

Deep carbon cycle from sediments in subduction zones

Jinhua Lai^{1,2} · Haiying Hu¹ · Lidong Dai¹

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Abstract Over 90% of Earth's carbon is stored in the mantle and core. The deep carbon cycle plays a critical role in regulating surface carbon fluxes, global climate, and the habitability of Earth. Carbon mainly residing within the sediments, altered oceanic crust, and mantle peridotite as carbonate minerals and organic carbon is transported to the deep Earth via plate subduction. A series of reactions (e.g., metamorphism, dissolution, and melting) occurring in the subducting slab drive the carbon removal. Some of the carbon is recycled to the surface via arc volcanism, while the rest is carried into the deeper Earth. More than two-thirds of the global subduction carbon input comes from sedimentary carbon, whose fate during subduction directly affects the flux in the global carbon cycle. Over the past two decades, the sedimentary carbon cycle in subduction zones has been extensively studied by experiments and computational approaches. Here, we provide a comprehensive review of the sources, species, decarbonation reactions, carbon cycle tracing, and fluxes of sedimentary carbon in subduction zones, and the role of sedimentary carbon subduction in climate evolution and mantle chemistry. Further research is required for our understanding of deep carbon cycle processes and their role in Earth's climate.

Keywords Deep carbon cycle · Subducted sedimentary carbon · Subduction zone · Global climate

1 Introduction

Carbon, a crucial volatile element on Earth, plays an important role in the planet's evolution, the origin of life, climate change, and resource utilization. It profoundly impacts the dynamic equilibrium and habitability of the Earth system (Stewart et al. 2019). Carbon migration within Earth's surface and interior is often individually referred to as surface and deep carbon cycles. The former occurs between the atmosphere, hydrosphere, and biosphere within relatively short timescales from days to thousands of years. The latter involves long-term carbon storage and transfers within the mantle and core over geological timescales of millions of years (Hayes and Waldbauer 2006; Orcutt et al. 2019). During the Earth's evolution, volcanic CO₂ emissions have governed the surface carbon reservoir. They contributed to the Neoproterozoic “Snowball” warming and triggered climate changes during the Cretaceous and Paleocene-Eocene epochs (Hoffman et al. 1998; Kerrick 2001). The solid Earth is estimated to contain over 90% of the planet's total carbon (Orcutt et al. 2019), which implies its importance for maintaining the stability of surface systems and understanding the global carbon cycle. Carbon stored in sedimentary, oceanic crust, and mantle layers can be carried into the deep Earth through slab subduction. The sedimentary carbon input flux is about 60 Mt/year at modern subduction zones, which constitutes over two-thirds of the total global carbon input at the trenches. The residual one-third primarily comes from the altered oceanic crust and subducting mantle peridotite (Kelemen and Manning 2015; Clift 2017; Dutkiewicz et al. 2019; Plank and Manning 2019). Water-rich fluids released from the altered oceanic crust and serpentinized lithospheric mantle during slab subducting can infiltrate sediments, causing sedimentary carbonate to dissolve (Wang et al. 2022), even triggering partial melting at sub-arc depths (Skora and

✉ Haiying Hu
huhaiying@vip.gyig.ac.cn

¹ Key Laboratory of High-Temperature and High-Pressure Study of the Earth's Interior, Institute of Geochemistry, Chinese Academy of Sciences, Guiyang 550081, China

² University of Chinese Academy of Sciences, Beijing 100049, China

Blundy 2010; Tsuno et al. 2012; Chen et al. 2023). These carbon-bearing fluids can migrate into the overlying mantle wedge, lowering the solidus temperature of peridotite and inducing partial melting, ultimately generating CO_2 -rich magmas (Behn 2011; Wang et al. 2022). Carbon dissolved in arc magma can return to the surface via volcanism. The trace elements and isotopic signatures in volcanic lava provide evidence of sediment recycling (Plank and Langmuir 1998; Walter et al. 2011; Atlas et al. 2022). In addition, light carbon isotopic signatures in ultradeep diamond inclusions reveal the potential deep subduction of organic carbon from sediments (Brenker et al. 2007; Bulanova et al. 2010). Ca and Mg isotope characteristics in basalts indicate that sedimentary carbonates can be subducted into the deep mantle (Huang et al. 2015; Li et al. 2017; Banerjee et al. 2021). Therefore, the fate and flux of sedimentary carbon are significant for the deep carbon cycle.

The deep carbon cycle within the Earth, associated with the petrological processes and carbon fluxes, has been summarized by recent studies (Hayes and Waldbauer 2006; Dasgupta and Hirschmann 2010; Kelemen and Manning 2015; Plank and Manning 2019). However, although subducted sedimentary carbon has been the subject of numerous studies, no comprehensive literature review has examined recent advances. Consequently, we conduct a review regarding its sources and forms, decarbonation mechanisms, tracer characteristics, and the flux of subducted sedimentary carbon in the mantle, as well as the influence of the sedimentary carbon cycle on global climate and mantle chemistry.

