



CHAPTER 2
REVIEW OF
LITERATURE



2.1. Soil Properties and Ecosystem Carbon Flux

An ecosystem's net carbon (C) balance is highly dependent on input and output C fluxes. The major processes that regulate the C cycling of an ecosystem are CO₂ uptake by photosynthesis, loss by soil respiration and CaCO₃ dissolution [1, 2]. The intricate interaction of factors such as vegetation, climate and soil properties govern the alteration in ecosystem CO₂ flux. Being a major sink of carbon, soil properties strongly regulate the ecosystem C fluxes. Soil parameters, particularly temperature and moisture, are shown to be the most important regulators of variation in soil CO₂ efflux [3, 4]. Previous studies have documented significant effects of soil organic carbon (SOC), dissolved organic carbon (DOC), soil pH, total nitrogen and salinity on ecosystem C fluxes [5, 6, 7, 8, 9].

2.1.1. Soil Physico-chemical Characteristics

Soil physicochemical parameters such as pH, SOC, soil N, soil P, and available nutrients are said to influence soil C mineralization [10]. An increase in C mineralization has been reported with the increase in these soil properties [11, 12].

2.1.1.1. Soil Texture

Soil texture and soil macropore structure play a key role in regulating the amount of CO₂ efflux [13]. Soil texture is an important soil characteristic that can influence the C cycle in forests, including soil organic matter retention and tree growth response. Fine-textured soils generally store higher amounts of C than coarser soils because of higher specific surface area and more reactive surfaces [14]. Furthermore, the effect of texture on aeration and water holding capacity influences soil microbial activity, which is also related to C stabilisation and storage. Under similar temperature and moisture conditions, respiration in fine-textured soils can be higher due to higher retention of water and substrate availability for microbes [15]. Soil texture also influences net primary production (NPP) by controlling availability of water to plants based on precipitation and the final balance of air and water in soil [16,17].

Ebrahimi *et al.* [18] used artificial neural network (ANN) and linear regression models to estimate soil respiration under different land uses in humid subtropical and semi-arid areas of Iran. The results of the ANN model revealed that soil texture, pH, EC, percent of calcium carbonate equivalent, and organic C explained 66% of the variation in basal respiration. According to Li *et al.* [19], soil texture regulates C flux by controlling soil moisture, which has an indirect impact on heterotrophic activity. Augustin and Cihacek

[20] documented strong correlation between silt content and soil organic C. This in turn, affect plant productivity and influences soil C sequestration. According to Yang *et al.* [21], soil texture was a significant contributor for formation of CO₂ concentration gradients and gas diffusion. They documented that in loamy soil, the CO₂ efflux is primarily controlled by CO₂ production. Whereas, in silty loam soil both CO₂ production and gas diffusion and in silty clay loam soil the gas diffusion is influenced.

2.1.1.2. Soil Moisture

Soil moisture is a potential capping factor for land C uptake. It can reduce gross primary production via ecosystem water stress, causes vegetation mortality, and aggravate climate extremes [22, 23, 24, 25]. Previous research has investigated the effect of soil moisture availability on variability in C-flux [22, 26, 27]. The availability of soil water can have a direct impact on C fluxes by regulating the opening and closing of stomata [28, 29]. Furthermore, in semi-arid grasslands, soil water availability is critical for plant growth [30], which is directly related to C fluxes. The availability of soil water also influences enzymatic activities of plant that regulate photosynthesis and respiration rates [31]. Su *et al.* [32] found that soil moisture had a greater influence on ecosystem C fluxes than soil temperature in a semi-arid grassland on the Loess Plateau, implying the dominant control of water availability on ecosystem C fluxes rather than air/soil temperature. Zhang *et al.* [33] reported similar findings in an alpine meadow on the Tibetan Plateau. A study by Kumar *et al.* [34] documented that soil moisture and soil temperature explained 66% variation in soil CO₂ efflux in moist temperate forest of Western Himalayas. Similarly, Weissert *et al.* [35] reported that soil temperature and soil water content explained 54% and 71% of the temporal variation in soil CO₂ efflux in urban parkland and urban forests, respectively. Furthermore, Kishimoto-Mo *et al.* [36] also reported that soil temperature and moisture elucidated 69–86 % and 10–13 % of the temporal variability of soil CO₂ efflux, respectively. They documented greater soil CO₂ efflux in the top soil layer due to increased soil moisture following episodic rainfall events during summer and early fall.

