AGU ADVANCING EARTH AND SPACE SCIENCES

Reviews of Geophysics

INVITED ARTICLE

10.1029/2023RG000829

The Impacts of Erosion on the Carbon Cycle

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Key Points:

- Quantification of erosion-induced carbon fluxes on decade–century timescales is vital for the global carbon budget
- The magnitudes of lateral and vertical carbon fluxes caused by physical and chemical erosion are synthesized
- Combining physical- and chemical-erosion-induced carbon dynamics can reduce global carbon budget biases

Supporting Information:

Supporting Information may be found in the online version of this article.

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Citation:

Zheng, H., Miao, C., Huntingford, C., Tarolli, P., Li, D., Panagos, P., et al. (2025). The impacts of erosion on the carbon cycle. *Reviews of Geophysics*, *63*, e2023RG000829. https://doi.org/10.1029/2023RG000829

Received 4 DEC 2023
Accepted 18 NOV 2024

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Abstract Physical and chemical erosion associated with water both affect land–atmosphere carbon exchanges. However, previous studies have often addressed these processes separately or used oversimplified mechanisms, leading to ongoing debates and uncertainties about erosion-induced carbon fluxes. We provide an overview of the on-site carbon uptake fluxes induced by physical erosion (0.05–0.29 Pg C yr^{-1}, globally) and chemical erosion (0.26–0.48 Pg C yr^{-1}). Then, we discuss off-site carbon dynamics (during transport, deposition, and burial). Soil organic carbon mineralization during transport is nearly 0.37–1.20 Pg C yr^{-1} on the globe. We also summarize the overall carbon fluxes into estuaries (0.71–1.06 Pg C yr^{-1}) and identify the sources of different types of carbon within them, most of which are associated with land erosion. Current approaches for quantifying physical-erosion-induced vertical carbon fluxes focus on two distinct temporal scales: short-term dynamics (ranging from minutes to decades), emphasizing net vertical carbon flux, and long-term dynamics (spanning millennial to geological timescales), examining the fate of eroded carbon over extended periods. In addition to direct chemical measurement and modeling approaches, estimation using indicators of riverine material is popular for constraining chemical-erosion-driven carbon fluxes. Lastly, we highlight the key challenges for quantifying related fluxes. To overcome potential biases in future studies, we strongly recommend integrated research that addresses both physical and chemical erosion over a well-defined timescale. A comprehensive understanding of the mechanisms driving erosion-induced lateral and vertical carbon fluxes is crucial for closing the global carbon budget.

Plain Language Summary Erosion—driven by internal forces such as mountain uplift and external forces such as water, wind, and human activities—plays a pivotal role in altering land carbon storage. It exerts intricate influences on carbon cycling involving physical and chemical transformations. Physical erosion affects soil organic carbon through processes like disaggregation, transport, deposition, and deep burial, while chemical erosion influences land carbon uptake or release through chemical weathering of minerals and rocks. Here, we thoroughly examine erosion-induced carbon dynamics, distinguishing between on-site processes (occurring at the original erosion site) and off-site processes (pertaining to the fate of carbon removed from its original, now-eroded location). We provide an in-depth analysis of how physical and chemical erosion impact carbon dynamics, then offer quantitative estimates of erosion-related carbon fluxes for key processes. Additionally, we develop a new conceptual framework for quantifying the erosion-related carbon fluxes. Although accurately quantifying the impacts of erosion on carbon cycling remains challenging, we believe that the use of modern research tools such as advanced monitoring tools and geostatistical modeling, remote sensing databases, and artificial intelligence offers promising solutions.

1. Introduction

The global soil carbon pool, storing ∼1,550 Pg of organic carbon and ∼950 Pg of inorganic carbon in the top meter of soil, is approximately three times the atmospheric carbon pool (Lal, 2004a). Even a weak disturbance to the soil carbon pool has the potential to cause a distinct fluctuation in atmospheric carbon levels (Houghton, 2003). It has been estimated that the carbon emissions from fossil fuel burning and land use change in

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2010–2020 were ~9.5 and ~1.1 Pg C yr^{-1}, respectively (Friedlingstein et al., 2022). Most of the emitted carbon is stored in the atmosphere (~5.2 Pg C yr^{-1}) and oceans (~2.7 Pg C yr^{-1}), while the remaining carbon (~2.7 Pg C yr^{-1}) is sequestered in soil and vegetation (Friedlingstein et al., 2022; Regnier et al., 2022). However, direct estimations of land carbon uptake have large uncertainties and often fail to balance the carbon budget (Ahlström et al., 2012; Bloom et al., 2016; Gurney et al., 2002), leaving room for the so-called "missing carbon sink" (Fang & Guo, 2007; Houghton et al., 1998; Lal, 2004b). It is frequently suggested that the missing sink probably has close associations with soil organic carbon (SOC) burial, carbon uptake by weathering processes, and carbon replacement by erosion (Hoffmann et al., 2013; Regnier et al., 2013; Stallard, 1998; Zondervan et al., 2023). The overall size of the sink is estimated to be approximately 1.0–1.4 Pg C yr^{-1} (Schimel, 1995; Stallard, 1998).

Erosion (the definition can be found in the Glossary) is typically ascribed to mountain uplift, water, wind, and human interference (Doetterl et al., 2016; Hilton & West, 2020). The physical erosion of soil has a critical influence on land–atmosphere carbon flux; physical weathering, by contrast, essentially lacks carbon exchange processes (White & Buss, 2014). Given the much smaller carbon flux caused by wind erosion in contrast with water erosion (Y. Liu et al., 2023), physical and chemical erosion associated with water have been identified as the predominant contributors to erosion-induced carbon fluxes (Hilton & West, 2020; Van Oost et al., 2012). The relationships among the various types of erosion mentioned above are described in Figure S1.

In the context of unprecedented human-induced atmospheric warming, parallel rapid expansion of urban and agricultural lands (IPCC, 2022), abrupt permafrost degradation, and other human disturbances, global water erosion has been accelerating (Borrelli et al., 2020; D. Yang et al., 2003). Erosion promotes carbon burial (Van Oost et al., 2012) and increases chemical weathering rates (Hilton & West, 2020; Willenbring & von Blanckenburg, 2010), thereby serving as a sink for atmospheric carbon (Berhe et al., 2008; Hilton et al., 2015; Van Oost et al., 2007). However, erosion also leads to decreased soil productivity and enhanced mineralization of migrated organic carbon, exerting an opposing effect and weakening the "draw down" of atmospheric CO_2 (Feng et al., 2018; Lal, 2019; Polyakov & Lal, 2008). To date, estimates of vertical carbon fluxes induced by water erosion span a wide range (from −1.2 to 1.5 Pg C yr^{-1}; see Figure 1). This indicates substantial uncertainty, and importantly, the range spans both signs (i.e., the net effect could be either positive—a sink effect—with net carbon movement from the atmosphere to the land, or negative—a source effect—with net carbon movement from the land surface to the atmosphere). The primary reasons for this large uncertainty are the absence of a unified, systematic assessment approach and the use of varying timescales (Berhe et al., 2008; S. Liu et al., 2003; Lugato et al., 2018).

The isolated study of either vertical or lateral carbon fluxes often makes the "sink" effect ambiguous, and the transport of riverine carbon to the ocean offers an indirect loss of (or sink for) atmospheric carbon. Rivers serve as vital connections between the products of physical and chemical erosion, linking biogeochemical processes across land, ocean, and atmosphere (Aufdenkampe et al., 2011; Ran et al., 2015). However, little attention has been given to understanding the relationships among riverine carbon, terrestrial erosion, and chemical weathering (Drake et al., 2018). In addition, almost all previous studies or reviews consider only physical or chemical erosion (Doetterl et al., 2016; Emberson et al., 2016; Hilton & West, 2020; Yue et al., 2016), without quantifying the interaction between them, which can lead to an incomplete understanding of the full impact of these erosion processes. Any incomplete assessment of erosion-induced carbon fluxes will likely generate conclusions that are biased (Doetterl et al., 2016). To avoid biased conclusions about erosion-induced carbon effects on global or continental scales, this review illustrates the full combined effects of overall geomorphic and geologic processes, as depicted in Figure 2.

Here, we provide a new, more holistic analysis of both on-site and off-site erosion-induced carbon fluxes based on an extensive literature review. First, we provide an overview of changes in on-site and off-site carbon dynamics induced by physical and chemical erosion separately. Next, we clarify the methods used to quantify carbon fluxes associated with physical and chemical erosion. Finally, we address key challenges for accurately quantifying erosion-caused carbon fluxes and outline future directions. A more precise calculation of erosion-induced carbon fluxes supports the estimation of changes in the soil–vegetation carbon pool and underpins policy guidelines for agricultural land management and carbon capture and storage in carbonate rocks to mitigate future climate change. By these means, our study contributes to advancing global carbon-neutral solutions.

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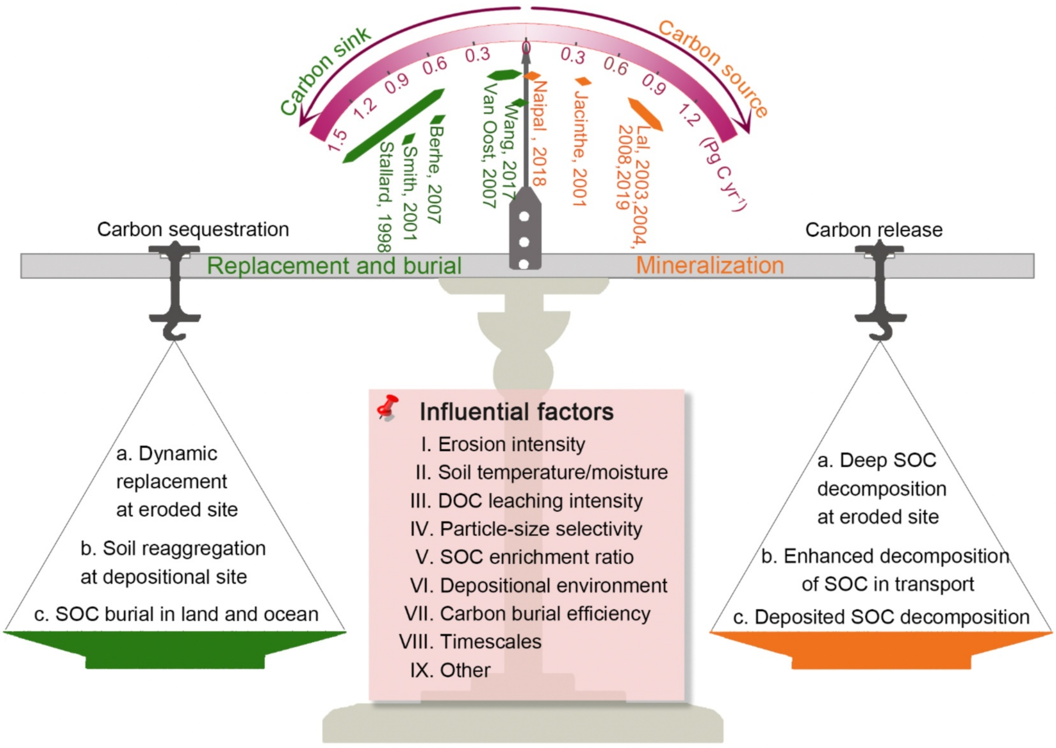

Figure 1. The balance between key processes of water-erosion-induced carbon sinks and sources in organic carbon processes. The middle pink box lists several important factors affecting the balance. DOC = dissolved organic carbon; SOC = soil organic carbon. The references beside the needle have ranges as follows: Stallard (1998), 0.6–1.5 Pg C yr^{-1}; S. V. Smith et al. (2001), ~1 Pg C yr^{-1}; Van Oost et al. (2007), 0.06–0.27 Pg C yr^{-1}; Berhe et al. (2007), 0.72 Pg C yr^{-1}; Z. Wang et al. (2017), 56–100 Pg C from 6000 BCE to 2015 CE; Jacinthe and Lal (2001), 0.37 Pg C yr^{-1}; Lal (2003, 2004b, 2008, 2019), 0.8–1.2 Pg C yr^{-1}; Naipal et al. (2018), 2 Pg C from 1850 to 2005. Note that the research targets of these studies, such as farmland, agricultural land, and global land, may vary.