2 Sources and storage of subducted sedimentary carbon

2.1 Primary sources of sedimentary carbon

Subducted sedimentary carbon can be classified into two types according to its origin: organic (OC) and inorganic carbon (IOC). Organic carbon includes material derived from both terrestrial and marine environments, such as river-borne detritus, marine plankton, and microbial biomass associated with hydrothermal systems (McCollom and Shock 1997; Kennedy et al. 2004; Galy et al. 2007). The OC flux is higher in deep-sea fans, which are characterized by high primary productivity and rapid sedimentation rates. Although the OC content in most sediments is less than 1 wt%, it can dominate carbon input in some deep-sea fans at convergent margins. For instance, the turbidite section of the Bengal fan, with 1.5 km thickness and an OC content of 0.35 wt%, contains more carbon than the average oceanic crust (Plank and Manning 2019). Inorganic carbon mainly resides within carbonate minerals, including (1) carbonate precipitation from chemical reactions between Ca^{2+} ,

Mg^{2+} , HCO_3^- , and CO_3^{2-} in seawater; (2) fish excretions, particularly high-Mg calcite, which contribute significantly to sedimentary carbonate content; and (3) skeletal remains of marine organisms, such as calcareous algae and foraminifera, which are the primary components of sedimentary carbonates (Perry et al. 2011; Turchyn et al. 2021). Plank and Manning (2019) suggested that the carbon content in a 100-m-thick microplankton fossil layer is almost equal to that of the underlying oceanic crust.

2.2 Organic carbon preservation

The burial of OC in marine sediments is the primary link between surface OC reservoirs (i.e., oceans and atmosphere) and long-term carbon pools (i.e., the lithosphere). It leads to the removal of CO_2 and the input of O_2 into the atmosphere through the balance between photosynthesis and respiration (Burdige 2007; Eguchi et al. 2019). Research on factors controlling OC burial in subduction zones, the total proportion of OC, and how the proportion of OC in subduction zones changed over geological timescales is essential for understanding the global carbon cycle.

Factors that control OC burial in subduction zones have been widely studied, primarily focusing on primary production, sediment transport, and bottom-water O_2 concentration. Key factors include organic matter sources (Prah et al. 1997; Burdige 2007), primary production (Kohfeld et al. 2005; Moore et al. 2013), sediment accumulation rate, microbial activity (Matsumoto 2007), oxygen exposure time (Hartnett et al. 1998; Jessen et al. 2017), co-precipitation with reactive ions (Lalonde et al. 2012), and sorptive preservation on mineral surfaces (Keil et al. 1994; Hedges and Keil 1995).

House et al. (2019) analyzed carbon content from Sunda Margin drilling sites, revealing that OC comprises approximately 75% of the total carbon delivered to the trench, while accounting for only 10%–25% of what is subducted beyond 20 km depth. Clift (2017) estimates that around 60 Mt/year of carbon is subducted beneath the outer forearc, with 20% in the form of OC. This estimate aligns with findings from Hayes and Waldbauer (2006). Given the unique compositional and physical characteristics of each subduction zone, as well as discrepancies in sampling sites and estimation methods, the estimated proportion of OC varies between nearly 10% and 25% (Li and Bebout 2005; Hayes & Waldbauer 2006; Clift 2017; House et al. 2019).

Global OC burial exhibits significant temporal and spatial variation, largely influenced by climate and oceanographic conditions. Over the past 150 kyr, burial rates increased markedly during glacial periods, especially in the tropical Atlantic, eastern Pacific, and Subantarctic regions (Cartapanis et al. 2016). During the Pleistocene, OC burial was up to two orders of magnitude higher than in the Neogene (Cartapanis et al. 2016; Li et al. 2023). Modern OC burial

remains elevated in regions with high productivity and low oxygen, such as the tropical/subtropical eastern Pacific, Arabian Sea, and subarctic zones (Duncan and Dasgupta 2017). Clift (2017) also highlights significant burial in the northeastern Indian Ocean and eastern Pacific, supported by drilling data from Java, Sumatra, the Andaman–Burma margin, Makran, Chile, and Peru.

2.3 Inorganic carbon in the crust and mantle

Carbon in the crust is primarily stored as carbonates or graphite within sedimentary, metamorphic, and igneous rocks (Wilkinson and Algeo 1989; Cook-Kollars et al. 2014). This carbon can be subducted into the mantle, where it may either return to the surface via arc volcanism and outgassing or be stored in deep reservoirs (Kelemen and Manning 2015). Due to its low solubility in silicates, mantle carbon

mainly resides in accessory phases such as carbonates, graphite, diamond, or carbides (Keppler et al. 2003; Panero and Kabbes 2008).

