, 2002). The first factor was the limited CO₂ absorption efficiency of the alkali medium and the resulting CO₂ 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 CO₂ 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_s under high CO₂ condition should be relatively mild with this protocol of Kirita (1971) compared to other alkali absorption methods because of its increased CO₂ absorptive efficiency. Following the publication of the research on the bias of alkali absorption method for R_s 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_s , 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 CO₂ 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 (N₂O) 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 CO₂ 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) (Liang *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

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 the field	<ul style="list-style-type: none"> ● Low cost ● No need for electrical power source in field 	<ul style="list-style-type: none"> ● Overestimation (at low R_s) or underestimation (at high R_s) for R_s ● Need for chemical analysis in laboratory
	Closed	Closed chamber method	From 1980s	<ul style="list-style-type: none"> ● Possible to concurrently measure several types of GHG fluxes (CO_2, CH_4 and N_2O) with GC ● High versatility in many ecosystems (e.g. forests, paddy fields, grasslands) due to the simplicity 	<ul style="list-style-type: none"> ● Need for GC or analyser (cost) ● Need for manual gas sampling several times in one measurement and gas analysis in laboratory*
	Closed	Closed dynamic chamber method	From 1990s	<ul style="list-style-type: none"> ● Possible to concurrently measure several GHG fluxes (depending on the analyser) and record data in the field ● Relatively simpler flow line compared with open dynamic chamber systems 	<ul style="list-style-type: none"> ● Need for analyser (cost) and power source (usually battery) in field measurements ● Pressure difference between inside and outside of chamber may produce large errors
Dynamic	Open	Open flow method	From late 1960s	<ul style="list-style-type: none"> ● Measurement in steady-state ● Possible to concurrently measure GHG fluxes (depending on the analyser) and record the data in the field 	<ul style="list-style-type: none"> ● Typically requires a buffer tank ● Unbalance between inflow and outflow of air in chamber may cause pressure difference and produce large errors for flux values
	Open	Open top chamber method	From late 1990s	<ul style="list-style-type: none"> ● Measurement in steady-state with minimum pressure difference between inside and outside of chamber ● No need for buffer tank 	<ul style="list-style-type: none"> ● Strong wind may produce errors for flux values ● Due to the potential for these errors, this method is rarely used in recent studies for soil C flux measurement
	Open/ Closed	AOCC method	From 1970s	<ul style="list-style-type: none"> ● Suitable for long-term continuous measurements for several GHG fluxes (depending on the analyser) because of minimal difference in environmental factors between inside and outside of chamber ● Efficient data collection with high resolution 	<ul style="list-style-type: none"> ● Expensive system installation ● Need for stable electrical power source (usually commercial power supply) for measurement

* 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_s 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; 2019; 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 CH₄ flux measurements using chamber methods

Soil CH₄ flux may be also measured using the chamber methods. Until recently, soil CH₄ 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 CH₄ analyser has enabled continuous field measurements for soil CH₄ 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 CH₄ flux using the CH₄ 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 CH₄, Yang *et al.* (2019a) developed equations using 154 site-years monthly CH₄ flux data measured using the chamber methods in China. Such an approach will contribute to the improved estimation of the regional and global soil CH₄ flux. It is also expected that the observation network in the Asian region for soil CO₂/CH₄ flux measurements will continue to expand, facilitating a better understanding of soil C dynamics.

2.6 CH₄ 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 CH₄ emissions. Rice cultivation (through paddy fields), is one of the major CH₄ sources in the agricultural

industry (Yan *et al.*, 2009; Zhang *et al.*, 2016), and more than 90% of CH₄ emissions from paddy fields has been suggested to originate from Asian monsoon regions (Yan *et al.*, 2009). The chamber method has also been applied to measure CH₄ flux in paddy fields. The most popular method to measure CH₄ 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 CH₄ and N₂O 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 CH₄/CO₂ analyser (G2201-i, Picarro Inc., Santa Clara, CA, USA) to obtain two-day continuous data in tropical paddy field located in Thailand. They evaluated CH₄ emissions by bubble ebullition and diffusion separately, and indicated the controlling factors for CH₄ ebullition (i.e., atmospheric air pressure and soil surface temperature). Continuous and high frequency measurements of CH₄ 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_s components and separation