2. On-Site Carbon Dynamics Induced by Physical and Chemical Erosion

In organic carbon cycling, the primary processes that maintain the carbon balance between land and atmosphere are organic carbon synthesis (through plant photosynthesis) and decomposition (through autotrophic and heterotrophic respiration, as well as nonbiological oxidation) (Hilton & West, 2020). Generally, the cycling period for organic carbon ranges from minutes to hundreds of years (i.e., relatively short timescales), and the involved carbon, approximately 100 Pg C yr^{-1}, is usually young and active (Mayorga et al., 2005). Physical erosion has a persistent effect on land carbon uptake, decomposition, and the transport in river systems (Figures 3a, 3c, and 3d). The magnitude of the effect varies significantly between natural and human-accelerated erosion (Regnier et al., 2022). The vertical carbon flux induced by human-accelerated erosion may range from −1.2 to 1.5 Pg C yr^{-1} worldwide, which corresponds to approximately 10% of the carbon emissions from fossil fuel combustion (global mean annual emitted flux of ~10 Pg C yr^{-1} during 2010–2019) (Berhe et al., 2007; Hilton & West, 2020). This highlights the importance of considering both human-induced climate change and direct land-use changes in future projections of large-scale environmental shifts.

In contrast with organic carbon cycling, inorganic carbon cycling takes place over a long, "geological" timescale (often from thousands to millions of years) (Berner & Caldeira, 1997), and the age of the associated carbon typically exceeds thousands of years (Mayorga et al., 2005). Inorganic carbon cycling processes include the dissolution of atmospheric CO_2 in rainfall, the chemical weathering of minerals in soil or rocks, the transport of weathered materials, carbonate sedimentation in oceans, and carbon release from the crust (such as volcanic degassing) (Figure 2). This balanced mechanism is relatively slow under natural erosion conditions, with the amount of inorganic carbon involved potentially being less than 0.1 Pg C yr^{-1} (Hilton & West, 2020; W. Li et al., 2022). However, the amount of carbon involved in accelerated chemical weathering, for which physical

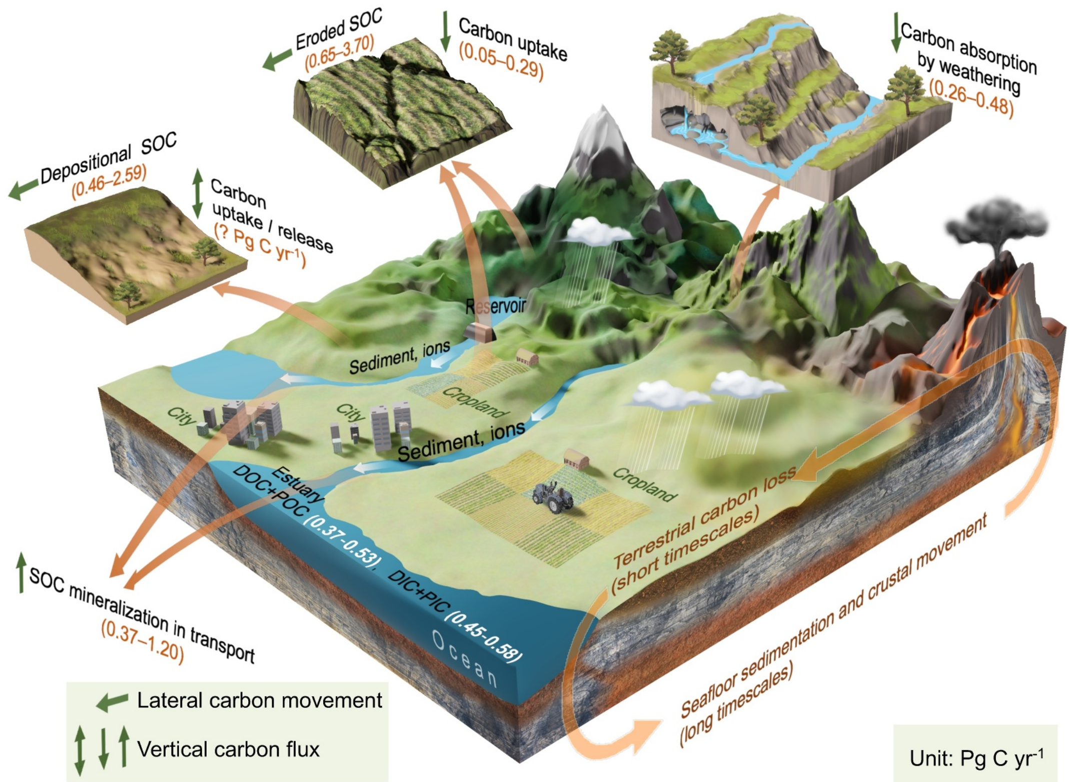

Figure 2. Overview of global erosion-induced lateral and vertical carbon fluxes (units of all fluxes: Pg C yr^{-1}). The carbon absorption estimate for chemical weathering is referenced (Hartmann et al., 2013; Hilton & West, 2020; Suchet & Probst, 1995; Willenbring & von Blanckenburg, 2010). Assuming global soil loss by water erosion is 35–200 Pg yr^{-1} and average soil organic carbon (SOC) content is 1.85% (Borrelli et al., 2021; Naipal et al., 2018; S. V. Smith et al., 2001), the total amount of global eroded carbon is estimated to be 0.65–3.70 Pg C yr^{-1}. Assuming the depositional ratio is 70% (Walling & Webb, 1996), the total deposited carbon is about 0.46–2.59 Pg C yr^{-1}. Vertically, assuming the net carbon absorption flux caused by SOC dynamic replacement is approximately 26% of the net eroded carbon (eroded carbon minus deposited carbon; Van Oost et al., 2007), the net flux is 0.05–0.29 Pg C yr^{-1}. Global mineralization originating from migrated SOC is about 0.37–1.20 Pg C yr^{-1} (Jacinthe & Lal, 2001; Lal, 2003). Dissolved organic carbon, particulate organic carbon, dissolved inorganic carbon, and PIC (particulate inorganic carbon) fluxes are referenced (Ludwig et al., 1996; Schlesinger & Melack, 1981). "? Pg C yr^{-1}" indicates a still-unknown flux.

erosion related to human activity constantly provides nonweathered minerals, may be four times higher than that under natural conditions (W. Li et al., 2022).

2.1. Physical Erosion Alters Organic Carbon Cycling Processes

SOC stabilization mechanisms control organic carbon detachability and decomposition during erosion phases to a large extent (Xiao et al., 2018). Organic carbon exists in various forms and is categorized into different pools based on its likelihood of decomposition or oxidation. These pools include active pools (such as root exudates and rapidly decomposing components of fresh litter like phenols and carbohydrates), slow pools (with turnover times ranging from years to hundreds of years, including substances like cellulose and hemicellulose), and passive pools (materials that persist in soils for thousands of years, such as lipids and lignin-derived compounds) (Trumbore, 1997). SOC stabilization relies mainly on (a) chemical linkages between SOC and soil mineral surfaces and (b) physical protection mechanisms of SOC, such as the packing of aggregates or burial under a mass of soil (Doetterl et al., 2016). The rate of SOC decomposition by microbes is influenced by chemical composition and physical protection mechanisms, as well as environmental factors such as soil temperature, humidity, and aeration (Chen et al., 2015; Lehmann & Kleber, 2015). Active and fresh organic matter in the soil carbon pool, typically found in the topsoil layer, is easily transported and decomposed during heavy rainfall events, making it crucial in erosion systems (Deumlich et al., 2018; Lin et al., 2019; Yue et al., 2012).

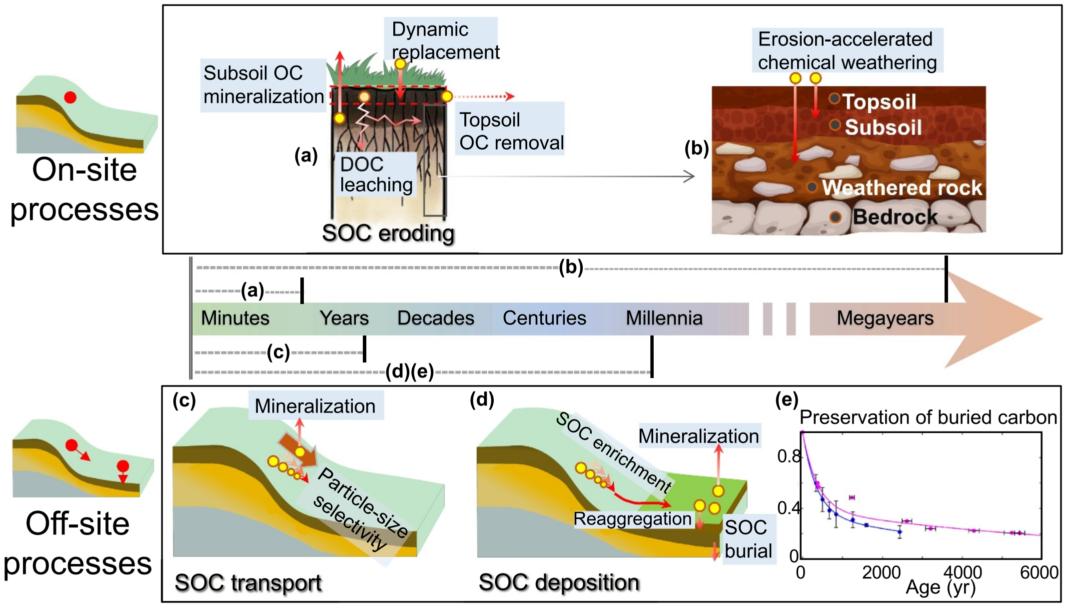

Figure 3. (a, b) Effective duration of on-site and (c, d) off-site processes induced by physical and chemical erosion. The colored arrow in the middle of the figure is a timeline, and the letters refer to the panels above and below it. Downward red arrows indicate carbon uptake; upward red arrows indicate carbon release; rightward (near-) horizontal red arrows indicate the lateral transport of eroded organic carbon. (e) Preservation ratio of buried soil carbon over time. The data shown with blue dots are from Van Oost et al. (2012) and refer to carbon burial in colluvial soils. The data shown with pink dots, referring to arable land at slope bottoms, are derived from Miao (2008). The preservation ratio is estimated through the carbon content in a typical soil depth divided by the carbon content in the surface soil layer. The age in (e) corresponds to the millennial timescale in (d).

SOC undergoes complex on-site dynamics at eroded areas on slopes (Figure 3a). First, soil erosion can lead to a significant loss of SOC (Dlugoß et al., 2012), and the affected soil is prone to progressive degradation. The organic layer and topsoil are often loose and have low bulk density due to the high levels of organic carbon (Lorenz & Lal, 2018), so surface organic carbon is preferentially carried away by surface runoff (Causarano et al., 2008; Stallard, 1998). SOC removal directly reduces carbon storage in eroded slopes, so soil productivity may be affected (Polyakov & Lal, 2008). Following surface soil removal, changes in subsoil moisture, temperature, and oxygen content on slopes are likely more favorable for soil formation once the carbon-depleted subsoil zone is exposed to the air (Alhassan et al., 2021; Bajracharya et al., 2000; B. Liu et al., 2020). Elevated soil temperature and moisture can accelerate the mineralization of deep SOC (carbon release) and promote the fragmentation and weathering of deep, less-weathered mineral soils (carbon uptake).