The stability and form of carbon species vary with depth, as shown in Fig. 1, controlled by pressure, temperature, and oxygen fugacity (Stagno and Frost 2010). In the shallow mantle layer, carbon exists primarily as carbonates, including aragonite, calcite, dolomite-ankerite, and magnesite (Dasgupta and Hirschmann 2010). At sub-arc depths, stable carbonate species in sediments are primarily calcite and dolomite, which transform into aragonite and magnesite at greater depths (Grassi and Schmidt 2011a; Brey et al. 2015; Chen et al. 2021). With increasing depth and temperatures exceeding the solidus, these carbonates decompose into CO_2 -rich melts or vapor (Thomsen and Schmidt 2008; Grassi and Schmidt 2011b; Tsuno et al. 2012). At depths beyond 150 km, extreme reduction leads to the formation of

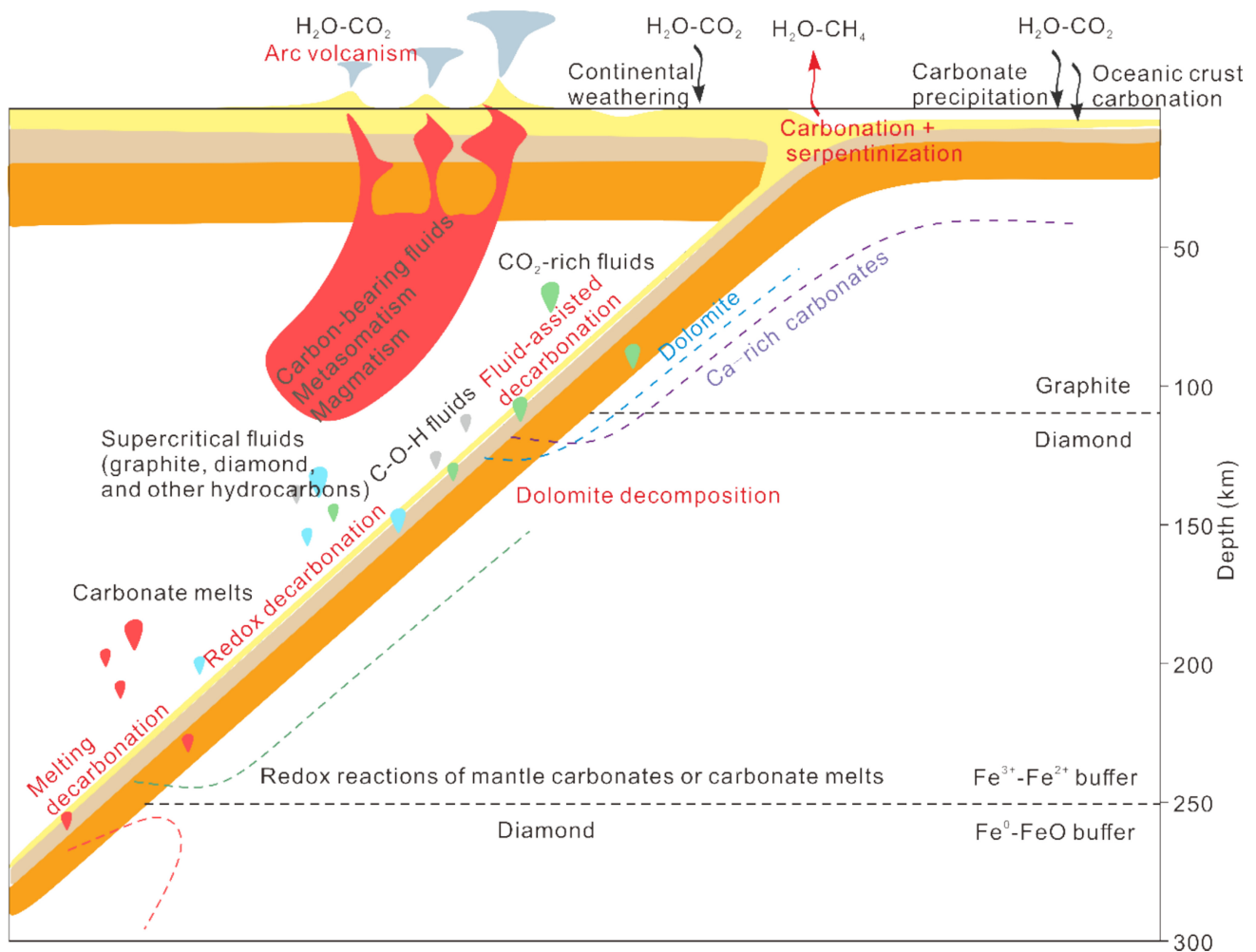


Fig. 1 The key processes of the deep carbon cycle in subduction zones (adapted from Zhang et al. 2017), including subduction of carbonates, decarbonation reactions that release carbon to fluids and melts (balloon-shaped shaded area), variation in carbon-bearing phases with depth, and arc volcanism. Dashed lines indicate the stability boundaries of various carbon-bearing phases, such as dolomite, Ca-rich carbonates, graphite, and diamond, with depth

diamond and C–H fluid (Brey et al. 2015; Hammouda and Keshav 2015).

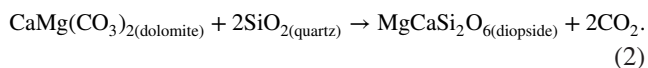
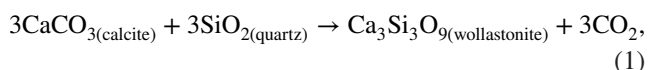
Reduced carbon species such as graphite, diamond, and carbides can persist in most regions of the lower mantle (Dasgupta and Hirschmann 2010). Diamonds, likely originating from the transition zone or deeper, may represent a major carbon reservoir in the convecting mantle (Smith et al. 2016; Nestola 2017). It has also been proposed that the upper mantle may be saturated with Fe–Ni alloys, while the lower mantle is predominantly Fe-saturated (Frost et al. 2004; Rohrbach et al. 2007). Furthermore, in iron-rich metal regions, carbon may exist as graphite, diamond, carbides, or as a dissolved component in the metal (Lord et al. 2009; Dasgupta et al. 2009; Liu et al. 2015; Chen et al. 2016; Smith et al. 2016).