The process of R_s is defined as the release of CO₂ from the soil to the atmosphere. R_s originates from different sources; microbial decomposition of SOC (R_h) and respiration by living roots (R_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_r to total R_s 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_r to R_s 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 phenology of vegetation. In addition, ecosystem and vegetation types may also contribute to the variation of the R_r contribution rate. Subke *et al.* (2006) suggested that contribution rate of R_r to R_s was relatively high in temperate and tropical forests compared with boreal coniferous forests. The factors, R_r and R_h , responded differently to temperature increase, resulting in different Q_{10} values (where Q_{10} is the multiple of the R_s increase ratio due to a 10 °C temperature rise, representing the temperature sensitivity of R_s) (Rey *et al.*, 2002). This potentially alters the net soil C flux, potential C sequestration, and climate change associated feedbacks. The

increase in R_r may reflect increased C inputs to the soil through photosynthesis (Högberg *et al.*, 2001), while increased R_h 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_h (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_s 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_s components. However, the complete *in-situ* separation of R_r (usually root and rhizosphere) from R_h is quite difficult. Many approaches focused on the separation of R_s 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, CO₂ effluxes of each flux component such as R_r , litter and SOC decomposition are measured separately in a laboratory to estimate contribution of R_r and R_h to R_s (Hanson *et al.*, 2000; Baggs, 2006). For example, Noh *et al.* (2011) incubated intact roots of *Pinus densiflora*, soil (including roots) and soil alone in root chambers in a laboratory. They estimated that R_r contributed 53–58% of R_s by measuring the CO₂ efflux in each component. Although this method is simple and low-cost, the *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_r to R_s based on the linear regression between the root biomass (x-axis) and R_s (y-axis), where the intercept of the y-axis is be estimated as R_h (Kucera 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_s . 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_r and R_h to R_s by comparing soil CO₂ 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, it will be discussed in further detail in the later part of this Section 3.1. In gap analysis, the R_r to R_s contribution is estimated by measuring soil CO₂ 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_s , 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_r contribution of 49–57% to R_s 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_s components.

Category	Method	Advantages	Disadvantages
Component integration	Component integration	Simplicity and low cost.	Difficulty in evaluating each flux component <i>in-situ</i> . Difficulty in evaluating the root-microbial interaction and substrate supply to microbial community in field conditions.
Root regression	Root regression	Simplicity and applicability to many study sites without disturbing the soil and rhizosphere structure prior to measuring R_s .	Uncertainty may be large when coefficient of regression is small.
	Root removal	Exclusion of R_r without any dead roots in measurement plot.	Disturbance of soil structure. Potential to increase soil moisture.
	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_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 ¹⁴ C analysis is expensive. Use of ¹⁴ C in the field is strictly limited due to safety concerns.
	¹³ C and ¹⁴ C labelling		
	Bomb ¹⁴ C analysis		

(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_r without disturbing the soil and rhizosphere; however, there is uncertainty as to the extent to which recently fixed C from trees contributes to R_r (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_r and R_h 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_r to R_s is calculated based on the difference of $\delta^{13}\text{C}$ (‰) in each flux component (R_r , litter and SOC decomposition) (Rochette *et al.*, 1999; Albanito *et al.*, 2012; Ogle 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_s in cropland (maize, the C4 plant, was planted on the cropland). For ^{13}C labelling (photosynthetic labelling with $^{13}\text{CO}_2$ to vegetation), ^{13}C -enriched (Subke *et al.*, 2009; Biasi *et al.*, 2012) or ^{13}C -depleted CO_2 (Lin *et al.*, 1999; Ogle and Pendall, 2015) was used. Analysis for $\delta^{13}\text{C}$ may be conducted with isotope ratio mass spectrometry (IRMS) or isotope ratio infrared spectroscopy (IRIS) (see Epron *et al.*, 2012). The ^{14}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 ^{14}C labelling is strictly limited in field studies due to safety concerns (Epron *et al.*, 2012). The ratio of ^{14}C ($\Delta^{14}\text{C}$), particularly bomb ^{14}C that is derived from nuclear bomb tests in the 1950s to 1960s, may be analysed using AMS in each R_s 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_h) 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_r and R_h 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 trench treatment 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_h (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_h (Cheng, 2009; Finzi *et al.*, 2015; Savage *et al.*, 2018). Whilst the trenching method is a convenient and practical means to separate R_r and R_h 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_r 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):