In general, the surface SOC content (typically ranging from 10 to 75 kg C m^{-3}) is considerably higher than the dissolved organic carbon (DOC) content (1.7–88.3 g C m^{-3}), so the majority of studies have not considered DOC changes caused by erosion (Van Oost et al., 2012; Yue et al., 2016). Nevertheless, DOC, the most labile component in soil, is an important intermediate variable in SOC formation and mineralization. The content of DOC is determined by organic carbon's solubility and particle size (Langeveld et al., 2020). DOC is typically lost through leaching into deeper soil layers or flowing into river systems through surface and subsurface runoff (Gommet et al., 2022; L. Wang et al., 2019). External environmental factors, such as rainfall intensity and runoff volume, can facilitate or hinder the vertical migration of DOC (Deirmendjian et al., 2019; Ma et al., 2014). The vertical migration of DOC intensifies with increasing rainfall intensity (Dlugoß et al., 2012). Erosion could alter the vertical distribution of DOC, and the content of DOC is higher at eroding sites than at depositional sites (X. Wang et al., 2013, 2014).

While soil erosion removes a portion of the topsoil's organic carbon, it may have only a limited impact on the continuous input of new organic carbon from plant photosynthesis into the soil (Feng et al., 2018; Van Oost et al., 2012). “Dynamic replacement” (Harden et al., 1999) at an eroded area refers to replacing lost soil carbon with the latest organic carbon from vegetation residues, and it reflects the capacity for SOC recovery at eroded sites. Since the carbon content of deeper soil is often lower than that in the humus layer and topsoil (Sinoga et al., 2012), subsoil substances are unsaturated with organic carbon and have available reaction sites, which can stabilize and transform newly generated organic carbon from plant residues in the soil (Harden et al., 1999; Kirkels et al., 2014). In other words, erosion promotes organic soil regeneration by removing part of the carbon-rich surface soil, leaving plants growing in relatively carbon-poor, less-weathered conditions, where the remaining soil still provides sufficient nutrients—such as nitrogen, phosphorus, potassium, and sulfur—for healthy plant growth. Therefore, eroded areas can continuously accumulate and supply organic carbon to deposition sites, and dynamic replacement maintains soil productivity, along with effective carbon burial in deposition areas, ultimately creating a persistent sink for atmospheric carbon (Berhe et al., 2014). The replacement rate expresses the percentage of eroded carbon replaced by new carbon, generally ranging from 19% to 100% (Quinton et al., 2006; Van Oost et al., 2007). However, it is important to further investigate and quantify how carbon sequestration potential depends on erosion intensity. In relatively fertile, slowly eroding, and poorly mineralized soils, plant growth is not significantly hindered by erosion, allowing carbon fixation to offset much of the carbon lost through erosion (Feng et al., 2018). In contrast, severe erosion events significantly reduce vegetation productivity, turning these areas into typical carbon source zones (Deng et al., 2019).

2.2. Physical Erosion Accelerates Chemical Erosion

Several important types of chemical erosion (weathering)—including carbonate weathering, silicate weathering, the weathering of carbonate minerals driven by sulfuric acid, and the oxidation of rock organic carbon—significantly influence carbon fluxes between the atmosphere and land on both short and long timescales. The first two types are crucial carbon sink processes, while the latter two, which primarily occur in glacier regions, act as carbon sources to the atmosphere. Carbonate and silicate weathering are the most prevalent forms of chemical weathering, with carbonate and silicate rock types covering 53% of the global land area, contributing to a carbon sink rate of 0.222–0.317 Pg C yr^{-1} (Xiong et al., 2022). Sulfuric acid, produced from sulfide oxidation, can dissolve carbonate minerals in karst terrains and fragmented glacial sediments, releasing CO_2 into the atmosphere (Martin, 2017). In a precipitation-dominated, glacierized basin in the central Himalaya, sulfuric-acid-driven carbonate weathering contributes to two-thirds of the dissolved load in the meltwater, and this process particularly dominates in the middle and later stages of glacier melting (Sundriyal et al., 2024). The effect of thermokarst in glacier regions on climate (i.e., carbon sink or release) depends on the mineral composition of permafrost soils (Zolkos et al., 2018). In addition, the annual carbon flux to the oceans from the weathering of sedimentary rock is about 0.043 Pg C, and the carbon release from fossil organic carbon in sedimentary rocks can be substantial when erosion is relatively strong (Copard et al., 2007). It has been demonstrated that rock organic carbon oxidation driven by glaciation led to an increase in CO_2 emissions in the Mackenzie River Basin in Canada and in the mountain watersheds of New Zealand (Horan et al., 2017, 2019). The oxidation of organic matter in sedimentary rocks could have contributed to an increase in atmospheric CO_2 of approximately 30–60 ppm during the last deglaciation (Blattmann, 2022).

Soil removal mixes the minerals of different soil layers and thus increases the probability of minerals weathering in the deep, less-weathered soil layer. Typical weathering processes, such as silicate mineral hydrolysis and sedimentary rock hydration, can extend to tens of meters below the soil (Buss et al., 2008). Fractured soil layers allow gases (O_2 and CO_2) and water to seep through, speeding up soil parent material disintegration and chemical weathering (Molnar et al., 2007). Micro-fracturing has been reported to cause both physical and chemical erosion by increasing gas influx (Gu et al., 2020). Compared with sheet and rill erosion, physical and chemical weathering processes respond more significantly to gully erosion and gravity erosion, such as landslides and debris slides (Soulet et al., 2018). Gully erosion, including rainstorm-driven and thermo-erosion gullies in permafrost landscapes, is a major cause of land degradation (Pal et al., 2022), and it accelerates degradation of the regolith in both the sides and bottoms of the gullies and enhances the exposure and weathering of weathered material. Chemical weathering can also indirectly promote gully development (Chakrabortty et al., 2022). During gully erosion of sedimentary rocks, the increased exposure of rock organic carbon may result in higher CO_2 emissions to the atmosphere (Copard et al., 2007). Most gully erosion development induced by overland flow scouring can promote slope failure and landslides (Lalitha et al., 2021). Landslides (or collapses), a common type of erosion in mountainous areas, are often triggered by earthquakes, successive storms, and human disturbance, and they can leave distinctive hillslope scars and masses of soil clastic materials. Freshly exposed, less-weathered landslide materials provide the reaction mass for chemical weathering, and appropriate temperature and humidity can create favorable conditions for chemical weathering (Moquet et al., 2021; C. H. Wang et al., 2021). The residence time of landslide material determines local carbon exchange fluxes, and it mostly depends on the transport capacity of surface runoff (Croissant et al., 2019). Thermo-erosion gullies in permafrost landscapes can accelerate the mineral weathering process and may release carbon after permafrost collapse (Turetsky et al., 2019, 2020).

Erosion provides the source material and creates better environmental conditions for weathering, establishing a clear, positive correlations among watershed erosion rate, oxidative weathering rate (carbon release), chemical weathering rate (carbon uptake), and riverine suspended sediment yield (Galy et al., 2015; Hilton & West, 2020; Soulet et al., 2018). Usually, a higher erosion rate intensifies the interaction between soil minerals and the atmosphere (Moore et al., 2013; Soulet et al., 2018). For example, nearly 40% of chemical denudation occurs in the steepest 10% of slopes in the Himalayan regions and other high-mountain areas (Larsen et al., 2014). However, when the erosion rate is extremely high, the chemical weathering rate may be influenced more by surrounding conditions (e.g., CO_2 concentration, temperature) than erosion intensity. This is because a large portion of materials have already met the basic requirement for chemical weathering (Hilley et al., 2010). It was found that chemical weathering of minerals produces maximum CO_2 drawdown at erosion rates of ~0.07 mm yr^{-1} (Bufe et al., 2024). Erlanger et al. (2021) found accelerated soil erosion to be the major driver of mineral weathering in an Italian mountain system of mixed silicate-carbonate rocks. W. Li et al. (2022) reported the carbon fluxes of carbonate weathering in the Pearl River in China during 1957–1980 were 4.6 times higher than those during 1893–1957. Global increased silicate weathering rates resulting from accelerate erosion have contributed to a carbon sink in the past thousands of years (Hilton et al., 2015). Moreover, silicate weathering rates in the mountains would increase by 0.4%–0.7% if annual surface runoff increased by 1% (Hilton, 2017), so amplified climate change and the enhanced erosion and sediment transport in high-mountain areas (D. Li, Lu, et al., 2021; Miao et al., 2024) will likely accelerate global chemical weathering processes.

3. Off-Site Carbon Dynamics Induced by Physical and Chemical Erosion

Off-site processes include the phases of transport, sedimentation, and burial (Figures 3c and 3d). Accelerated erosion induces significant downstream landscape transformations—the amount of sediment transported by streams and rivers increases substantially, and colluvial areas become more widespread (Cendrero et al., 2022; Tarolli & Sofia, 2016). As a result, a considerable amount of the deposited organic carbon is gradually buried and kept away from short-term cycling and then becomes a nonnegligible part of geological carbon cycling.

3.1. Depositional and Burial Characteristics of Eroded Materials

Deposition and burial of carbon on land typically refer to the behavior of sediment organic carbon (the behavior of organic carbon transformed from dissolved inorganic carbon (DIC) in water bodies is discussed in Section 3.2). Predicting sediment deposition trajectories is one of the most challenging aspects of quantifying off-site carbon fluxes caused by erosion (Hu & Kuhn, 2014; Kirkels et al., 2014). However, a substantial part of eroded soil tends to be deposited along the surface runoff pathway, and soil deposition in areas without human activities is largely predictable over the course of several years. It was reported that more than half of eroded SOC is deposited within the local watershed, while 10%–30% of SOC is transported to distant downstream or coastal areas (Lal, 2020; Stallard, 1998; Walling & Webb, 1996).

Previous observations have confirmed SOC enrichment occurs at depositional sites—where sediment carbon content is higher compared to the original sites (Schiettecatte et al., 2008a)—because of preferential transport of fine-grained loose soil components, particularly SOC, by surface runoff. Fiener et al. (2015) highlighted that SOC enrichment during interrill erosion is remarkable in small erosion events and overlooking the enrichment may lead to an inaccurate estimation of vertical carbon release. H. Zhang et al. (2014) reported SOC enrichment ratios (ER, the ratio of the SOC concentration of the eroded sediment to that in the original soil) in migrated sediment ranging from 1.3 (25th percentile) to 2.6 (75th percentile), with a median value of 1.8. ER is probably less than unity in some specific cases (Hu & Kuhn, 2014; Hu et al., 2013). ER is correlated with various factors, such as erosion type, sediment delivery ratio (negative correlation) (Schiettecatte et al., 2008b), sediment carbon and nitrogen content (positive correlations), and rainfall intensity and duration (negative correlations) (Schiettecatte et al., 2008b; Strickland et al., 2005). The negative correlations with rainfall intensity and duration are due to particle sorting; that is, larger, carbon-poor particles are transported together with fine, carbon-rich soil when heavier rainstorms occur. In contrast, most fine, carbon-rich soil is transported in less-intense rain events (Nie et al., 2015).