Although the mantle becomes increasingly reduced with depth, carbonates may remain stable in localized oxidizing zones, as evidenced by carbonate mineral inclusions in diamonds (Brenker et al. 2007; Bulanova et al. 2010) and preserved sedimentary carbonate components in deep-sourced magmas (Chen et al. 2017; Xue et al. 2018).

3 Decarbonation mechanism

3.1 Metamorphic decarbonation

As plate subduction progresses, increasing temperature and pressure drive metamorphic reactions in silicate and carbonate minerals, leading to mineral transformations and CO₂ release. In the absence of silicate minerals, carbonates may decompose directly into mineral oxides and CO₂. Typical metamorphic decarbonation reactions include the following:



Increasing temperature facilitates decarbonation (Stewart et al. 2019). Carbonate–silicate mineral assemblages (e.g., calcite and quartz) can remain stable under specific pressure–temperature conditions and fluid composition. However, once the temperature exceeds a critical threshold, prograde metamorphism triggers CO₂ release, as shown in Fig. 2. Phase equilibria calculations of sediments, altered oceanic crust, and mantle peridotites indicate that (1) dehydration and decarbonation occur independently; (2) in high-heat-flow plates, dehydration occurs before the slab reaches the arc, while decarbonation remains limited; and (3) in low-heat-flow plates, dehydration and decarbonation reactions are nearly absent (Kerrick and Connolly 1998, 2001a, b).

The variations in H₂O and CO₂ content during the metamorphism of different oceanic sediments are shown in Fig. 3

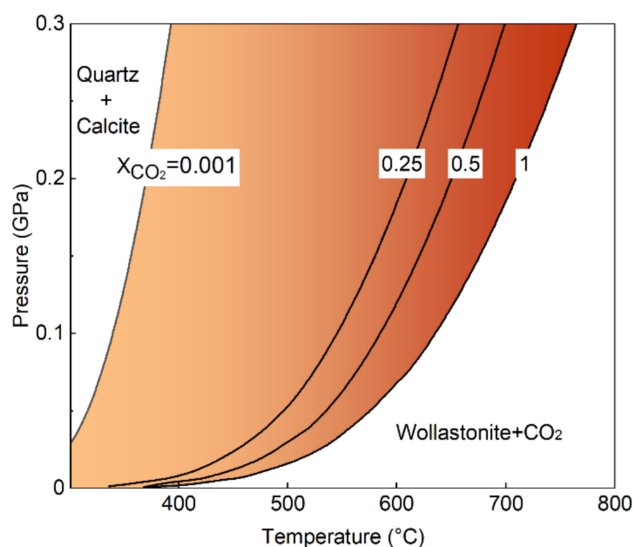


Fig. 2 Pressure–temperature diagram depicting the stability of the quartz and calcite assemblage in contrast to wollastonite and CO₂ (from Stewart et al. 2019). X_{CO_2} represents the mole fraction of CO₂ in the fluid, and the curves in the figure delineate the stability fields of phase equilibrium between the minerals and fluids with varying CO₂ content

(Kerrick and Connolly 2001a). At sub-arc depths, siliceous limestones with richer carbon release approximately 1 wt% CO₂ and nearly all H₂O in hot slabs, while in cold slabs, almost all H₂O is released and CO₂ is largely retained. Thus, most CO₂ in siliceous limestones is stored beneath the sub-arc mantle. In comparison, mudstones with lower carbonate and higher clay content release nearly all volatiles in the forearc regions under the context of hot subduction, but they undergo significant dehydration and no decarbonation in cold slabs. Therefore, despite lower initial carbon content, mudstones may release carbon more efficiently at sub-arc depths. In carbonate-bearing protoliths (e.g., mudstones), the isotherms of H₂O and CO₂ follow the geothermal gradient between 80 and 180 km depth, implying volatile retention to greater depths. Metamorphism, especially when aided by water-rich fluids, can release 40%–60% of slab carbon (Kerrick and Connolly 2001a; Ague and Nicolescu 2014; Stewart and Ague 2020; Arzilli et al. 2023). Metamorphism works in combination with dissolution to significantly enhance carbon loss, providing a plausible mechanism for the substantial CO₂ emissions observed in carbonate-rich arc magmas.