$$R = \alpha e^{\beta t} \quad (1)$$

where R is R_s ; α and β 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):

$$Q_{10} = R_{(t+10)} / R_t \quad (2)$$

$$Q_{10} = e^{10\beta} \quad (3)$$

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 (Saunio *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 $144.39 \pm 12.90 \text{ Tg C yr}^{-1}$ with an annual increasing trend ($0.43 \pm 0.06 \text{ Tg C yr}^{-1}$) 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 (Saunio *et al.*, 2016).

In contrast, upland soils are a major CH_4 sink (-16.13 to $-17.76 \text{ Tg C yr}^{-1}$, 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 CH₄ flux in the Asian region will contribute to a better understanding on changes in soil CH₄ flux under the influence of climate change and its response mechanisms in this region.

4. Soil C flux research in different ecosystems

4.1 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_s has been reported in no less than 70 study sites in natural and disturbed grasslands in Asia; most of these R_s 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_s in Asian temperate grasslands averaged $516 \pm 436 \text{ g C m}^{-2} \text{ yr}^{-1}$, 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_s of tropical grasslands ($620 \pm 191 \text{ g C m}^{-2} \text{ yr}^{-1}$), was higher than that of temperate grasslands, the CO₂ efflux of temperate grasslands was as high as $2407 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Zhai *et al.*, 2017). Wang and Fang (2009) reported a global R_s from temperate and tropical natural grasslands of 390 ± 46 and $601 \pm 46 \text{ g C m}^{-2} \text{ yr}^{-1}$ (mean \pm 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_s of the Loess Plateau ($803 \pm 528 \text{ g C m}^{-2} \text{ yr}^{-1}$) and the Qinghai-Tibet Plateau ($605 \pm 321 \text{ g C m}^{-2} \text{ yr}^{-1}$).

The R_s 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 $158 \pm 94 \text{ g C m}^{-2} \text{ yr}^{-1}$. In arid and semi-arid environments, biological soil crusts significantly influence ecological processes, including soil C flux. The CO₂ flux of a sandy ground with a lichen and

algae crust has been reported to be 307 and $168 \text{ g C m}^{-2} \text{ yr}^{-1}$, respectively. This is higher than that of a mobile dune without a crust, the value of which is $115 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Wang *et al.*, 2019a). Aside from biotic respiration, abiotic processes are also considered to be important for CO₂ flux. A study in the Mu Us Desert in China found that natural desert soils abiotically absorb CO₂ at a mean rate of $0.08 \text{ g C m}^{-2} \text{ d}^{-1}$, 73% of which is stored in the solid phase of the soil (Liu *et al.*, 2015a).

There has been much less reporting on the CH₄ flux than CO₂ 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 CH₄ at a rate of $0.84 \text{ g C m}^{-2} \text{ yr}^{-1}$ (Table 3). A study on the Hobq Desert indicated a lower CH₄ uptake ability for deserts compared to grasslands, the former having an uptake rate of $0.15 \text{ g C m}^{-2} \text{ yr}^{-1}$ (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_s of Asian boreal forests is $275 \pm 169 \text{ g C m}^{-2} \text{ yr}^{-1}$ (ranging from 27 to $576 \text{ g C m}^{-2} \text{ yr}^{-1}$, 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_s is also lower than that of global natural boreal forests, of $483 \pm 242 \text{ g C m}^{-2} \text{ yr}^{-1}$ based on 178 records (Table 4). When evaluating the regional differences, the R_s of boreal forests in Asian Russia was the lowest ($217 \text{ g C m}^{-2} \text{ yr}^{-1}$) 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_s and methane uptake in grasslands and arid environments.