2.1.1.3. Soil pH

Soil pH regulates CO₂ emissions to a larger extent by changing the rates of microbial C turnover [37]. Reth *et al.* [38] observed a significant correlation of spatial variation of soil CO₂ emission in the field with the soil pH and fine root mass, explaining up to 24% and 31% of the variability. Several studies also revealed significant effects of soil pH on soil

respiration [39, 40, 41] due to increased microbial activity with the rising pH values. Bååth and Högberg *et al.* [42, 43] demonstrated the direct positive effect of soil respiration with pH tolerance ability of the bacterial community. Optimum biological activity of soil microorganisms with nearly a linear increase of soil CO₂ emission was observed between the soil pH of 3 to 7 and 7 to 8 [44]. Similar findings reported by Courtois *et al.* [45] documented soil water content and soil pH as the main drivers of soil CO₂ effluxes across topographical positions in tropical forests of the Guiana Shield. Wanyama *et al.* [46] also found that soil pH, along with the C: N ratio and bulk density explained the majority of the variation in CO₂ fluxes in an African tropical montane region. Similarly, Singh *et al.* [47] also reported that soil pH, soil organic C, microbial biomass and soil moisture as the key regulators of soil CO₂ efflux in a tropical dry riparian ecosystem.

2.1.1.4. Soil Organic Carbon (SOC)

Soil organic carbon contains three times the amount of C found in the atmosphere, and its decomposition is likely a key source of climate change feedback and uncertainty in climate projections [48]. Soil carbon is the largest organic carbon stock in terrestrial ecosystems [49], accounting approximately two-thirds of terrestrial C [50] and about 75 Pg C year⁻¹ is being respired back to the atmosphere [51]. The SOC pool accounts for roughly half of the total forest C pool. However, in boreal and temperate forests, the SOC pool exceeds the C content of forest biomass [52]. SOC plays a critical role in regulating soil biological productivity [53]. It is critical to comprehend the temporal and spatial variation of SOC and its feedback to the atmosphere in different ecosystems [54]. A small change in the rate of SOC decomposition could have a significant impact on the atmospheric CO₂ concentration [55]. Efflux of soil CO₂ is an intricate process encompassing plant root respiration, microbial respiration as well as decomposition of soil organic matter [56, 57]. Soil CO₂ efflux is primarily governed by heterotrophic microorganisms responsible for oxidising soil organic matter during litter decomposition and plant roots respiration [58, 59, 60]. Thus, the population dynamics of soil microorganisms (e.g., bacteria, fungi, actinomycetes etc.) and soil abiotic factors (moisture, temperature, organic matter content) are the major determinates influencing soil CO₂ emissions [61, 62, 63]. Several researchers have also reported that composition of soil organic matter plays an important role in soil CO₂ efflux [64, 65, 66]. Thangavel *et al.* [67] reported a positive correlation of labile SOC fractions (POC, ROC, and MBC) with TOC and CO₂ efflux in Indian subtropical soil (Meghalaya). Similarly, influence of soil organic carbon on CO₂ efflux was also reported

for a desert steppe in Northern China and the semi-arid Loess Plateau [68, 69, 70]. Sanjita *et al.* [71] documented a strong positive correlation of seasonal SOC content with soil CO₂ efflux in a subtropical sacred grove of Manipur, North-East India. High clay content and soil organic carbon content were the key factors resulting in high CO₂ efflux from the mangrove floor in Pichavaram as reported by Gnanamoorthy *et al.* [72]. Furthermore, a positive correlation of TOC with soil CO₂ efflux rate was reported by Mishra *et al.* [73] in a representative tropical deciduous forest at Katarniaghat Wildlife Sanctuary (KWLS) of Uttar Pradesh, India.

2.1.1.5. Soil nutrients

Nutrient availability has a significant impact on ecosystem processes. In forest ecosystem, the productivity, soil microbial biomass and activities are influenced by available nutrients and in turn effect soil respiration. The availability of nutrients such as nitrogen (N) and phosphorus (P) influences SOC mineralization [74-75]. Total nitrogen and labile carbon were identified as the major determinate to regulate soil CO₂ efflux variation in Brazilian coffee agroforestry systems [76]. Liu *et al.* [77] reported that total nitrogen (0.63) exerted a positive effect on CO₂ emission rate in paddy soil as indicated by structural equation modelling. According to Mori *et al.* [78], the availability of P influences both autotrophic and heterotrophic components of soil respiration. Phosphorus not only promotes fine root growth [79] but also regulates decomposition of organic matter [80]. Fisk *et al.* [81] reported P induced enhanced mineralization of C in the surface organic soil horizon of northern hardwood forests in United States. This is further corroborated by results from Wang *et al.* [82] who reported a spatial variation of soil respiration at 10 m × 10 m scale as primarily determined by total soil phosphorus, ammoniacal nitrogen, pH and soil water content ($p < 0.05$). Kadulin *et al.* [83] reported an increase in soil respiration with increase in available soil carbon, nitrogen, phosphorus, and potassium contents.