3.2 Fluid-driven decarbonation

In the presence of fluids, carbonate minerals may undergo dissolution, allowing carbon species to be mobilized with the fluid phase. For example,

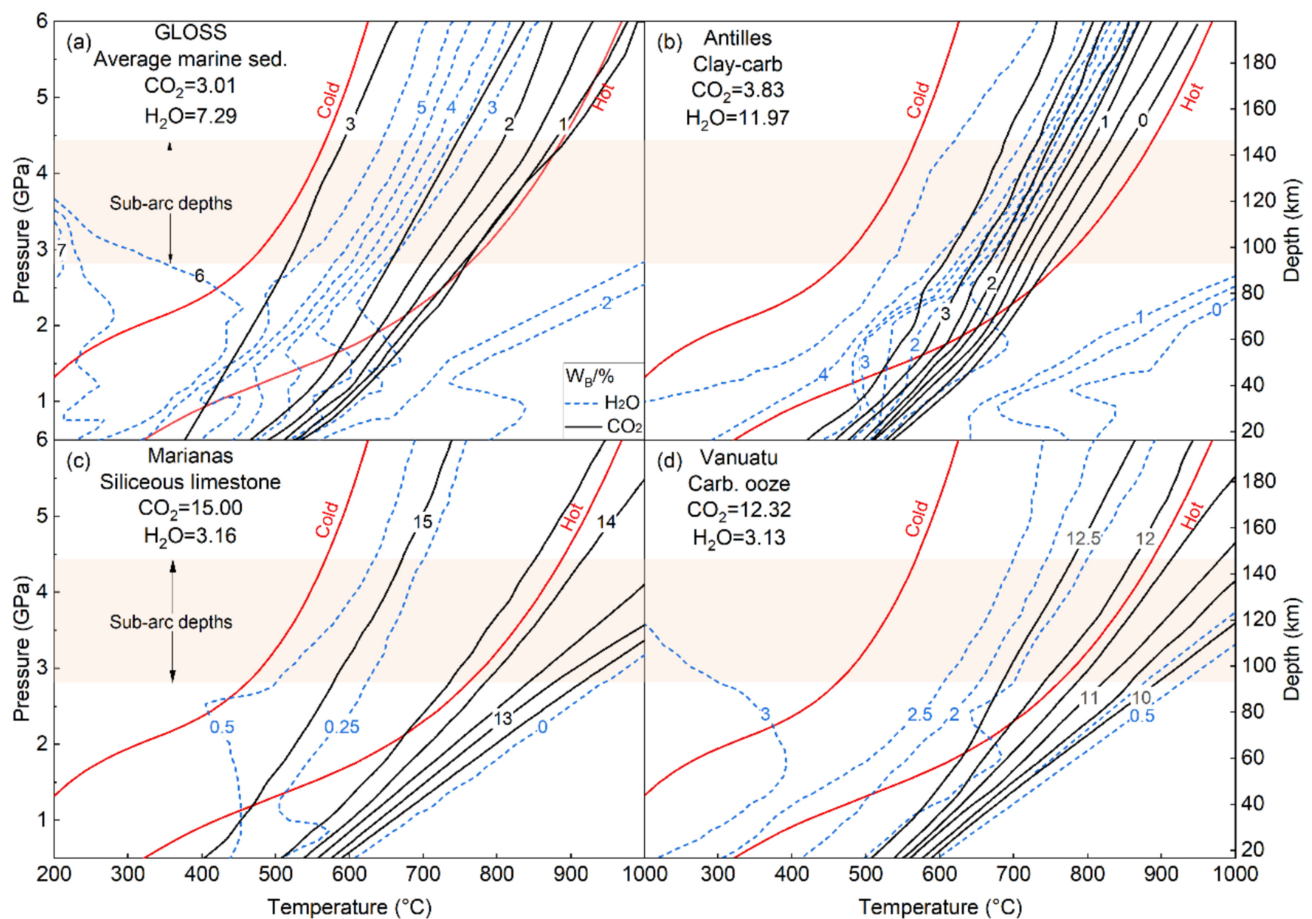
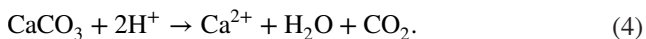


Fig. 3 Variations in H₂O and CO₂ content during the metamorphism of different marine sediments (from Kerrick and Connolly 2001a). **a** GLOSS, **b** Antilles, **c** Marianas, and **d** Vanuatu. The initial CO₂ and H₂O content of the materials are indicated. Blue lines show wt% H₂O and black lines show wt% CO₂ retained in the solid phases, respectively. The orange area denotes the sub-arc depths



Fluid-driven dissolution has been observed in various subduction settings. At the Mariana forearc seamount, fluids released from serpentinite diapirs precipitate carbonate minerals, forming authigenic aragonite (Haggerty 1987). At depths of 10–50 km, CO₂-rich fluids derived from subducted sediments metasomatize the overlying peridotite, producing listvenites composed of magnesite + quartz or dolomite + quartz. This indicates that carbonated peridotite may serve as an effective carbon reservoir at the front of the mantle wedge (Falk and Kelemen 2015). At greater depths, fluid and solid inclusions in diamonds from ultrahigh-pressure rocks in the Western Alps of Italy contain HCO₃[−], CO₃^{2−}, and carbonate crystals but lack gaseous CO₂ (Frezzotti et al. 2011; Falk and Kelemen 2015). This suggests that diamonds precipitate from fluid and that carbon is transported via dissolution rather than degassing.