The depositional environment, characterized by high spatial heterogeneity, is a crucial external factor influencing the stabilization of deposited organic carbon. In low-oxygen depositional sites, such as anaerobic areas of wetlands and lake bottoms, deposited SOC is minimally oxidized by microbes, resulting in high carbon sequestration (Lal, 2019; Sasmito et al., 2020; Z. Wang et al., 2014). A 5-year field experiment on terraced lands reported an 8% decrease in the mean carbon release from disturbed soil compared to its original state, which was attributed to changes in the soil's biochemical properties (W. Li et al., 2020). Note that in low-oxygen settings, flood conditions and high levels of SOC stimulate the growth of methanogens, leading to the conversion of SOC into CH_4, which acts as an atmospheric source (Lal, 2020; Worrall et al., 2016).

Most depositional environments, such as widespread colluvium and floodplains, however, offer well-ventilated (aerobic) conditions for microbial activities, resulting in continuous carbon emission (Lal, 2019). The influx of fresh materials at deposition sites not only increases the storage of organic carbon but also changes the population structure of soil microbes and boosts the activity of the soil microbial community, leading to increased mineralization (Kuzyakov & Bol, 2006). Experimental data showed that the cumulative release of CO_2 increased by 27% at depositional zones compared to noneroded areas (Mariappan et al., 2022; X. Wang et al., 2014).

Organic carbon burial is a long-term process following the deposition of organic carbon, and it plays a crucial role in carbon sequestration, warranting greater attention. SOC burial predominantly occurs in low-lying areas (such as slope toes, reservoirs, lakes, floodplains, and oceans). Initially buried organic carbon becomes gradually overlain by newly deposited materials over time, leading to slow carbon sequestration. Figure 3e illustrates that the total amount of preserved buried SOC exhibits exponential decay over time as burial depth increases, particularly over a timescale of thousands of years in well-ventilated environments (Van Oost et al., 2012). In other words, carbon sequestration at depositional areas over short timescales is inconclusive due to the time-dependent effect on the mineralization of deposited SOC. Observations show that ∼23% of sedimentary carbon is mineralized to CO_2 (77% preserved) within the first 100 days (Van Hemelryck et al., 2010), 50% is preserved after 500 years (Van Oost et al., 2012), and only 17% remains after 1,000 years (Z. Wang et al., 2014). Given that persistent decomposition over the long term limits carbon sink potential in colluvial soil, it is essential to accurately define the initial and terminal conditions of erosion in quantitative research. Carbon burial efficiency can be used to show this burial characteristic, which is defined as the ratio of the current carbon content of the topsoil to the original value found in the subsoil immediately beneath the topsoil at depositional sites (Doetterl et al., 2016). The efficiency eventually stabilizes at a constant value and is positively correlated with the sedimentation rate (Z. Wang et al., 2015; Worrall et al., 2016). However, the eroded and deposited carbon content from topsoil to deep layers rarely follows a strictly monotonic decreasing trend.

In addition, the reformation of soil aggregates (reaggregation) can improve soil carbon sequestration and boost soil productivity, making it an important indirect carbon sink (Blanco-Canqui & Lal, 2004). The dispersed clay from the breakdown of aggregates may interact with organic carbon and cations to promote the formation of new soil aggregates after compacting and altering soil structural units (Lal, 2022). The reformatted aggregates can stabilize buried organic carbon through encapsulation (Schomburg et al., 2019; Steger et al., 2019; Wade et al., 2020). With the addition of deposited nutrients such as nitrogen and phosphorus, which support plant growth, deposition sites often demonstrate higher plant productivity compared to non-depositional areas.

Overall, the amount of global buried carbon is estimated to be 0.6–1.6 Pg C yr^{-1} (Stallard, 1998). The carbon burial rate of check dams—important engineering structures designed to intercept water and sediment to create cultivated land—in channels over the past 50 years has been investigated, and a value of 1,773 t C yr^{-1} was reported for an agricultural watershed of 187 km^2 located on the Loess Plateau of China (Zeng, Fang, & Shi, 2020). The capacity of check dams to store carbon relies strongly on intercepted sediment volume at watershed scales (Yao et al., 2022). In addition, widely distributed floodplains have a high capacity to store sediment organic carbon. For example, SOC is stored at a rate of 266 t C ha^{-1} within the upper 2 m of floodplains in California's Central Valley, with the majority of carbon being buried at depths of approximately 0.8 m (Steger et al., 2019). Nevertheless, over 50% of carbon stored in the top layers of floodplains is expected to be lost within a century (Omengo et al., 2016). Certainly, lakes and reservoirs are the most important continental environments for efficient carbon burial. Over the past 8,000 years, SOC burial in lakes and reservoirs worldwide has amounted to approximately 41.25 Pg C (Z. Wang et al., 2017). Another study shows a mean burial rate of 0.06–0.25 Pg C yr^{-1}

for global lakes and reservoirs during the last 150 years (Mendonça et al., 2017). Temperature positively affects the mineralization of organic carbon in lake sediments, and future temperature rise could lead to a 4%–27% reduction in organic carbon burial in boreal lakes (Gudasz et al., 2010).

3.2. Transport Characteristics of Eroded Materials

The transportation of eroded materials delivers DOC, particulate organic carbon (POC), DIC, and PIC in sediment runoff to new locations. During the transport of sediment OC, particle size sorting and enhanced decomposition of OC are dominant processes in most cases (Figure 3c). However, the sorting characteristics are common in sheet and interrill erosion, but not typically observed in rill erosion (Schiettecatte et al., 2008a), gully erosion (J. Zhang et al., 2006), and landslides (Dialynas et al., 2016). The magnitude of ER is related to the sorting characteristics (Polyakov & Lal, 2004). Furthermore, the active organic carbon in the surfaces of soil particles and aggregates (Issa et al., 2006; Zhai et al., 2019) is easily exposed and decomposed into CO_2 during transport (Feng et al., 2018). The decomposition of transported SOC is mostly affected by the actual aggregate size (Hu & Kuhn, 2014; T. Liu et al., 2023). Nevertheless, estimating mineralization of mobilized soil remains a crucial and challenging task (Hu & Kuhn, 2014).

The mineralization of in-transit SOC varies in different experimental environments and geographic positions. Several geoscientists have estimated that 20%–30% of migrated carbon is mineralized (Jacinthe et al., 2002; Lal, 2005; S. V. Smith et al., 2005), while up to ∼56% of eroded carbon may be oxidized during transport (Z. Wang et al., 2014); others have reported minimal carbon loss during transport as evidenced by experiments showing only a slight decrease in the concentration of recently deposited SOC (Galy et al., 2007; Z. Wang et al., 2017). Experiments conducted on sloping farmland with red soil showed that the SOC content in runoff and sediment was higher than in the original soil but SOC content in depositional areas of the basin was lower than in the eroded soil (M. Zhang & Liu, 2009); this demonstrates that carbon-rich soil migrated preferentially and that the migrated SOC underwent strong decomposition. Temperature is also an important factor affecting the rate of SOC mineralization, as demonstrated by research showing that turnover time is much longer—by more than an order of magnitude—in a cold climate than in a warm climate (Koven et al., 2017). Overall, researchers tend to concur that accelerated mineralization of SOC during transport has been identified as a source of atmospheric carbon (de Nijs & Cammeraat, 2020).

Identifying the sources of DIC and PIC caused by chemical weathering in rivers (Figure 4) is crucial for revealing the relationships among riverine material fluxes, slope erosion, and chemical weathering rates. Generally, atmospheric CO_2 does not directly enter static freshwater rivers or streams, except for through photosynthesis by aquatic vegetation, because the carbon concentration in these waters is already supersaturated relative to atmospheric carbon (disregarding carbon exchange at the water–air interface) (Billett et al., 2007; S. Liu et al., 2016; Raymond et al., 2013). Thus, the sources of carbon—POC, DOC, DIC, and PIC—in inland waters are mostly derived from terrestrial ecosystems (Figure 4). Riverine POC originates from soil erosion in the biosphere (abbreviated POC_{bio}), from physical erosion of the lithosphere ($POC_{petro\text{-}direct}$), and from photosynthetic assimilation by aquatic autotrophs ($POC_{petro\text{-}indirect}$), and none of these components can be ignored (Einsele et al., 2001; Z. Liu et al., 2017; Tao et al., 2004; Zeng, Fang, Shi, Lu, & Wang, 2020). To explore the relationship between the land erosion rate and riverine POC, it is necessary to distinguish riverine lithogenic carbon (POC_{petro}, which equals $POC_{petro\text{-}direct}$ plus $POC_{petro\text{-}indirect}$) from biospheric carbon (POC_{bio}) through radiocarbon activities (reported as fraction modern [*Fm*] or $\Delta^{14}C$) (Hilton & West, 2020). Usually, the radiocarbon activity is low for POC_{petro} but high for POC_{bio} (Blair et al., 2003). It is estimated that the global POC_{bio} flux is 0.11–0.23 Pg C yr^{-1} and the POC_{petro} flux is 0.018–0.10 Pg C yr^{-1} (Galy et al., 2015). In addition, positive relationships have been found between POC_{bio} yield and suspended sediment yield, as well as between POC_{petro} yield and suspended sediment yield (Galy et al., 2015). The total POC flux ($POC_{bio} + POC_{petro}$) is largely controlled by the erosion rate, sediment yield (Hilton et al., 2012), and the sediment carbon content (Stallard, 1998).

Some DOC is produced from surface erosion and lateral migration through overland flow, and the remaining DOC originates from soluble SOC that enters river network systems through leaching and delivery (Figure 4) (Billett et al., 2007). DOC is labile in aquatic environments, and new DOC can be produced from POC trapped by lakes and reservoirs (Ittekkot et al., 1986), and DOC is easily decomposed into CO_2 (S. Liu et al., 2016). Previous studies have revealed strong relationship between DOC and several critical factors in catchments, such as river discharge (closely correlated), labile organic carbon content, soil respiration rate, soil erosion rate, average slope,

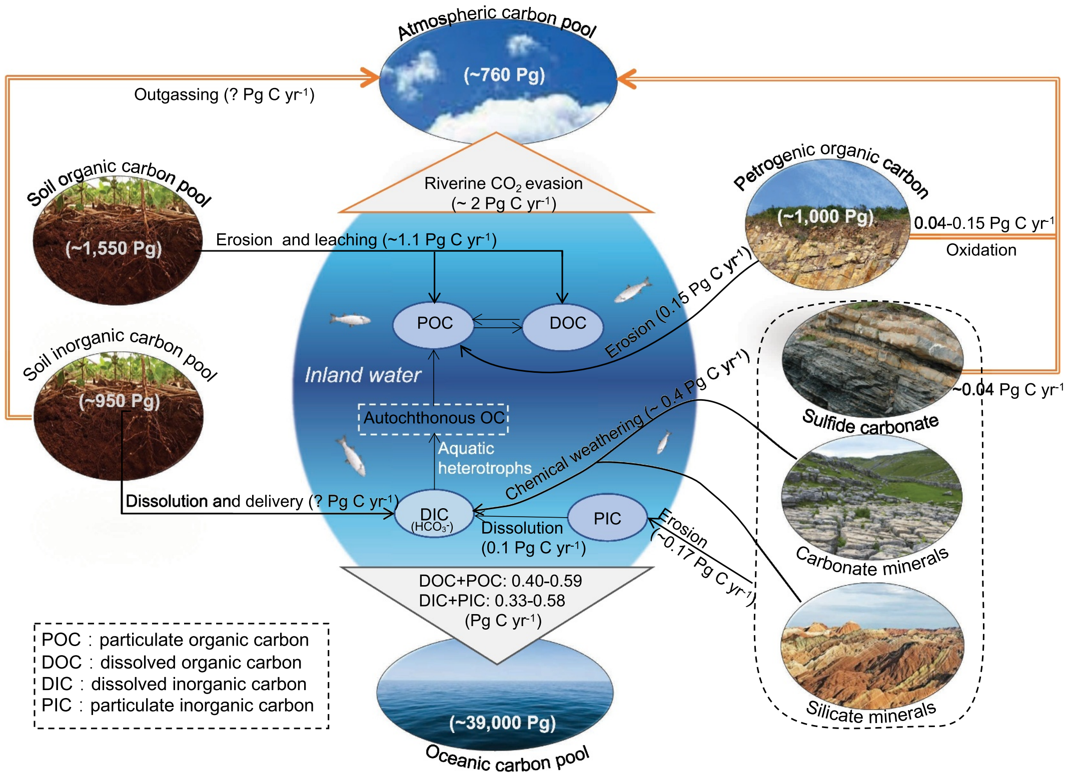

Figure 4. The terrestrial sources and major transformations of dissolved organic carbon, particulate organic carbon, dissolved inorganic carbon (DIC), and PIC in inland waters; the fluxes are in units of Pg C yr^{-1}. The storages of soil organic carbon, inorganic carbon, and petrogenic organic carbon refer to the amount contained within the top meter of soil. PIC, which generally does not involve terrestrial–atmospheric carbon exchange, is not discussed in this review. "? Pg" indicates a still-unknown flux. Autochthonous OC, as opposed to allochthonous OC, refers to organic carbon produced through photosynthesis by aquatic autotrophs absorbing DIC originating from the terrestrial lithosphere, rather than from the atmosphere. The data used in this figure are from Hemingway et al. (2018), Hilton and West (2020), Lal (2003), S. Liu et al. (2016), and Regnier et al. (2022).