2.1.2. Soil Biological Characteristics

2.1.2.1. Soil enzymes

Microbial processes as reflected by soil enzymatic activities and microbial biomass have been shown to improve global predictions of soil C fluxes [84, 85]. Extracellular soil enzymes catalyse the initial step of microbial decomposition of both labile and recalcitrant C (C) compounds [86, 87]. Labile C decomposes quickly and provides a high energy yield to microbes, whereas recalcitrant C decomposes slowly over years and requires multiple

enzymatic steps [88, 89, 90]. This increase in enzyme-driven decomposition could boost soil respiration, releasing more CO₂ from soil C stocks and providing positive feedback to accelerate global surface temperatures [91, 92, 93, 94, 95, 96, 97, 98]. Thus, extracellular soil enzyme driven decomposition of native organic matter (depolymerization, e.g., peroxidase) produces simple soluble compounds for microbial assimilation [99, 100, 101] that get mineralized to CO₂ with the production of ATP [102]. Sinsabaugh *et al.* [103] observed a positive correlation of β -1,4-glucosidase, cellobiohydrolase, β -1,4-N-acetylglucosaminidase, and phosphatase activities with soil organic carbon content. These enzymes catalyse the breakdown of organic matter and the mineralization of nutrients. β -1,4- glucosidase hydrolyzes the cellobiose residue in the final stage of the cellulose degradation process [104]. The final product of these reactions is glucose, an important C energy source for growth and activity of soil microorganism [105], which ultimately regulates C mineralisation. Auwal *et al.* [106] reported a positive correlation of α -glucosidase, β -glucosidase, β -xylosidase and cellobiohydrolase enzyme activities with cumulative CO₂ emissions at 25°C implying that SOM decomposition is regulated by microbes. Zhang *et al.* [107] also reported a positive correlation ($p < 0.05$) of β -glucosidase and dehydrogenase activities with cumulative soil CO₂ emission in a bamboo forest soil amended with pyrogenic and fresh organic matter. Similar findings were reported by Mohamed *et al.* [108]; Halmi and Simarani, [109] where the rate of SOC mineralization was closely related to the activities of soil -glucosidase/dehydrogenase. β -D glucosidase activity also showed the highest contribution in regulating the soil CO₂ emissions of Sundarban mangrove forest, India as reported by DE *et al.* [110]. Li *et al.* [111] reported that cellobiohydrolase ($P < 0.001$) and β -glucosidase activities ($P = 0.007$) accounted for 50% and 31% of the variation in soil respiration, respectively, in a subtropical plantation of China. A significant positive correlation of β -glucosidase and cellobiohydrolase with soil CO₂ flux was reported by Gao *et al.* [112] during both the snow cover and the snow melt periods in Larch and Chinese pine plantation forests. Zhao *et al.* [113] also documented a significant correlation of cumulative soil respiration with β -D-glucosidase ($p < 0.05$) and soil MBC ($p < 0.01$). Hugh *et al.* [114] hypothesized that the activity of β -D glucosidase, which is involved in the breakdown of polysaccharides would increase in response to elevated CO₂ in a California annual grassland.

Furthermore, under elevated CO₂, the activities of the enzymes N-acetylglucosaminidase and acid phosphatase increased significantly in a temperate forest at Changbai Mountains in North-eastern China, as documented by Zheng *et al.* [115].

2.1.2.2.MBC

Soil microbial biomass is the most active component of SOC, regulating biogeochemical processes in terrestrial ecosystems [116]. It is the labile fraction of soil organic matter (SOM) that is highly sensitive to changes in biotic and abiotic factors. Despite accounting for a small portion of total soil organic carbon (TOC), microbial biomass has been found to be a more sensitive indicator of changes in SOM levels than SOC content [117]. Several studies have also found a significant relationship between labile fractions of SOC and soil respiration, particularly the microbial biomass carbon pool [118, 119, 120]. Microbial biomass carbon (MBC) was found to be significantly related to variation in soil respiration ($P = 0.043$) across China's subtropical forests during the warm season [121]. Similar findings were reported by Wang and Gu [122], where a strong correlation between soil heterotrophic respiration with microbial biomass carbon ($r = 0.761, 0.801, 0.923$) was noted in subtropical forests in China. Iqbal *et al.* [119] reported that microbial biomass carbon ($R^2 = 0.39-0.86$) and dissolved organic substances ($R^2 = 0.50-0.95$) have significant effect on temporal variation of soil CO₂ fluxes in the Three Gorges Reservoir Area of South China. Han and Zhu [123] reported a positive correlation of soil CO₂ efflux with soil microbial biomass carbon and root biomass, within or across all types of primary forest conversions of China. According to Wei *et al.* [124], MBC and fine root biomass were positively correlated with soil respiration. Furthermore, the MBC and fine root biomass co-explained 70% of the variation in soil respiration, suggesting that the MBC and fine root biomass co-controlled soil respiration. According to Feng and Zhu [125], autotrophic and heterotrophic soil respiration increased in lockstep with MBC, SOC, and belowground biomass. Similar findings were reported by Zhao *et al.* [30]; Ding *et al.* [126]; Yang *et al.* [127] where SOC, MBC, and DOC found to contribute towards rising CO₂ level. Mirzaei *et al.* [128] observed a positive correlation of CO₂ efflux with higher soil temperature (0.43–0.79) and MBC (0.66–0.89) in a semi-arid region of Iran under corn-wheat rotation.