In the altered marble complexes of Syros and Tinos Islands in Greece, carbon–oxygen isotope analysis reveals

that 60%–90% of the carbon is removed by dissolution, significantly exceeding the predictions from simple metamorphic devolatilization (Ague and Nicolescu 2014). Additionally, inclusions of calcite, magnesian calcite, and dolomite in diamonds suggest that water–carbonate interaction may extend the 660-km boundary (Brenker et al. 2007; Tschauner et al. 2018).

Carbonate dissolution is influenced by temperature, pressure, mineral type, fluid flux, and solute composition (e.g., NaCl) (Liu et al. 2019; Farsang et al. 2021b; Lan et al. 2022). The solubility of carbonates increases with pressure and temperature, peaking near melting conditions (Caciagli and Manning 2003; Farsang et al. 2021a). Brines significantly enhance carbonate solubility, which scales quadratically with salinity (Newton and Manning 2002). In subduction zones, the solubility of carbonates follows the trend: strontianite > calcite/aragonite > dolomite, rhodochrosite, smithsonite ≫ magnesite (Pan et al. 2013; Farsang et al. 2021b). The difference in solubility between calcite/aragonite and magnesite can exceed two

orders of magnitude, increasing further with pressure (Pan et al. 2013; Farsang et al. 2021b, c). Highly soluble calcite/aragonite is likely to be dissolved in water-rich metamorphic fluids released from the subducting slab and may be recycled to the surface at shallow depth (Newton and Manning 2002; Caciagli and Manning 2003; Facq et al. 2014). In contrast, dolomite and magnesite are more stable and insoluble. Thus, they can be transported into the deep mantle along with the slab, potentially making an important contribution to the deep carbon cycle (Shen et al. 2018; Farsang et al. 2021b).

3.3 Melting decarbonation

Carbonates in water-rich fluids melt when slab geotherms reach the carbonate solidus temperature, with carbon dissolving in the melt and migrating upward into the overlying mantle. Melting experiments suggest that marine sediments are typically the first materials in subducting slabs to undergo partial melting, compared to the other two lithologic components in the subduction zone (Thomsen and Schmidt 2008; Grassi and Schmidt 2011b; Shatskiy et al. 2019; Chen et al. 2021). The fate of carbonate-bearing sediments is influenced by volatile content (H_2O and CO_2), major cations (Ca^{2+} , Mg^{2+} , etc.), alkali metals (K^+ , Na^+ , etc.), oxygen fugacity, and pressure–temperature paths (Dasgupta et al. 2005; Dasgupta and Hirschmann 2010).

As shown in Fig. 4, the presence of water significantly enhances partial melting of carbonate sediments, particularly at sub-arc depths and within the transition zone (Grassi and Schmidt 2011a, b; Tsuno et al. 2012). The influence of CO_2 on the sediment system is more complex, as it may either elevate or lower the melting temperature, likely modulated by the content of alkali metals (Spandler et al. 2010; Tsuno and Dasgupta 2011, 2012). This trend is consistent with observations in other carbonate systems (Dasgupta and Hirschmann 2006; Duncan and Dasgupta 2014; Hammouda and Keshav 2015). In dry carbonate-bearing mudstones, adding 1 wt% H_2O lowers the solidus by $\sim 210^\circ\text{C}$, while 5 wt% CO_2 raises it by $\sim 75^\circ\text{C}$, indicating a much stronger melting effect of water than CO_2 (Tsuno and Dasgupta 2012).

Sedimentary composition critically influences melting behavior. Dolomite-rich rocks, characterized by a $\text{Ca}^\#$ of 0.4–0.5, are among the first components to melt (Dasgupta et al. 2005). Under identical pressure conditions, the melting temperatures of Na_2CO_3 and K_2CO_3 are approximately 200°C lower than that of CaCO_3 (Dasgupta and Hirschmann 2007; Litasov et al. 2013). Consequently, K^+ and Na^+ significantly reduce the solidus temperature of carbonate–silicate systems. In addition, low-aluminum carbonated pelites melt more easily than high-alumina counterparts (Grassi and Schmidt 2011a, b; Tsuno et al. 2012).