and anthropogenic activities (e.g., industrial wastewater release and basin population density) (M. Li et al., 2017; D. Liu et al., 2020; Ludwig & Probst, 1999). For example, M. Li et al. (2017) proposed the following simple empirical relationship for DOC in global rivers:

$$F_{\mathrm{DOC}} = 0.0081 + 0.0044 \times Q + 0.050 \times \mathrm{SOC}, \quad r^2 = 0.95,\ n = 109 \tag{1}$$

where F_{DOC} (Tg C yr^{-1}) is DOC flux in rivers, Q ($km^3\ yr^{-1}$) is discharge, SOC (Pg C) is total soil organic carbon in a basin, r^2 is the coefficient of determination, and n is the total number of involved basins. But in Ludwig et al. (1996), the average steepness of basins (Slope, unit of radians) is another important empirical factor determining DOC (t $km^{-2}\ yr^{-1}$):

$$F_{\mathrm{DOC}} = 0.004 \times Q - 8.76 \times \mathrm{Slope} + 0.095 \times \mathrm{SOC},\ r^2 = 0.90,\ n = 29 \tag{2}$$

where Q (mm) is runoff depth in a specific basin and SOC (kg m^{-3}) is organic carbon content in the given basin.

DIC, predominantly as bicarbonate (HCO_3^-), is primarily generated through the chemical weathering of carbonate and silicate minerals, as well as from gaseous carbon dissolution and delivery in soil (where CO_2 in soil is produced by plant root respiration, microbial decomposition, and direct exchange between the soil and the atmosphere). A close relationship between DIC and weathering rate was proposed by M. Li et al. (2017):

$$F_{\mathrm{DIC}} = 0.50 + 2.47 \times f_{\mathrm{CO2}} - 0.0038 \times Q,\ \ r^2 = 0.77,\ n = 111 \tag{3}$$

Table 1
The Global Carbon Flux From Continents to Oceans in Published Studies

	Specific flux (Pg C yr^{-1})						
Total flux	OC	DOC	POC	IC	DIC	PIC	References
0.95	0.37	0.20	0.17	0.55	0.38	0.17	Meybeck (1982, 1993)
0.71	0.38	0.21	0.17	/	~0.33	/	Ludwig et al. (1996, 1998)
0.82	0.53	0.23	0.30	/	0.29	/	Stallard (1998)
0.80–0.9	0.40	0.20	0.20	/	0.40–0.50	/	Aufdenkampe et al. (2011)
1.06	0.48	0.24	0.24	0.58	0.41	0.17	M. Li et al. (2017)
0.85–0.90	~0.40	/	/	0.45–0.50	/	/	Drake et al. (2018)
<0.83	<0.45	0.21	<0.24	/	0.38	/	Syvitski et al. (2022)

Note. "/" means no figures reported in the cited research. The unit of all figures is Pg C yr^{-1}. OC, organic carbon; IC, inorganic carbon; DOC, dissolved organic carbon; POC, particulate organic carbon; DIC, dissolved inorganic carbon; PIC, particulate inorganic carbon.

where F_{DIC} (Tg C yr^{-1}) is DIC flux in rivers, Q (km^3 yr^{-1}) is discharge, f_{CO2} (Tg C yr^{-1}) is the gross CO_2 consumption by mineral weathering in a basin, r^2 is the coefficient of determination, and n is the total number of involved basins.

Chemical weathering supplies autotrophs with bicarbonate to synthesize organic carbon, and this process is often the most overlooked aspect of terrestrial inorganic carbon output (Billett et al., 2007; Raymond et al., 2004). In the estuary of the Mississippi River, the organic matter is dominated by autochthonous carbon sources (60%–83%) (Waterson & Canuel, 2008). In carbonate-dominated rivers, a large portion of DIC originates from the lithosphere (radiocarbon-depleted) (Z. Liu et al., 2017); autotrophs can utilize this DIC to grow up, rather than allowing the DIC to combine with Ca^{2+} to release carbon ($Ca^{2+} + 2HCO_3^{-} = CaCO_3 + CO_2\uparrow + H_2O$). Therefore, the contribution of carbonate-fixed carbon has often been underestimated in previous assessments. Z. Liu et al. (2018) reported a carbon sink of approximately 0.5 Pg C yr^{-1} caused by carbonate mineral weathering coupled with aquatic photosynthesis for the world's continents. So, aquatic burial of carbon originating from terrestrial minerals represents a stable sink of atmospheric carbon (Z. Liu, 2012).

3.3. Organic and Inorganic Carbon Fluxes Transported From Eroding Areas to Oceans

Material fluxes in river systems provide abundant information about physical erosion and chemical weathering products (L. Yang et al., 2022), such as water discharge, various types of carbon, and the concentrations of ions such as Ca^{2+}, Mg^{2+}, Na^{+}, K^{+}, SO_4^{2-}, and Cl^{-}. Oceans receive carbon at a rate of 0.71–1.06 Pg C yr^{-1} (Table 1). The discrepancies in calculations of carbon fluxes in coastal oceans mostly stem from the differences in estimation methods, the limited data sources, and the specific research periods considered (Drake et al., 2018). For example, predicting POC fluxes in estuaries in most studies begins with the sediment load and eroded soil amounts (Ji et al., 2016; M. Li et al., 2019; Ludwig et al., 1998); but the global eroded soil amount estimated by multiple methods has large uncertainties, basically ranging from 35 to 200 Pg yr^{-1} (Borrelli et al., 2017; Ito, 2007; Lal, 2003; S. V. Smith et al., 2001; D. Yang et al., 2003).

Most terrestrial carbon eventually settles to the ocean floor, where it becomes buried and sequestered. Even though ~1.0 Pg C yr^{-1} is transported to estuaries, carbon burial in the open ocean is ~0.75 Pg C yr^{-1} (Drake et al., 2018). In the coastal ocean of New Zealand, 80% of SOC is deeply buried in the open ocean, with a burial rate of 0.0031 Pg C yr^{-1} (Dymond, 2010). The burial rate of SOC in the marginal seas of eastern China is approximately 0.007 Pg C yr^{-1} (Zhao et al., 2021). The burial efficiency in the coastal Bohai Sea reaches 43%, but it is much lower in the Yellow Sea (11%) and East China Sea (16%) (Zhao et al., 2021). The land organic carbon buried in oceans is approximately 0.17–0.20 Pg C yr^{-1} (Galy et al., 2007; R. W. Smith et al., 2015).

4. Quantification of Erosion-Induced Vertical Carbon Fluxes

4.1. Framework for Quantification of Physical-Erosion-Induced Vertical Carbon Fluxes

Even though efforts have been made to quantify erosion-induced vertical carbon fluxes (Lal, 2019; Quine & Van Oost, 2007; Yue et al., 2016), accurately quantifying relative fluxes is challenging, particularly at large scales, due to the complex interplay of multiple processes and factors influencing carbon dynamics (see Sections 2 and 3). This section proposes new approaches for quantifying vertical fluxes over short timescales (ranging from minutes to hundreds of years) and long timescales (millennia and beyond). The conceptual frameworks reflect current state-of-the-art knowledge to provide numerical outputs. Specifically, for short timescales, the focus is on net changes in vertical carbon flux during on-site and off-site processes, compared to noneroded areas or the land surface prior to erosion (Harden et al., 1999; Van Oost et al., 2012; Yue et al., 2016). Producing precise estimates is possible, but it remains challenging. The approach for quantification of on-site processes is presented in Figure 5 and can be expressed as follows:

$$F_{\text{on-site}} = ST_{E,T} - ST_{NE,T} - \sum_{i=1}^{T} v_{\text{ero}} \times c_{\text{top}(i)} \tag{4}$$

where $F_{\text{on-site}}$ is the erosion-induced vertical carbon flux after T years from the initial state at eroding sites, $F_{\text{on-site}} > 0$ means carbon uptake; ST represents carbon storage, with the subscripts E and NE referring to eroded areas and noneroded areas (i.e., reference sites), respectively; v_{ero} is the annual mean erosion rate; and c_{top} is the carbon content of topsoil (c_{top} will change over time). Similarly, the vertical carbon flux at deposition sites (F_{dep}) can be expressed as follows:

$$F_{\text{dep}} = ST_{D,T} - ST_{ND,T} - \sum_{i=1}^{T} C_{\text{dep}(i)} \tag{5}$$

where the subscript D represents deposition areas and ND represents nondeposition areas (i.e., reference sites) and C_{dep} is the deposited organic carbon in the ith year at deposition sites. $F_{\text{dep}} > 0$ means carbon uptake.

Although this quantification framework is straightforward, it relies heavily on precise experimental data, which is often hard to obtain. For example, accurately determining SOC storage at reference sites—which are difficult to identify—and at erosion and deposition sites is a complex task. In addition, measuring soil loss rates using experimental methods like ^{137}Cs is expensive and labor intensive.