Gas	Ecosystem/Region	Mean ($\text{g C m}^{-2} \text{ yr}^{-1}$)	S.D.* ($\text{g C m}^{-2} \text{ yr}^{-1}$)	MIN ($\text{g C m}^{-2} \text{ yr}^{-1}$)	MAX ($\text{g C m}^{-2} \text{ yr}^{-1}$)	Sample size
R_s	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
CH ₄	Desert	158	94	33	307	10
	Temperate grassland	0.84	2.77	-0.17	16.00	41
	Desert	0.15	0.04	0.12	0.20	3

*Standard deviation

Up to 80% of boreal forests in the world are located in continuous, discontinuous, sporadic, and isolated permafrost zones (Helbig *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 threaten 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_s 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_s 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_s 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_s 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_s for

Table 4. R_s 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	MAT ¹ (°C)	MAP ² (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

¹ Mean annual temperature

² Mean 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_s 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_s (Chen *et al.*, 2018). Increasing precipitation is speculated to be the main factor for declining CH_4 uptake in forests located within $0\text{--}60^\circ\text{N}$ 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_s of tropical forests in the Asian region (i.e., China, India, Indonesia, Malaysia and Thailand) was $1318 \pm 538 \text{ g C m}^{-2} \text{ yr}^{-1}$. 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_s in tropical forests located between $23^\circ 26'\text{S}$ to $23^\circ 26'\text{N}$. 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_s 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_s in Asian tropical forests vary largely in each study. For example, Zhou *et al.* (2019b) reported the annual R_s as $925 \text{ g C m}^{-2} \text{ yr}^{-1}$ (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_s as $2560 \text{ g C m}^{-2} \text{ yr}^{-1}$ in an evergreen forest in northern Thailand. Several factors are related to the spatial variation of R_s in tropical forests. Katayama *et al.* (2009) showed a positive relationship between the mean tree diameter at breast height in measurement plots and R_s in a lowland mixed-dipterocarp forest in Malaysia. Their results were indirectly supported by Saner *et al.* (2009), who found decreased R_s 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_s 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_s in tropical forests. For example, Takahashi *et al.* (2011) indicated that the R_s 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_s in relation to the soil moisture variation and

Table 5. R_s and CH_4 uptake of temperate forests. We removed studies in which the exact sampling year from the calculations for each decade were difficult to obtain.

Gas	Ecosystems/Periods	Mean ($\text{g C m}^{-2} \text{ yr}^{-1}$)	S.D.* ($\text{g C m}^{-2} \text{ yr}^{-1}$)	MIN ($\text{g C m}^{-2} \text{ yr}^{-1}$)	MAX ($\text{g C m}^{-2} \text{ yr}^{-1}$)	Sample size
R_s	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_h . 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 the 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; Adji *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; Adji *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; Saunio *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

(Tokida *et al.*, 2010).

Based on 942 observations on CH₄ emissions, the average CH₄ emissions for the entire cultivation season varies 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; Saunio *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 R_s 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 R_s to changing temperatures, there is inconsistency regarding the impacts of R_s 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 R_s to warming is based on the complex soil processes regulating R_s associated with R_r and R_h (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 acclimates to warming (Melillo *et al.*, 2017; Bradford *et al.*, 2019), and contradictory results on the acclimation of R_s to elevated temperature (Wang *et al.*, 2014b). These factors also contribute to the uncertainty in predicting the response of R_s to ongoing

Table 6. CH₄ emissions from rice paddy fields. We removed studies in which the exact sampling year from calculations for each decade that were difficult to obtain. Flux data were derived from Wang *et al.* (2018).