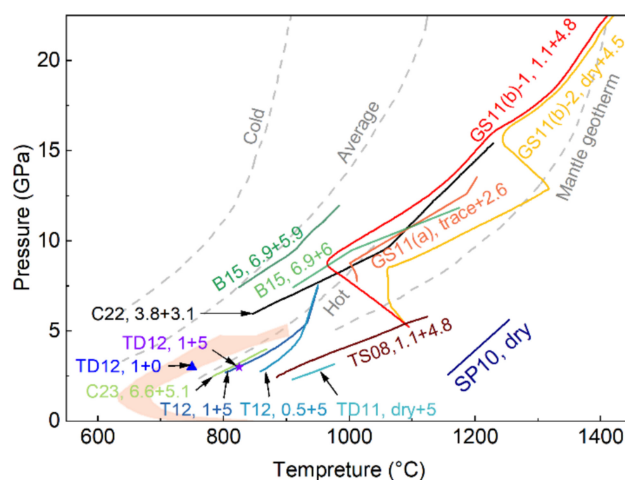


Fig. 4 Melting curve of carbonated pelite from high-temperature and high-pressure experiments. The numbers indicate the weight percentage of H_2O and CO_2 in the initial materials used in each study. The orange area represents the solidus of water-saturated sediments. Data source: geothermal gradient curves of cold and hot subduction zones and the mantle geotherm (Akaogi et al. 1989; Kincaid and Sacks 1997; Van Keken et al. 2002; Peacock 2003), C23 (Chen et al. 2023), C22 (Chen et al. 2021), B15 (Brey et al. 2015), TD12 (Tsuno and Dasgupta 2012), T12 (Tsuno et al. 2012), GS11(b)-1&2 (Grassi and Schmidt 2011a), GS11(a) (Grassi and Schmidt 2011b), TD11 (Tsuno and Dasgupta 2011), SP10 (Spandler et al. 2010), TS08 (Thomsen and Schmidt 2008), sediment (Nichols et al. 1994; Hermann and Spandler 2008)

3.4 Diapirs of subducted sediment

Most carbonate-rich sediments are expelled from the slab during subduction, forming buoyant diapirs due to their lower density and viscosity relative to the overlying mantle (Behn et al. 2011). These diapirs ascend, undergo decompression or heating, and subsequently generate carbonate melts (Liu et al. 2015; Chen et al. 2016; Wang et al. 2024). The buoyant melts efficiently extract carbon from the sediments, percolate through the altered mantle wedge (Kono et al. 2014; Wang et al. 2024), and eventually feed the magma reservoirs.

Recent geophysical studies suggest that scatterers detected at 60–95 km depth above the Ryukyu Plate may represent diapirs in the mantle wedge (Lin et al. 2020, 2021). Comparable features have also been observed in the forearc regions of Cascadia, Alaska, and the Aleutian Islands, which may serve as significant carbon reservoirs (Barry et al. 2019; Scholl 2021).

Diapirism is more efficient than metamorphic or melting-induced decarbonation in subduction zones (Marshall and Schumacher 2012; Chen et al. 2016). Wang et al. (2024) estimate that diapirism is responsible for about 80% of carbon removal from the subducting slab and may

be the dominant release mechanism in regions such as the Cyclades (Greece) and Costa Rican forearcs.

A geodynamic model by Klein and Behn (2021) indicates that sediment thickness, feldspathic content, and thermal gradients control diapir formation. Thicker, feldspar-rich sediments in warmer settings promote diapirism, enhancing forearc carbon storage. In contrast, diapirism is inhibited in colder arc regions with magnesium- and iron-rich sediments, with thinner sediments, such as in the Tonga–Kermadec and Mariana–Izu–Bonin arc systems (Behn et al. 2011; Wang et al. 2024). Furthermore, carbon-bearing materials within a 100-m-thick, wet quartz rheological sediment layer can rise as buoyant diapirs to sub-arc depths, forming substantial carbon reservoirs in the lithosphere or even the lower crust (Behn et al. 2011; Kelemen and Behn 2016). Consequently, sediment diapirism may significantly reduce carbon transport into the deep mantle and alter deep Earth carbon reservoirs.

4 Evidence for sedimentary carbon in the mantle

4.1 High-pressure experiments and numerical simulations

High-pressure experiments and thermodynamic modeling provide compelling evidence that sedimentary carbon can be recycled into Earth's mantle through subduction processes. As subduction progresses, carbonate minerals in the slab undergo decarbonation, releasing CO_2 that may be returned to the surface through arc volcanism (Gorman et al. 2006; Tsuno and Dasgupta 2011). However, not all carbon is released at shallow depths. Melting experiments on carbonated pelites suggest that 70%–80% of carbonates typically bypass the volcanic arc region and are transported into the deeper mantle (Grassi and Schmidt 2011a, b; Chen et al. 2021, 2023). According to the steady-state hypothesis proposed by Javoy et al. (1982), approximately 1.8×10^{24} g of carbon have been transferred from deep reservoirs to the surface over 4 billion years. This amount is equivalent to 22 times the present surface carbon inventory, implying substantial recycling of surface-derived carbon back into Earth's interior. Furthermore, thermodynamic modeling predicts that nearly all CO_2 is released in hotter subduction zones, whereas in colder and intermediate-temperature settings, slabs may retain up to 80% of their initial carbonate content (Connolly 2005; Gorman et al. 2006). This behavior is consistent with Earth's thermal evolution. During the Archean, when mantle temperatures were approximately 200 °C higher than today (Abbott et al. 1994), subducting slabs likely experienced complete decarbonation. The subsequent increase in carbonate preservation through the Proterozoic and Phanerozoic correlates with the planet's gradual

cooling and supports greater retention of sedimentary carbon in the mantle over time.