Further, Equation 3 can be reshaped:

$$F_{\text{on-site}} = \left(ST_{E,T} - \sum_{i=1}^{T} v_{\text{ero}} \times c_{\text{top}(i)} - ST_0 \right) - (ST_{NE,T} - ST_0) \tag{6}$$

where ST_0 is the original carbon storage, the first term in parentheses on the right-hand side of Equation 6 is the net vertical carbon flux at eroded sites, and the second term in parentheses is the net vertical carbon flux at noneroded sites (reference sites). Therefore, the framework for quantifying physical-erosion-induced carbon flux at on-site positions can be expressed by the changes in vertical carbon flux (Figure 5):

$$F_{\text{on-site}} = \Delta F_{\text{net}} = F_{\text{net},E} - F_{\text{net},NE} = \sum_{i=1}^{T} (F_{\text{upt},E} - F_{\text{min},E})_i - \sum_{i=1}^{T} (F_{\text{upt},NE} - F_{\text{min},NE})_i \tag{7}$$

where the subscripts "upt" and "min" represent carbon uptake and mineralization processes, respectively. Similarly,

$$F_{\text{off-site}} = \Delta F_{\text{net}} - F_{\text{tran}} = F_{\text{net},D} - F_{\text{net},ND} - F_{\text{tran}} = \sum_{i=1}^{T} (F_{\text{upt},D} - F_{\text{min},D})_i - \sum_{i=1}^{T} \left(F_{\text{upt},ND} - F_{\text{min},ND}\right)_i - F_{\text{tran}} \tag{8}$$

and

$$F_{\text{short}} = F_{\text{on-site}} + F_{\text{off-site}} \tag{9}$$

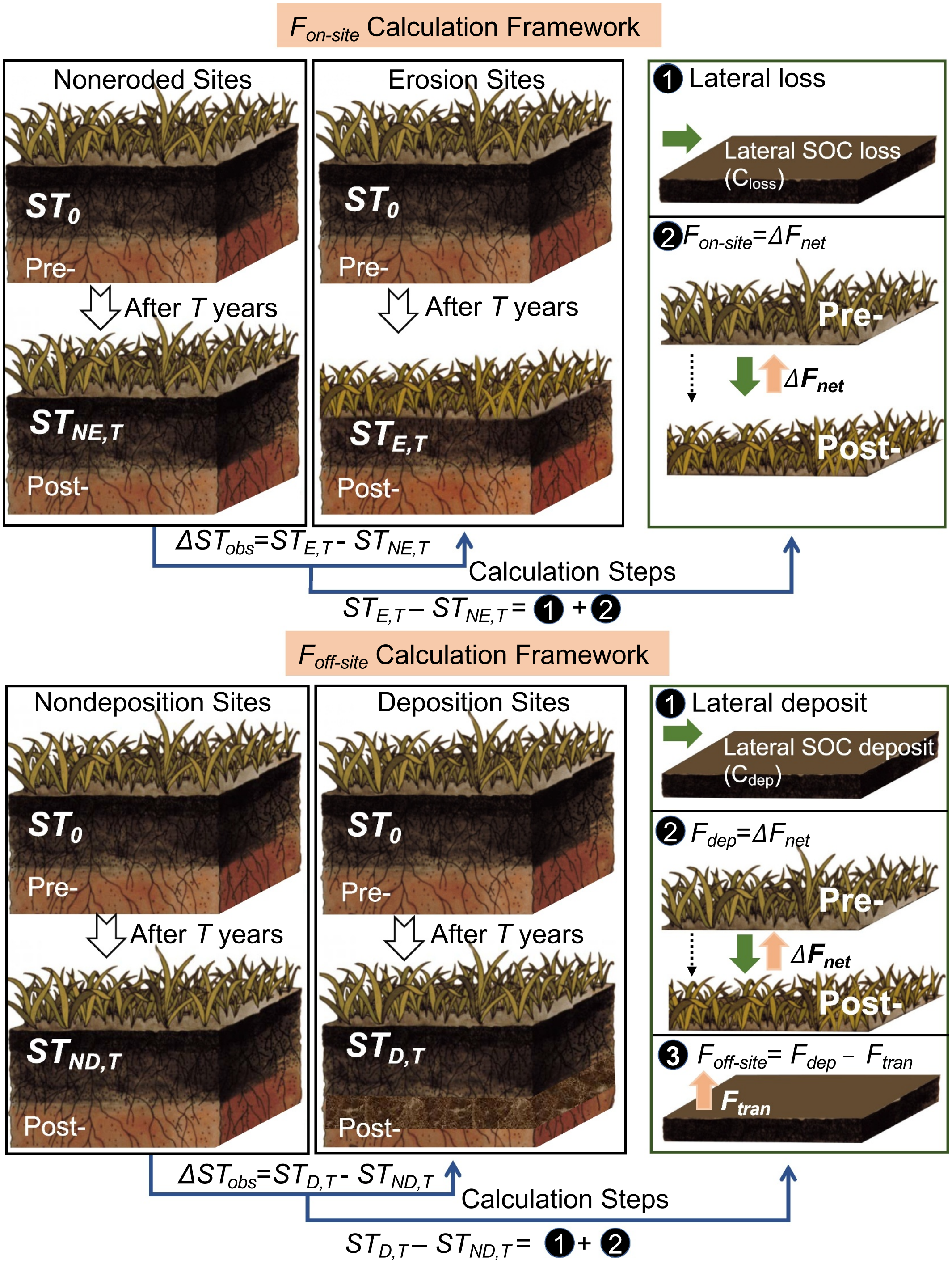

Figure 5. The framework for quantifying the physical-erosion-induced carbon flux for short timescales (the upper part corresponds to $F_{\text{on-site}}$ and the lower part to $F_{\text{off-site}}$). The change in soil organic carbon storage (ΔST_{obs}) at eroded (or depositional) areas, compared with noneroded (or nondepositional) areas, consists of two components: lateral SOC loss (or deposition) and vertical carbon flux change (i.e., ΔF_{net}). F_{tran} is the erosion-induced carbon release during sediment transport on slopes and in rivers. Noneroded and nondepositional areas can be regarded as reference sites and also as equivalent to the previous state ("Pre-") before erosion and deposition occurrence, while "Post-" refers to the state after soil erosion and deposition.

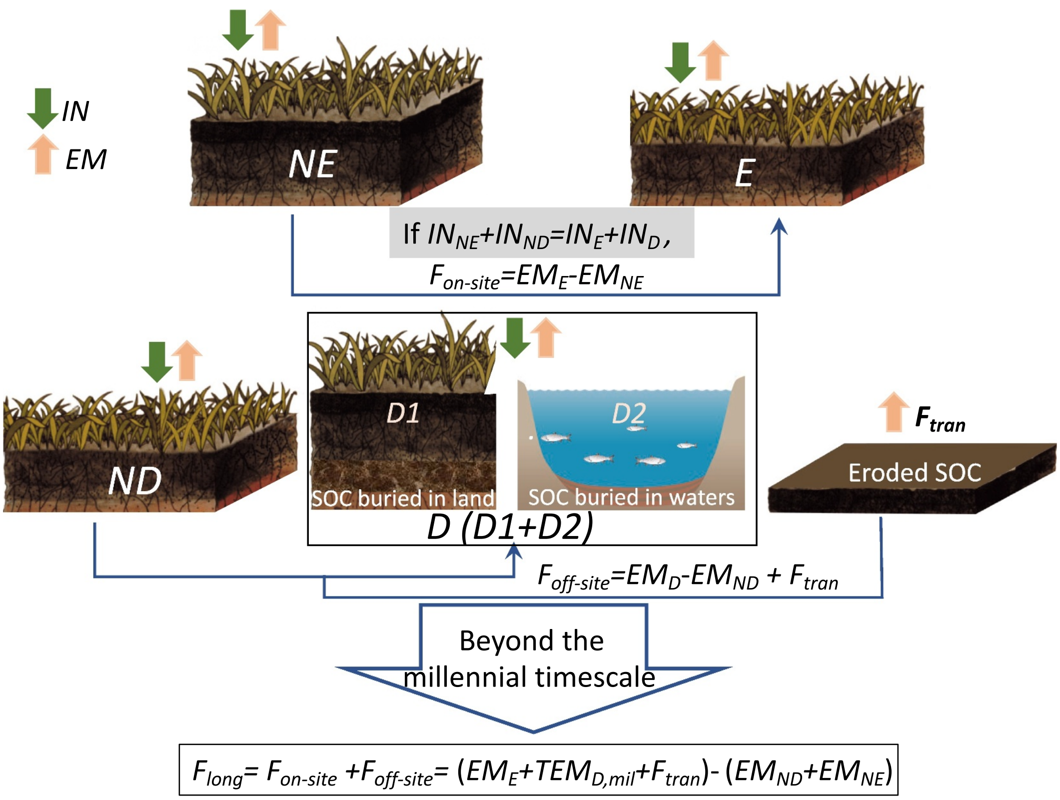

Figure 6. The framework for quantifying the physical-erosion-induced carbon flux over long timescales (beyond the millennial scale). *IN* and *EM* indicate carbon input and carbon release processes, respectively. F_{tran} is the carbon release during transport on slopes and in rivers. *E* and *D* indicated eroded and depositional sites, and *NE* and *ND* indicated noneroded and nondepositional sites, respectively. In this framework, both *NE* and *ND* can be regarded as reference sites. *D*1 and *D*2 represent two states of depositional locations.

Here, depositional areas cover widespread colluvium, alluvium, floodplains, wetlands, and coastal oceans. F_{short} is the erosion-induced carbon flux on short timescales, where $F_{\text{short}} > 0$ indicates a carbon sink effect (carbon recovery), while $F_{\text{short}} < 0$ indicates a source effect. F represents vertical flux. In other words, when erosion or deposition occurs, the impact of eroded soil outflow or inflow on the synthesis and decomposition of local SOC is profound, yet often difficult to quantify. However, by using our understanding of the underlying mechanisms, this method can accurately quantify erosion-induced carbon flux through mathematical simulation of the evolution of SOC, offering an alternative to experimental measurements, especially over large scales.

Another quantification method focuses on longer time periods and aims to quantify the changes in carbon release from soil particles before and after erosion, placing emphasis on the fate of the eroded soil material (Figure 6). While these estimates are rough, they are relatively simple to produce. If the carbon uptake rate at eroded and depositional areas, whether individually or collectively, remains constant, the erosion-induced carbon flux is primarily determined by the combined changes in on-site and off-site emitted fluxes:

$$F_{\text{on-site}} = EM_E - EM_{NE} \tag{10}$$

$$F_{\text{off-site}} = EM_D - EM_{ND} + F_{\text{tran}} \tag{11}$$

$$F_{\text{long}} = F_{\text{on-site}} + F_{\text{off-site}} \tag{12}$$

where F_{long} is the erosion-induced carbon flux over long timescales, $F_{\text{long}} < 0$ represents a sink effect, and $F_{\text{long}} > 0$ indicates a source effect; EM_E and EM_{NE} represent carbon releases when erosion does and does not occur, respectively (or before and during erosion activity); EM_D and EM_{ND} represent carbon releases when deposition action does and does not occur, respectively (or before and during deposition); and F_{tran} represents the amount of carbon released during transport processes. Over longer timescales, spanning millennia and beyond, the subsequent EM_D of deeply buried organic carbon tends to approach zero because of strong stabilization effects (Van Oost et al., 2012; Z. Wang et al., 2014). Therefore,

$$F_{\text{long}} = (EM_E + \text{TEM}_{D,\text{mil}} + F_{\text{tran}}) - (EM_{ND} + EM_{NE}) \tag{13}$$

where $\text{TEM}_{D,\text{mil}}$ is the total carbon release in burial processes during millennial timescales.

Generally, erosion-induced carbon sinks require two key conditions: dynamic replacement at eroded areas and reduced mineralization rates of buried carbon at depositional areas (Harden et al., 1999). To prove the occurrence of dynamic replacement in erosion areas, Remus et al. (2018) and Berhe et al. (2008) conducted experiments and measured net primary productivity (NPP) and carbon displacement on four different slopes, and they found a decline in the decomposition rate of SOC at both eroded and depositional areas. Using the ^{137}Cs isotope to trace the carbon movement in erosion processes, Van Oost et al. (2007) demonstrated eroded sites act as carbon sinks, and estimated a global erosion-induced sink of 0.12 Pg C yr^{-1}. If the total soil erosion rate is less than 91 t km^{-2} yr^{-1} and eroded carbon is completely replaced by new soil organic matter, and if no more than 50% of the eroded material is transported into rivers, then erosion processes act as a sink (Worrall et al., 2016), note that these preconditions are nonuniversal. From 6000 BCE to 2015 CE, agricultural erosion led to a net sink of ~78 Pg C, as calculated based on a global database (Z. Wang et al., 2017). Furthermore, an increase in temperature or erosion rate can promote an erosion-induced carbon sink (Z. Wang et al., 2023).