2.1.2.3.Soil metagenomics

Soil microorganisms play a pivotal role in regulating C dynamics of soil through control on decomposition of soil organic matter (SOM) and nutrient-cycling rates [129]. Microbial diversity has been linked to a variety of soil functions, including soil respiration [130, 131]. Furthermore, because of the strong positive relationships that exist between microbial composition and the functional genes that regulate soil respiration [132],

community composition (relative abundance of phylotypes) may improve our prediction of soil respiration in addition to diversity [133]. Several microbial taxa regulate changes in SOC fractions by participating in the decomposition of soil organic matter, thereby influence soil CO₂ efflux [134-135]. Many molecular techniques such as polymerase chain reaction-denaturing gradient gel electrophoresis (PCR-DGGE), terminal restriction fragment length polymorphism (T-RFLP), and high-throughput sequencing are now widely used to study the soil microbial community [136, 137, 138]. Although common high-throughput sequencing can provide information on microbial communities, it does not provide detailed information on microbial metabolic function. However, metagenomic sequencing, which sequences the extracted DNA directly, can furnish elaborate information on both microbial classification and their metabolisms [139]. Padhy *et al.* [140] reported that metagenomic analysis of soil bacteria and archaea could give valuable insight of the structural feature and microbial functionality of the ecosystem related to GHGs emissions and C-dynamics.

Several researchers believe that linking C mineralization to taxonomic and functional characteristics of microbial communities could improve predictions on soil respiration dynamics [141, 142, 143, 144]. Furthermore, various experiments demonstrated that copiotrophs (such as *Alphaproteobacteria*, *Gammaproteobacteria* and *Bacteroidetes*) and oligotrophs (such as *Deltaproteobacteria*, *Acidobacteria* and *Actinobacteria*) played distinct ecological functional roles in utilisation of C for respiration [145, 146, 147, 148]. Liu *et al.* [144] identified the major bacterial phyla such as *Alphaproteobacteria*, *Deltaproteobacteria* and WPS.2 for predicting soil respiration ($P < 0.05$) using random forest modelling. Significant relationships were found between respiration and MBC, relative abundance of bacteria, bacterial community and richness. Similarly, *Acidobacteria* and *Alphaproteobacteria* were identified as the major microbial taxa governing changes in soil respiration along elevational gradients of alpine forests in China [149]. Colombo *et al.* [150] reported that relative abundance of bacteria is most crucial predictor of basal respiration ($\Delta AICc = 11.090$). Ren *et al.* [151] observed a significant correlation of bacterial abundance (*Proteobacteria*, *Bacteroidetes*, *Actinobacteria*, *Chloroflexi*, and *Nitrospirae*) and bacterial alpha diversity with SOC, DOC, MBC, and soil respiration components ($p < 0.001$). Tardy *et al.* [134] reported microbial richness as the principal predictor of the C cycling activities, with bacterial and fungal richness accounting for 32.2 and 17% of the SOM mineralization, respectively. Liu *et al.* [152] observed a significant ($P < 0.05$) relationship between bacterial community composition

and rate of soil respiration in both paddy and upland soils. Furthermore, random forest modelling suggested that the phylotypes within the phyla/classes *Alphaproteobacteria*, *Bacteroidetes*, *Betaproteobacteria*, *Actinobacteria*, *Gammaproteobacteria* and *Firmicutes* were among the numerous bacterial predictors regulating soil respiration.

2.2. Soil Mineralogy and Carbon Sequestration

Physical and chemical associations of soil C with minerals play an important role in storage of C [153]. Available minerals have been found to have a strong correlation with soil C and its long-term stability [154]. Interactions between soil organic carbon (OC) and mineral soil particles reduce OC accessibility to degraders, thereby controlling its susceptibility to mineralization and subsequent release into the atmosphere [155, 156, 157, 158]. Phyllosilicates (layer-lattice aluminosilicates), metal oxides, hydroxides, and oxyhydroxides (e.g., hematite, gibbsite, goethite), and short-range ordered aluminosilicates (e.g., imogolite, allophane) are the most common minerals in soil environments that contribute to stabilisation of organic matter. Several factors such as size, shape, polarity, surface topography etc. govern the surface interaction of organic molecules with these minerals. A substantial number of studies infers that soil OM storage capacity is largely determined by the mass proportion of fine mineral particles with high surface area, such as phyllosilicates and Fe/Al oxides and hydroxides [159, 160, 161, 162, 163, 164].