Regions/Periods	Mean (mg CH ₄ m ⁻² h ⁻¹)	S.D.* (mg CH ₄ m ⁻² h ⁻¹)	MIN (mg CH ₄ m ⁻² h ⁻¹)	MAX (mg CH ₄ m ⁻² h ⁻¹)	Sample size
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_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 (Aguilos *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_r and R_h 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_h due to sufficient soil moisture conditions or the negligible depletion of C substrate (Aguilos *et al.*, 2013; Teramoto *et al.*, 2016a; 2018; Wu *et al.*, 2016). In the Asian monsoon region, the sustained stimulatory warming effects on R_h 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_4 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_4 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_4 uptake may be inconsistent across different ecosystems. Previous studies have reported that experimental warmings enhanced CH_4 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_4 emissions from alpine meadows on the Qunghai-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_4 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; Edburg *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_2 and CH_4 flux response to extreme drought and precipitation (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_s 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_s and R_h due to suppressed microbial activity and fine root metabolic capacity for a temperate forest in the Jigong Mountains National Nature Reserve, central 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_s due to enhanced leached dissolved organic C production (Deng *et al.*, 2018). In contrast, in a temperate forest located in the Changbai Mountains Natural Reserve, northeast China, reduced precipitation significantly increased R_s 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_4 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_s 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_2 fluxes have been investigated to understand the response of soil and understory CO_2 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_s 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_s 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_s 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_s in birch and larch forests in China, and Hu *et al.* (2017) examined the effects of fire severity on R_s in larch forests to better estimate R_s in boreal ecosystems; they found that wildfires decreased R_s 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_s 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_s largely from the reduction in R_r leading to an increase in the R_h to R_s ratio. Wildfires may have positive and negative effects on R_s , 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 (Irvine *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_s 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 CH₄ uptake in the Asian region to reliably assess post-fire impacts on CH₄ 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 CH₄ uptake, finding no significant alterations in soil properties such

as soil moisture, pH, and inorganic N concentrations. However, severe fires increases soil CH₄ uptake through an increase in gas diffusivity induced by the removal of the organic layer, thereby enhancing CH₄ 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 CH₄ efflux 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 CH₄ 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 (Edburg *et al.*, 2012; Hicke *et al.*, 2012; Fraterrigo *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 al.* Buhner) 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 CO₂ and CH₄ flux. Jeong *et al.* (2013) reported the effect of PWD on R_s and C storage in *Pinus densiflora* stands in Korea. This study indicates that severe PWD infestation had significantly decreased R_s due to the reduction in living root biomass and R_r . 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_s , its source components, and CH₄ flux (Borkhuu *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_s (Tian *et al.*, 2009; Teramoto *et al.*, 2019). R_s 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_r 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_s due to enhanced R_r 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_r and R_h to the occurrence of thinning and the thinning intensity (Pang *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_h and decreasing R_r . 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_r and R_h 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_s 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_{ss} , 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_s in a semi-arid grassland ecosystem in which N addition significantly increased R_r 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.*, 2019d; 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 Haibei 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; Jauhainen *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 (Couwenberg *et al.*, 2010; Jauhainen *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 McDaniel *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 rubber 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 CH₄ emissions from tropical peatlands in Malaysia. Yang *et al.* (2019b) also concluded that conversion from grassland to cropland (e.g., paddy field) increased CH₄ uptake in an agro-pastoral ecotone of Inner Mongolia, China. However, the lowered water table level in wet ecosystems may trigger the CH₄ 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 Thomson, 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-CO₂ GHG fluxes

The non-CO₂ GHG (e.g., CH₄ and N₂O) 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 N₂O flux in East Asia, determining that the high annual total N₂O emission was 2.03 Tg N₂O yr⁻¹; this was mainly due to the intensive use of fertiliser (Ito *et al.*, 2018). The study suggested