The contrasting perspective, and also the prevailing view in this field, is that erosion is a source of atmospheric CO_2. Jacinthe et al. (2002) quantified for the first time the percentages of carbon mineralization during transport: Approximately 50% of migrated SOC was decomposed within the first 20 days following erosion. CO_2 release from severely or moderately eroded plots was significantly higher than from slightly eroded and depositional plots because of the substantial differences in temperature and humidity (Bajracharya et al., 2000). Decreased soil moisture and increased soil temperature at eroded sites leads to the decomposition of over 20% of SOC (Lal, 2003). In a depositional environment, soil redistribution leads to an increase in carbon release equivalent to 2%–12% of the total carbon content (Van Hemelryck et al., 2010). During the period 1850–2015, the total release associated with the expansion of global croplands was approximately 98.4 Pg C (~0.6 Pg C yr^{-1}) (Lorenz & Lal, 2018). The mineralization rate in soil is higher than under static water conditions, but the mineralization rate is enhanced up to 10-fold in turbulent water (T. Liu et al., 2023). This may lead to an overestimation of buried organic carbon in static water bodies.

In short, the status of erosion as a net source or sink for atmospheric CO_2 is still under debate because of gaps and inconsistencies in the existing research. Common issues include neglecting to consider all relevant processes, differences in timescales and spatial domains, and differences in the calculation methods employed, which include experimental observations and various modeling techniques, such as physics-based, conceptual, and empirical models. Therefore, it is necessary to conduct improved in-situ field observations and laboratory experiments that consider all relevant processes across diverse environments. For example, experiments should be conducted to (a) measure the effect of erosion intensity on soil quality and crop yield (Mandal et al., 2023), (b) quantify erosion-induced changes in soil carbon storage in agricultural and nonagricultural lands (Mariappan et al., 2022), (c) reveal erosion-induced soil respiration (mineralization) dynamics (Berhe, 2012; T. Li et al., 2019; Novara et al., 2016; Van Hemelryck et al., 2010), (d) find more evidence to support where dynamic replacement is occurring in eroded areas (Remus et al., 2018), and (e) determine relevant factors and mechanisms affecting ER of sediment organic carbon (Nie et al., 2015; Schiettecatte et al., 2008b). Gaining deeper understanding of the theoretical processes and measuring critical parameters are both crucial for the development of effective models.

4.2. Quantification of Chemical-Erosion-Induced Vertical Carbon Fluxes

Carbonate minerals constitute Earth's largest carbon reservoir (Hartmann & Moosdorf, 2012). Chemical weathering of silicates (calcium silicates and magnesium silicates) and carbonate minerals are the most crucial processes contributing to carbon sinks from chemical weathering (Hilton & West, 2020). Calcium-silicate weathering can form stable carbon sinks and regulate climate change over geologic timescales (see the first process in Figure 7, net reaction: $CaSiO_3 + CO_2 = CaCO_3 + SiO_2$). The annual rate of global carbon absorption through silicate weathering is approximately 0.09–0.14 Pg C yr^{-1} (Moon et al., 2014), basically counterbalancing

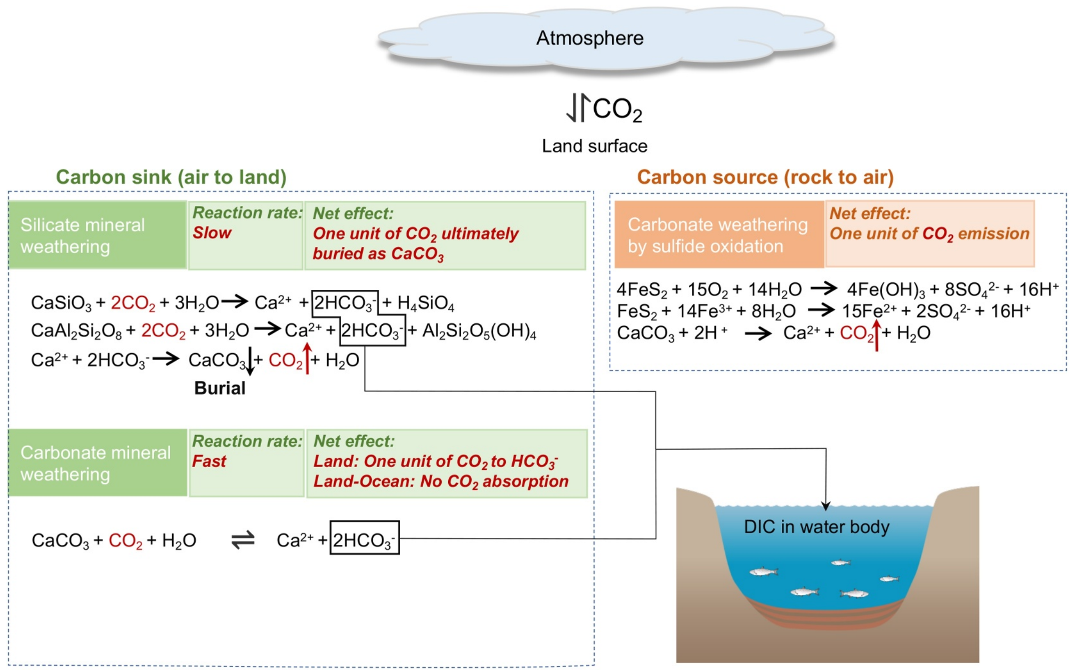

Figure 7. The processes and net effects of silicate and carbonate chemical weathering. The chemical equations show specific examples of mineral reactions.

the carbon released from global volcanoes. Traditionally, the carbon sink created by carbonate weathering (i.e., karstification processes) is considered unstable and reversible (see the second process in Figure 7). However, carbonate minerals weather more rapidly and have much higher solubility than silicate minerals (Z. Liu et al., 2011). It is estimated that the carbon absorption from carbonate weathering reaches 94% (0.477 Pg C yr^{-1}) of total carbon absorbed from global chemical weathering (Z. Liu et al., 2011).

Quantifying the rates of mineral weathering and quantifying the associated carbon absorption fluxes are equally important. The primary methods include the following: (a) the kinetic method, which reveals relationships between product concentrations and reaction times, obtains reactive kinetic parameters, and reveals underlying mechanisms (Braun et al., 2016; Gao et al., 2022; Pedrazas et al., 2021); (b) the dissolution measurement method, which quantifies the amount of dissolution and establishes a dissolution rate model to estimate the consumption of CO_2; (c) the chemical equation method, which involves the measurement of solute concentrations and fluxes in water (such as Ca^{2+}, Mg^{2+}, K^+, Na^+, SO_4^{2-}, and Cl^-) to estimate the amount of carbon consumed by mineral weathering based on the principle of mass conservation (this method considers that weathered products are ultimately transported with runoff to the watershed outlet (Jacobson & Blum, 2003; Larsen et al., 2014; Pedrazas et al., 2021)); and (d) model-based estimates, such as the Global Erosion Model for CO_2 fluxes (Suchet & Probst, 1995) and weathering front propagation models (Braun et al., 2016). The weathering rates of silicate and carbonate minerals can be affected by various factors, including lithological, climatic (especially temperature and precipitation) and hydrological conditions, plant growth, and agricultural activities (Gu & Brantley, 2022; Ludwig et al., 1996; Song et al., 2014; Szramek et al., 2007). The chemical weathering rates at the bottom of a hillslope may be influenced by the concentration of soil organic material and kaolinite (W. Liu et al., 2016). Table 2 lists the relative contributions of silicate and carbonate mineral weathering to carbon sinks across different basins and shows that carbonate weathering predominates in most basins.

5. Key Challenges for Quantifying Erosion-Induced Carbon Fluxes

Earlier in this review, we examined the processes associated with on-site and off-site erosion-induced carbon fluxes and discussed the methods for quantifying these fluxes. The central question in quantification is whether erosion facilitates the net transfer of carbon from the atmosphere to the land. Despite recent advances outlined in

Table 2
The Weathering Rates and Contributions of Carbonate and Silicate Minerals Across Different Basins

	Carbonate weathering sink		Silicate weathering sink			
Basin, country name	Unit (kmol km^{-2} yr^{-1})	Percentage contribution	Unit (kmol km^{-2} yr^{-1})	Percentage contribution	Carbonate weathering by sulfide (kmol km^{-2} yr^{-1})	References
Seine River, France	400	~95%	15–24	~5%	/	Roy et al. (1999)
Northern Okinawa Island, Japan	/	Slightly lower	334–471	Slightly higher	/	Vuai and Tokuyama (2007)
Brahmaputra catchment, international river	/	~80%	/	15%–20%	/	Hren et al. (2007)
Globe	/	94%	/	6%	/	Z. Liu et al. (2011)
Guijiang River, China	/	80%	/	20%	/	Tang et al. (2014)
Yellow River, China	/	74%	/	26%	/	Ran et al. (2015)
Mackenzie River, Canada	/	/	3,241	/	/	Hilton et al. (2015)
Globe	/	/	/	/	22.4	Torres et al. (2016)
Laval catchment, France	/	/	/	/	30.4	Soulet et al. (2018)
Taiwan, China	/	/	/	/	718	Blattmann et al. (2019)
Xijiang River Basin, China	/	76%	/	14%	/	Y. Zhang et al. (2021)
Northern Apennines, Italy	/	90%	/	10%	/	Erlanger et al. (2021)
Jialing River, China	2,042	75%	675	25%	425	W. Li et al. (2022)
Kali River, India	575	14%	3,442	86%	/	Arun et al. (2022)

Note. "/" means no figures in the cited research. Contribution refers to the ratio of absorbed carbon during carbonate or silicate weathering to the total absorbed carbon.

Section 4, achieving precise quantification of each carbon flux component on a large scale remains a big challenge (Hilton & West, 2020; Yue et al., 2016).

5.1. Challenges in Fundamental Research

In the study of organic carbon cycling, typical challenges include identifying and predicting the position, depth, and extent of depositional sites that receive migrated soil particles, as well as clarifying the physiochemical environments within these sites. When determining the characteristics of lateral soil particle redistribution, delineating the catchment area is crucial for identifying the routing scheme toward river basins and sea outlets and for determining the pathway of sediment movement, rather than using the sediment delivery ratio as the foundation of soil redistribution patterns (X. Wang et al., 2014). In addition, extending the mechanisms of soil redistribution from specific areas to other regions often faces numerous limitations (Lugato et al., 2016; Yue et al., 2016). The instability of migrated organic carbon—with its variable turnover rate, as evidenced by variation in carbon content with transport distance and time—greatly affects quantification outcomes (Bailey et al., 2019; Lal, 2005; Q. Sun et al., 2021; Y. Sun et al., 2021; Xiang et al., 2023). The decomposition of organic carbon likely increases with transport distance (Lal, 2019), and the preservation of buried carbon shows a logarithmic decline over time (Van Oost et al., 2012). Furthermore, establishing a conceptual link between slope lateral carbon fluxes and riverine carbon fluxes is paramount. This connection is pivotal for predicting one aspect of the relationship based on observations of the other.

The rapidly warming climate is having profound impacts on erosion-induced carbon fluxes. The abrupt permafrost degradation and the associated increase in erosional processes—such as thaw slumps, thermo-erosion gullies, and active layer detachments—must be taken into account. There is clear evidence of a noticeable upward trend in sediment yield within glacial and periglacial regions (Keller et al., 2021). The thermokarst process has recently gained increasing importance within the global carbon community, although it remains poorly understood (D. Li, Overeem, et al., 2021; Schuur et al., 2015; Turetsky et al., 2019, 2020). Continuing climate change is reshaping regional precipitation patterns, with a trend of increasingly frequent extreme precipitation events (Q. Sun et al., 2018). Panagos et al. (2022) reported that an increase in rainfall intensity (and erosivity) is expected to

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increase global soil losses by 30%–66% in the near future. In addition, the collapse of large dams caused by more frequent rainstorms may trigger the exposure of large amounts of buried organic carbon and promote a carbon source effect (Keller et al., 2021). So, understanding the response of terrestrial erosion-induced carbon dynamics to climate change is an important challenge.