2.2.1. Clay minerals

In soils, organic matter (OM) can be stabilized via three mechanisms: (1) by biochemical recalcitrance, (2) formation of organomineral complexes through chemical interactions with minerals and metal ions, and/or (3) physical protection owing to occlusion within soil aggregates [165]. It is also worth noting that not only the clay content, but also the type of clay, has a significant impact on SOC stabilisation. Soil clay minerals (metal oxides, oxyhydroxides, and hydroxides) are the most reactive soil fractions and form associations with organic matter [166, 167, 168]. In addition to phyllosilicate clay minerals, the Fe and Al sesquioxide also play an important role in soil OM (SOM) protection. Numerous studies have found a positive correlation of SOM content with the amount of Fe and Al sesquioxides in soils [169, 170, 171]. It has been reported that soils rich in sesquioxide clays stabilise more C than those without them [169, 172]. The Fe oxide coating on soil clay particles may affect the surface charge of clay minerals [157, 167]. As a result, a

strong complexation between positively charged clay and negatively charged OM may occur. This complex formation reduces the accessibility of microorganisms to OM as a food source and regulate soil respiration [167].

Nguyen and Marschner [173] reported that addition of smectite clay soil decreased cumulative respiration per g organic C in the sand-clay mixes compared to sandy soil alone at 22.5 and 30% clay addition rates. Addition of kaolinite clay soil had no effect at any clay addition rate. The low cumulative respiration of the mixes with smectite clay soil can be explained by its high surface area and CEC which reduces the accessibility of OC to microbes [174]. Singh *et al.* [172] reported that the smectitic soil recorded the highest basal respiration followed by the kaolinitic-illitic and allophanic soils. A strong inverse correlation ($R^2 = 0.90$ at $p < 0.05$) was observed between CO_2 emission rate and total sesquioxides (Fe and Al oxides) content. Saïdy *et al.* [175] reported that clay mineralogy and the presence or absence of goethite significantly influence the cumulative C mineralisation ($P < 0.05$). Sagggar *et al.* [176] recorded that the amount of C remaining after a 35-days of incubation period was highest in soils dominated by smectite. The same was found to be lowest in soils dominated by vermiculite and kaolinite. The differences in the specific surface area (SSA) of the clays reflects the observed variation in the rate of C mineralisation. The higher SSA of the illitic and smectitic clays compared to the kaolinitic clay have been suggested to enable more and stronger SOC-clay interactions and thereby facilitate SOC stabilisation [177, 178]. Kirsten *et al.* [179] documented that clay minerals, metal oxides and hydroxides are the most reactive soil mineral constituents controlling the long-term stability of organic carbon (OC) in terrestrial ecosystems. Churchman *et al.* [180] reported that adding clay to soils increased OC, potentially increasing net C sequestration. Rakhsh *et al.* [181] observed that the interaction of OC and microbial biomass with clay minerals leads to their stabilisation in soils. They also reported that addition of a small amount of clay (5%) significantly delays the decomposition of OC thereby reducing soil carbon dioxide emission. Yu *et al.* [182] showed that in a coniferous forest, stabilisation of SOC was due to sorptive protection or microaggregate formation of soil clay fraction with SOC ($r^2 = 0.89$, $p < 0.05$). Whereas, in a mixed or broadleaved forest, SOC stabilisation was due to chemical protection through formation of organo-mineral complexes.

2.3. Weather Variables and Ecosystem Carbon Flux

Global climate change is expected to aggravate and cause regional drought events, particularly in mid-latitude and subtropical dry regions due to rising global temperatures and changing precipitation regimes [183]. This changing climatic scenario may have a pronounced effect on terrestrial C cycling due to alteration in the structure and function of forest ecosystems [184, 185, 186, 187]. Majority of field studies propound that global warming generally stimulates plant growth and ecosystems C flux (e.g., NPP, ecosystem respiration and ecosystem photosynthesis). Soil temperature and water availability have a strong influence on soil respiration, and these two environmental factors are closely related to climate change and changing precipitation patterns [188, 189, 190].