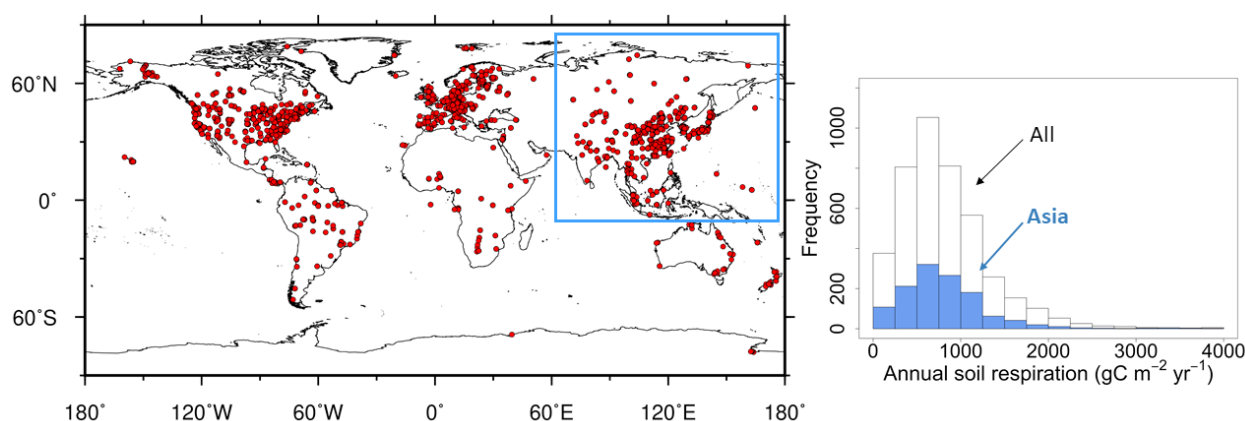


Fig. 1. Geographical distribution of all R_s records in the database (SRDB v4.0, Bond-Lamberty and Thomson 2018) (left), and histogram of annual R_s records around the world, and the contributions of data records from the Asian region (right). Blue rectangle (left) denotes the Asian region in Section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

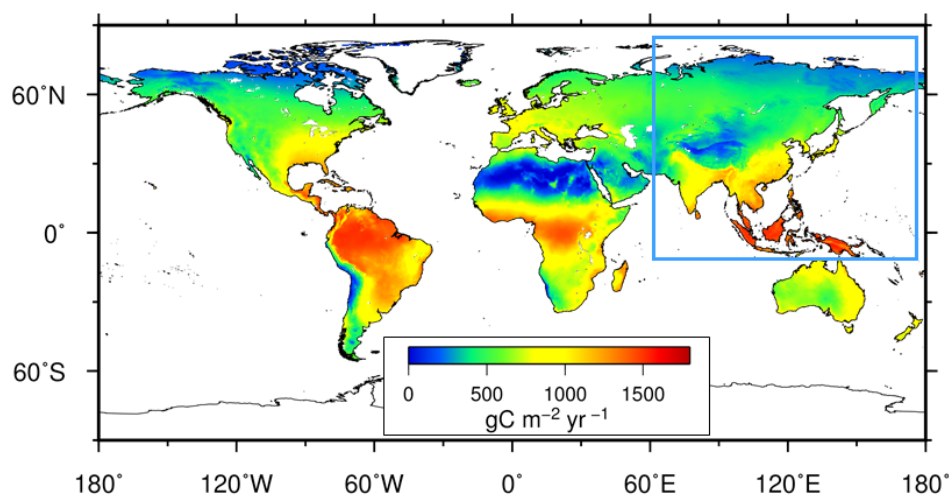


Fig. 2. Modelled global R_s (after Hashimoto *et al.* 2015) at a spatial resolution of 0.5°. The blue rectangle denotes the Asian region in Section 6.

that the soil N₂O emissions were likely to offset 40–74% of the mitigation from CO₂ sequestration by terrestrial ecosystems in this region. The importance of CH₄ is also similar to N₂O in the calculating the GHG budget in this region (Akiyama *et al.*, 2005; Katayanagi *et al.*, 2017). Ito *et al.* (2019) had evaluated the CH₄ budget of East Asia using inventory data and a process-based model; the estimated net CH₄ emissions from East Asia was 67.3 ± 14.0 Tg CH₄ yr⁻¹ for 2000–2012, equivalent to 525 Tg CO₂-C yr⁻¹ and accounting for 13% of the total global CH₄ emissions. In the budget, soil oxidation and wetland emissions were -2.35 and 9.43 Tg CH₄ yr⁻¹, respectively, and the emission from agriculture was 15.84 Tg CH₄ yr⁻¹. This highlights the importance of soil CH₄ fluxes in the GHG budget in this region. Modelling CH₄ emissions from paddy fields is an important issue in East Asia (Akiyama *et al.*, 2005; Katayanagi *et al.*, 2017). These Asian-scale evaluations of CH₄ and N₂O fluxes underline the important contributions from non-CO₂ 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_s , Warner *et al.* (2019) applied a Random Forest machine learning algorithm and estimated global R_s at a 1 km resolution. They used 2657 observations as input for the global annual R_s based on the global R_s 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., CO₂, CH₄, and N₂O), 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 CH₄ and N₂O, than CO₂. 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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