In the context of inorganic carbon cycling, although chemical weathering models provide insights into weathering profiles at plot scales (Lebedeva et al., 2010; D. Li et al., 2014), achieving model accuracy remains a significant challenge because of incomplete understanding of weathering mechanisms and the complexity of landscape erosion at geologic timescales (D. Li et al., 2014; G. Li et al., 2016). Indirect effects triggered by erosion still deserve attention; for example, erosion provides critical rock-derived nutrients (e.g., nitrogen, phosphorus, and potassium) for the terrestrial and marine biosphere (Hilton & West, 2020), which, to some extent, promotes vegetation growth in water bodies and the absorption of atmospheric carbon and further increases the difficulty of quantification. As a result of the easy transformation of DIC into DOC, POC, and PIC in aquatic environments, the question of whether the DIC flux at the watershed outlet can accurately represent the weathering rate remains open for further investigation (Z. Liu et al., 2018). Floodplains, a prevalent type of depositional landscape, receive SOC and semi-weathered minerals, creating an ideal environment for biogeochemical cycling. However, accurately quantifying the net carbon mineralization and weathering processes in floodplains remains challenging (Hilton & West, 2020; Lupker et al., 2011, 2012).

Disentangling the response of terrestrial inorganic carbon dynamics to human activity is also challenging. Human-induced alterations of the land surface significantly influence chemical weathering rate and DIC dynamics. Several activities include deforestation (Drake et al., 2020), coal and mineral mining, dredging and quarrying (Syvitski et al., 2022; Tarolli & Sofia, 2016), large dam construction (Keller et al., 2021; Mendonça et al., 2017), landfill construction (Porowska, 2015), and agricultural practices like fertilizer application (Drake et al., 2020). Global climate warming, coupled with rising atmospheric CO_2 levels, leads to greater CO_2 dissolution in rainwater, subsequently increasing DIC concentration in rivers. Acid rain caused by air pollution can accelerate mineral chemical weathering rates (Huang et al., 2019). Consequently, there is considerable room for improving the accuracy of estimates related to carbon fluxes resulting from global chemical weathering.

5.2. Challenges in Research Methods

The acquisition of observational data is the most crucial step in simulating and analyzing erosion-driven carbon dynamics. There are several new tools to gather observed data, such as tracer proxies, molecular fingerprinting, biomarkers, multispectral imaging, radar remote sensing imaging, and combinations of these tools (Doetterl et al., 2016; Polyakov et al., 2009; Walling, 2013). However, soil data collection is usually challenging, time-consuming, and resource intensive. Sometimes, selecting appropriate methods is necessary to assess the population characteristics based on a limited number of observed samples. Assessment accuracy can be affected by measurement errors, sampling biases (especially originating from data extremes), and variability in environmental conditions. Reproducibility is critical for validating findings and ensuring the robustness of scientific conclusions, and robustness can be enhanced through documenting and standardizing experimental procedures, employing standardized protocols and methodologies, and replicating measurements or observations. Additionally, sensitivity and uncertainty analysis techniques can be used to assess the robustness of findings and identify sources of error (Wu et al., 2024).

The major challenge in accurately quantifying erosion-induced sinks and sources lies in narrowing uncertainty. This can be achieved through an improved understanding of mechanisms, the development of advanced monitoring tools, the application of excellent interpolation methods, and the use of robust simulation models. Coupling research and extrapolation methods at different scales and for different geomorphic and geologic processes is necessary and crucial. For example, long-term observation provides extensive information on the quantities and classifications of dissolved and particulate materials in rivers, allowing us to estimate the weathering rates of silicate and carbonate minerals and also to calculate the atmospheric CO_2 absorption flux. However, monitoring every river worldwide is clearly time-consuming, costly, and impractical. Therefore, a key challenge is learning how to generalize and extrapolate chemical weathering patterns in small watersheds to larger watershed scales. Machine learning algorithms offer substantial benefits in efficiency, automation, and data processing capabilities, especially in building models (e.g., for classification, regression, and interpolation). However, the algorithms demand a substantial number of high-quality explanatory variables with long time series, are sensitive to outliers, and are prone to overfitting. More importantly, their ability to extrapolate and predict values in unsampled areas is highly limited (Hassani et al., 2023; Leirvik & Yuan, 2021). Consequently, when employing machine-learning-based interpolation methods to estimate global erosion-induced carbon fluxes from available measurements, the uniformity and representativeness of samples from field experiments are crucial. High-quality input data can significantly reduce the uncertainty associated with the estimates.

6. Summary and Future Perspectives

6.1. Summary

Global atmospheric CO_2 can be removed through silicate weathering, carbonate dissolution, and organic carbon burial, all of which are linked to erosion. However, accurate quantification of erosion-induced carbon fluxes has long been a crucial challenge. Traditional studies of erosion often emphasize physical processes, while physical-erosion-accelerated chemical erosion (chemical weathering) is also an important process influencing carbon exchanges between the land and the atmosphere. The carbon uptake associated with chemical weathering (0.26–0.48 Pg C yr^{-1}) is almost comparable to the carbon flux (-1.2 to 1.5 Pg C yr^{-1}) from physical erosion. Furthermore, soil-erosion-induced carbon flux estimates vary widely due to the lack of a comprehensive understanding of mechanisms, high uncertainty in erosion and deposition rate prediction, strong human interference in land surface processes, and the various timescales and spatial domains considered in different studies. This review examines relative carbon fluxes, from on-site to off-site, from slopes to rivers, from vertical to lateral views, and from organic to inorganic carbon cycling. On-site carbon dynamics primarily involve the removal of topsoil organic carbon, carbon dynamic replacement, subsoil mineralization, and the accelerated chemical weathering of minerals. Off-site carbon dynamics predominantly include the sorting and mineralization of migrated carbon, the generation and transformation of organic and inorganic carbon during transport, and organic carbon deposition and burial at depositional areas. We have summarized the carbon fluxes entering estuaries and proposed new conceptual frameworks for quantifying the carbon fluxes induced by physical and chemical erosion over short and long timescales.

This review discusses the importance of considering both physical and chemical erosion and emphasizes the strong time dependency (delays) of sink and source processes at colluvial sites. Because of this time dependency, it is important to define initial and terminal conditions in quantitative studies. Within the context of inorganic carbon cycling, we focus on the carbon sink effect generated by carbonate weathering because DIC in water mostly originates from carbonate minerals in carbonate-dominated watersheds, and to some extent, promotes aquatic vegetation growth. In addition, short timescales (from minutes to hundreds of years) should be prioritized, as short-term annual carbon exchanges are much larger than those occurring over longer timescales (Hilton & West, 2020) and are more relevant to interactions with the modern living environment. The overarching goal in this field is to accurately quantify erosion-induced sinks and sources, but more progress is still needed. Further research into erosion-induced carbon fluxes is essential to help policymakers formulate informed policies to address future climate change.

6.2. Future Perspectives

Research priorities for understanding erosion–carbon interactions should focus on several key areas: the mineralization characteristics of SOC during transport processes (how it varies with distance and time); the turnover rate variance of organic carbon in eroded and depositional environments; the relationship between physical and chemical erosion rates for soil on slopes and material fluxes in watersheds; the effect of erosion on land productivity or NPP in the presence or absence of agricultural activities (such as irrigation, fertilization, and plowing); and soil redistribution patterns following erosion events (Van Hemelryck et al., 2011). Understanding these issues is crucial for accurate estimation of erosion-induced carbon fluxes. Additionally, because cropland management is necessary for supporting food security and human survival, mitigating climate change, and improving the ecological environment, global croplands should be studied at high spatiotemporal resolutions (Apezteguía et al., 2009; Laamrani et al., 2021).

We recommend combining modern research tools—such as soil profile examination and sampling, radiogenic isotopes (Francke et al., 2020), laboratory sample analysis, artificial intelligence, and advanced geostatistical modeling—and integrating a global, high-precision database containing water, sediment, carbon, and other solute fluxes with a detailed catchment delineation database. This review demonstrates the need for integrated research into physical and chemical erosion at small scales, such as slopes, fields, and watersheds, while also supporting research at larger scales, such as continents or the globe, using advanced models and algorithms. We urge Earth scientists worldwide to share relevant data and engage in collaborative, comparative research through a global "carbon-erosion" initiative, and we call for in-depth discussions of divergent views to seek a consensus. To this end, it would be valuable to establish an interactive repository for erosion-induced carbon fluxes, which should be managed by specialized organizations (e.g., relevant scientific societies). Finally, we recommend expanding future multidisciplinary studies that integrate fields such as climate, geology, chemistry, and biology, which would greatly improve the understanding and modeling of erosion-induced carbon dynamics and global biogeochemical cycles.

Glossary

Erosion	Erosion is a geological process in which earthen materials are eroded away and transported by internal and external forces. It typically involves depositional processes, where the eroded materials eventually settle in new locations.
Soil erosion	Soil erosion is the loosening and displacement of soil from the land due to external forces, such as water, wind, gravity, and human activity. The processes involve the disaggregation, transport, and deposition of soil particles.
Weathering	Weathering refers to the breakdown and dissolution of rocks and minerals on Earth's surface through contact with water, atmospheric gases, sunlight, and biological organisms. Unlike erosion, weathering does not involve material movement.
Physical erosion	Physical erosion, also known as mechanical erosion, refers to the disaggregation and removal of rock or soil as clastic sediment. It contrasts with chemical erosion, in which rock or soil material is removed from an area through dissolution.
Chemical erosion	Chemical erosion refers to the removal of rock or soil particles through the dissolution or detachment of material caused by chemical interaction between acidic water and rocks or minerals, including processes like hydrolysis, hydration, and acidification. It also involves the transportation of the dissolved or particulate particles away from their original locations.
Physical weathering	Physical weathering is the process that leads to the disintegration of rocks, minerals, and soils without any chemical change. It can occur due to many factors, such as aqueous phase change, temperature fluctuations, pressure, frost, root growth, and activity of burrowing animals.
Chemical weathering	Chemical weathering is a process in which rocks and materials are broken down by chemical reactions. Unlike "chemical erosion," it does not include material transport and emphasizes instead the on-site alteration of the chemical composition of materials.
Erosion associated with water	These are specific types of erosion driven by or involving water, including water erosion, gravitational erosion, freeze–thaw erosion, and chemical erosion.
On-site processes	On-site processes refer to the dynamics of organic carbon that take place at the original, now-eroded site.
Off-site processes	Off-site processes refer to the dynamics of organic carbon after it has been removed from its original, now-eroded site; this included its transport, deposition, and burial.

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Data Availability Statement

Data were not used, nor created for this research.

Acknowledgments
This research was supported by the National Natural Science Foundation of China (No. U24A20572), the National Key Research and Development Program of China (No. 2024YFF0809301) and the National Postdoctoral Researcher Program of China (No. GZC20230249). The work was also supported by Agritech National Research Center and received funding from the European Union's NextGenerationEU initiative (Piano Nazionale di Ripresa e Resilienza [PNRR]—Missione 4 Componente 2, Investimento 1.4—D.D. 1032 17/06/2022, CN00000022), and it reflects only the authors' views and opinions; neither the European Union nor the European Commission can be considered responsible for them.

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