2.3.1. Temperature

Temperature is a major driver of soil respiration -a key process of C efflux to the atmosphere [191, 192, 193]. In general, increase in soil temperature below 35 °C increases soil respiration, as long as soil moisture is not a limiting factor (in the range from 0.20 to 0.30 cm³ cm⁻³) [194, 195]. The correlation between soil temperature, soil CO₂ efflux and rate of CO₂ production have been depicted by various researchers using the Q₁₀ model [195, 196]. Liu *et al.* [197] found a parabolic relationship between soil respiration and soil temperature during the growing season in a desert steppe of Northern China, accounting for about 19% of the total variation in soil respiration ($p < 0.01$). Dacal *et al.* [198] observed a consistent and significant positive effect of assay temperature (5 °C, 15 °C and 25 °C) on soil respiration across the eight chrono sequences studied ($p < 0.001$). Bao *et al.* [199] observed that rising temperature resulted in increased soil respiration on the Tibetan Plateau. A meta-analysis reported that experimental warming significantly increased soil respiration in forest ecosystems, owing to rapid decomposition of SOC [184, 185]. Soil warming considerably increased soil respiration by 32.0% in 2011 and 46.3% in 2012 under ambient precipitation as reported by Liu *et al.* [200]. Yu *et al.* [201] observed a significant correlation of soil temperature with soil CO₂ efflux and rate of CO₂ production in an oasis cotton field in arid North-western China. Yang *et al.* [202] documented a positive correlation of air temperature with CO₂ fluxes ($r = 0.444$, $P < 0.01$) in two subtropical estuaries of China. Wang *et al.* [203] indicated that the interactive effects of soil organisms on microbial biomass, respiration, and carbon use efficiency are regulated by temperature, providing a foundation for understanding the soil C cycle in forest ecosystems. Zhou *et al.* [204] reported a significant positive correlation between accumulated effective soil temperature and accumulative soil CO₂ efflux on a monthly

scale in temperate forests of Northern China. Soong *et al.* [205] observed a sustained $30 \pm 4\%$ increase in soil CO₂ efflux due to warming over a period of five years through the whole-soil profile in an experimental forest of California. Similarly, Lin *et al.* [206] documented warming induced hike in surface soil CO₂ efflux by 40% in a young Chinese fir plantation. Zou *et al.* [207] reported an increase in average soil CO₂ efflux by 23% due to rise in soil temperature in a forest from central Ireland.

2.3.2. Precipitation

Intensification of precipitation rates may affect soil CO₂ emissions from forests to grasslands having a direct impact on the C balance of the ecosystem [208]. One of the key factors influencing soil CO₂ efflux in terrestrial ecosystems is the changing precipitation. Increased precipitation has been shown to promote soil respiration in water-stressed ecosystems. This is primarily due to stimulation of many ecological processes such as plant growth, soil microbial growth and activity, and the temperature sensitivity of soil respiration [209, 210, 211, 212, 213, 214]. Moreover, recent findings have revealed that increased precipitation has a substantial impact on soil respiration in dry condition than that in the wet condition resulting in a saturated relationship between soil respiration and precipitation [209, 215]. Ahlström *et al.* [216] have reported that the high variability in precipitation regulates the interannual variability of the global C sink. A myriad of studies has reported that intensification of precipitation usually results in higher soil moisture contents, as well as increased soil respiration thereby accentuating soil CO₂ efflux in various ecosystems [217, 218, 219]. Zhang *et al.* [220] observed that total soil respiration and its components increased nonlinearly with increase in precipitation (all $P < 0.01$). They reported a 70%, 74% and 40% (marginal R²) variations in total soil respiration, heterotrophic soil respiration and autotrophic soil respiration due to change in precipitation while considering year as a random effect. The findings of Huang *et al.* [221] illustrated a linear increase in soil respiration with increased precipitation in a temperate desert ecosystem. Legesse *et al.* [222] found that wet spring and wet summer precipitation treatments increased soil respiration by 24.9% and 24.1%, respectively, compared to dry spring and dry summer precipitation treatments in mowed grassland of Inner Mongolia. Soil respiration was strongly affected by precipitation changes in a switchgrass field of the United States [223]. The findings of Ngaba *et al.* [224] showed that changing precipitation significantly increased soil respiration by 51 % and 17 % ($P < 0.05$), in forests and croplands, respectively, across diverse biomes. According to Liu *et al.* [68], increased

precipitation had a significant effect on soil respiration during the growth seasons ($p < 0.05$) in a desert steppe of northern China, accounting for approximately 16.3% of the variation in soil respiration ($p < 0.01$). Yue *et al.* [225] showed that precipitation and nitrogen deposition significantly elevated soil respiration in a temperate desert of China. Zhang *et al.* [226] reported a positive correlation of soil respiration with precipitation-induced change in above-ground plant biomass in the desert-grasslands of Inner Mongolia, China. They found an increase in C efflux from the soil with an increase in precipitation. Wang *et al.* [227] documented a stronger and faster precipitation pulse effects on soil microbial respiration in Beijing secondary forest soils. Pan *et al.* [228] used the Community Land Model Version to simulate soil heterotrophic respiration over a 33-year period (1980-2012) in the central Tibetan Plateau. The model output revealed a significant linear regression between seasonal fluctuations in heterotrophic respiration and precipitation. Furthermore, a meta-analysis of all field experiments revealed that a 31.81% increase in precipitation significantly increased soil CO₂ efflux by 12.74% [229].

2.4. Ecosystem Characteristics and Carbon Flux

2.4.1. Grass land ecosystem

Grassland ecosystems, which cover up to 40% of the global land surface, are critical modulators of the C cycle and climate. The eddy covariance technique and model efforts revealed that most of the grasslands have gross primary production (GPP) that exceeds ecosystem respiration and thus act as C sinks [230, 231, 232, 233, 234]. However, various factors such as vegetation, growing season, environmental factors and soil factors play a key role in regulating the C flux. Niu *et al.* [235] indicated that the sandy grassland behaved as a net CO₂ source on an annual scale with a mean annual net ecosystem exchange (NEE) of 49 ± 8 g C m⁻² yr⁻¹. On a seasonal scale, the sandy grassland absorbed net CO₂ during the summer but released net CO₂ during the other seasons. During the dry study periods, a grassland in New Mexico, USA, was a net source of 31 g C m⁻² yr⁻¹ [236]. Arredondo *et al.* [237] documented that mixed grassland treatment exhibited higher soil respiration rates than the monospecific *Bouteloua gracilis* grassland during the dry months (November to April). This could be attributed to greater plant interspace extensions in mixed grassland, which leads to increased radiation exposure and accelerated cycles of desiccation, elevated soil temperature, sensible heat flux, and re-wetting. Song *et al.* [238] reported that the grassland ecosystem of central Qinghai-Tibet Plateau shifted from a C sink during growing season (68.8 ± 8.7 g C m⁻²) to a C source in nongrowing season

($-41.1 \pm 2.4 \text{ g C m}^{-2}$). Whereas, the net ecosystem carbon budget ($27.7 \pm 6 \text{ g C m}^{-2} \text{ yr}^{-1}$) demonstrated a net C sink at an annual basis. The findings of Zhu *et al.* [239] indicated that northern China's temperate grasslands can significantly alter vegetation biomass, soil temperature, and soil water content, resulting in temporal and spatial variation in CO₂ fluxes. De Long *et al.* [240] demonstrated that shoot biomass is an important factor in grassland ecosystem respiration in both monocultures and mixed communities. This could be because of total plant biomass as plants are the primary source of C fixation via photosynthesis. A meta-analysis study found that warming generally stimulated ecosystem C and N cycles, but warming had little effect on soil C and N stocks in alpine grassland ecosystems on the Tibetan Plateau [241]. Wagle *et al.* [242] demonstrated that temporal and spatial variation of water use efficiency, evapotranspiration and C uptake in grassland ecosystem were strongly related to canopy coverage and greenness, as indicated by the enhanced vegetation index (EVI) in 12 grasslands across the United States. A grassland in Brazil's semi-arid region served as an atmospheric C source during drier periods with lower normalised difference vegetation index (NDVI) and as a C sink during wetter periods with higher NDVI as reported by da Silva *et al.* [243]. Chen *et al.* [244] reported that warming increased above-ground plant respiration and total autotrophic plant respiration by 28.7% and 19.9%, respectively, while decreased heterotrophic respiration by 10.4% in a Tibetan Plateau meadow grassland.

Furthermore, several other studies [245, 246] found grassland as a C sink during the growing season. Thus, diverse grassland ecosystems will respond differently to changing climatic scenario in near future.

2.4.2. Forest Ecosystem

Forests, extending from tropical to temperate to boreal cover approximately 30% of the total land surface. They play a vital role as soil C sinks by storing significant amount of stable organic matter [247]. Tropical and subtropical forests have made a significant contribution to gross forest C fluxes, accounting for 78% of total emissions ($6.32.4 \text{ Gt CO}_2\text{e yr}^{-1}$) and 55% of total removals ($8.67.6 \text{ Gt CO}_2\text{e yr}^{-1}$). Thus, only 30% of the global net C sink came from tropical and subtropical forests. Whereas, temperate (47%) and boreal (21%) forests accounts for roughly two-thirds of the global net C sink, owing to significantly lower gross emissions in these regions [248]. Modelling and experimental studies have been carried out to gain a better understanding of the possible impacts of elevated atmospheric CO₂, changing precipitation patterns and global warming on

ecosystem processes of these forests' types [249]. Climate change has been found to have both positive effects (e.g., increased forest growth and water use efficiency under elevated CO₂) and negative effects (e.g., reduced growth due to the combined effects of climate change and O₃) on forests, and these effects may vary depending on biotic and abiotic factors [250, 251].

The Amazon rainforest was identified as a C sink by Rödig *et al.* [252], with a net removal of 0.56 Gt C yr⁻¹. This C sink is driven by an estimated mean gross primary productivity (GPP) of 25.1 t C ha⁻¹ yr⁻¹ and 4.2 t C ha⁻¹ yr⁻¹ of woody aboveground net primary productivity (WANPP). The findings emphasised the importance of taking forest structure into account when simulating C dynamics. According to the findings of Phillips *et al.* [253], the average annual C sink into mature forests of the Amazon nations (Suriname, Peru, Guyana, French Guiana, Ecuador, Columbia and Bolivia) has been at least twice the magnitude of C emissions from fossil fuel burning of the same nations since 1980. Yang *et al.* [254] reported that mean annual precipitation, mean annual temperature and LAI accounted for 74% of variation in stem CO₂ efflux of forest ecosystems at global scales. Verduzco *et al.* [255] discovered that a monsoon precipitation threshold of 350 to 400 mm caused a shift in the annual C balance in tropical dry forest ecosystem of Mexico from a net source (+102 g C m⁻² yr⁻¹) to a net sink (-249 g C m⁻² yr⁻¹). Martins *et al.* [256] documented that functional microbial groups, along with water and substrate availability are the most important predictors of GHG emissions in an Australian dryland forest ecosystem. Sarma *et al.* [257] reported that the estimated annual net ecosystem productivity of 92.93 ± 1.7 g C m⁻² year⁻¹ for a semi-evergreen forest of Northeast India indicating the forest as a moderate sink of CO₂.

2.4.3. Wetland Ecosystem

Wetlands, which account for only 6% of the Earth's land surface [258], have been identified as a significant source and sink of GHG [259]. GHG emissions from wetland ecosystems vary greatly in space and time, depending on wetland type, soil properties, and climatic conditions [258, 260, 261, 262]. Wetlands play a key role in the C cycle, storing upto 15% of the total C in terrestrial ecosystems globally [263, 264, 265]. C sequestration in wetlands as determined by the net balance of their C fluxes has gained prominence in recent years. Accumulation of organic matter is more in wetlands than other ecosystems due to their higher net primary productivity (NPP) and slower rate of organic matter decomposition [266, 267, 268, 269, 270]. Mitsch *et al.* [271] and Mitsch [272]

demonstrated that despite accounting for only about 5-8% of the terrestrial landscape, the world's wetlands may be the net C sinks accounting about 0.83 Pg yr⁻¹ of C, with an average net C retention of 118 g C m⁻² yr⁻¹. Warmer temperatures and drier conditions associated with climate change are expected to shift the balance between ecosystem photosynthesis and respiration, potentially reversing a peatland's function from sink to source [273, 274]. Barbera *et al.* [275] reported increased microbial activity and plant respiration with rising air temperature in constructed wetlands. Salimi *et al.* [276] documented that drought might shift the role of wetlands from C sinks to C sources. However, higher temperature and rainfall can help to keep wetlands functioning as C sinks. According to Cao *et al.* [277], an alpine wetland ecosystem in Qinghai Lake area of China acts as a C sink throughout the year with an average net ecosystem CO₂ exchange (NEE) of -904.42 g CO₂/m². Pugh *et al.* [278] demonstrated that over multiyear timescales, annual temperature is the better predictor for interannual variation of wetland net C fluxes in northern Wisconsin rather than water table fluctuations. Cui *et al.* [279] found that mangrove wetlands of subtropical China could sequester more CO₂ from the atmosphere than nearby terrestrial forests due to higher gross ecosystem production (GEP) and lower ecosystem respiration (R_e) values. Lu *et al.* [280] used 143 site-years of eddy covariance data from 22 inland wetland and 21 coastal wetland sites around the world to conduct a meta-analysis comparing ecosystem CO₂ fluxes among different types of wetlands. They found that coastal wetlands were large CO₂ sinks compared to inland wetlands or the later was nearly CO₂ neutral. Furthermore, they observed that annual CO₂ fluxes were mainly regulated by mean annual precipitation (MAP) and mean annual temperature (MAT) which explained 71%, 57% and 54% of the variations in gross primary productivity (GPP), net ecosystem productivity (NEP) and ecosystem respiration (R_e), respectively.

Thus, ecosystems behave differently in regulating the CO₂ efflux based on the ecosystem characteristics and changing climatic conditions.

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