4.2 Natural samples

Natural high-pressure metamorphic rocks provide direct mineralogical and isotopic evidence for the subduction of sedimentary carbonates into the mantle. Marbles containing microdiamond inclusions and/or other ultrahigh-pressure (UHP) metamorphic features, particularly from orogenic belts, are considered ideal samples for studying deep sedimentary carbonate recycling (He et al. 2017). In the Kokchetav Massif, microdiamonds have been discovered within garnet and clinopyroxene in dolomitic marbles. Fe–Mg partitioning between these minerals indicates peak metamorphic conditions of 800–1000 °C and pressures exceeding 4 GPa (Ogasawara and Aoki 2005; Schertl and Sobolev 2013). In situ oxygen isotope analyses of dolomitic marble and garnet samples further support a sedimentary origin for these rocks, revealing isotopic heterogeneities linked to initial sediment composition and isotope exchange processes (Sobolev et al. 2011). Similar evidence has been reported in the Dabie Mountains, where diamond and coesite inclusions were identified in eclogite–marble assemblages (Okay 1993). These findings confirm subduction of sedimentary rocks to depths greater than 100 km and exhumation along steep P–T trajectories (Schertl and Okay 1994; Zhang and Liou 1996). UHP metamorphic marbles from the Bohemian Massif also provide comparable evidence (Becker and Altherr 1992). Additional support comes from carbonatitic xenoliths in basalts from the Dalihu volcanic field. These samples contain microscopic diamonds and exhibit O–Sr–Pb isotopic and trace element composition similar to sedimentary limestones, suggesting carbonate sediments can be subducted to depths exceeding 120 km (Liu et al. 2015). This represents the first direct geochemical and mineralogical evidence for deep recycling of sedimentary carbonates within subduction zones.

4.3 Carbon isotopes

Carbon isotopic composition in diamonds provides key evidence for the recycling of sedimentary carbonates into the mantle. The light carbon isotopic signature observed in diamonds from kimberlites, along with their silicate inclusions, supports this hypothesis (Brenker et al. 2007; Bulanova et al. 2010). There is a significant distinction in the carbon isotopic composition of IOC and OC in sediments: $\delta^{13}\text{C} \approx 0\text{‰}$ for IOC and $\delta^{13}\text{C} < -15\text{‰}$ for OC (Schidlowski 1988; Cartigny 2005; Galy et al. 2010). The typical mantle $\delta^{13}\text{C}$ value is $-6\text{‰} \pm 2\text{‰}$ (Hoffman et al. 1998; Mason et al. 2017), as shown in Fig. 5. Diamonds with $\delta^{13}\text{C}$ values $< -15\text{‰}$ in mantle rocks or inclusions suggest the presence of crustal

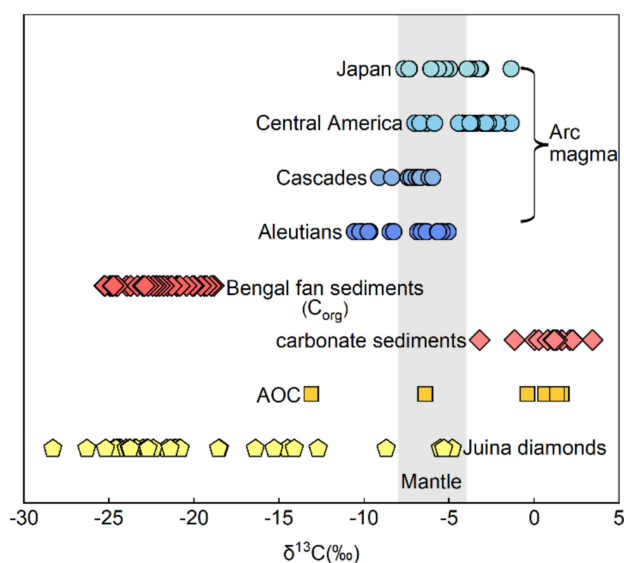


Fig. 5 Carbon isotope comparison: subducted inputs, diamonds, and arc volcanic gas. The shaded area is the $\delta^{13}\text{C}$ range of typical mantle values. Data sources: carbonate sediments (House et al. 2019), total OC in Bengal fan sediments (Galy et al. 2010), upper volcanic layers of altered oceanic crust (AOC) (Alt 2003; Li et al. 2019), mantle (Hoffman et al. 1998), Juina diamonds (Walter et al. 2011; Thomson et al. 2014), arc volcanic gases (Mason et al. 2017)

material, as the mantle is typically devoid of biogenic carbon. For instance, Bulanova et al. (2010) reported diamonds from the Juina-4 kimberlite in Brazil with $\delta^{13}\text{C}$ of -25‰ , with host inclusions resembling high-pressure phases. This suggests these diamonds may have formed from subducted marine sediments that melted in the transition zone. Walter et al. (2011) observed similar isotopic patterns in diamonds from the region, with $\delta^{13}\text{C}$ values ranging from -1‰ to -24‰ , indicating that these diamonds originated from the lower mantle, which supports the hypothesis of deep subduction of organic carbon. However, diamonds from the Jericho kimberlite in northern Slave show an extremely low $\delta^{13}\text{C}$ of -38‰ , which is challenging to explain solely by subducted sediments, as no known sediments exhibit such low values (Shirey et al. 2013). This suggests the involvement of an unknown isotopic process. Furthermore, distinguishing the carbon source based solely on carbon isotopes is challenging due to the slight difference between sedimentary carbonates and primitive IOC in the mantle (Thomson et al. 2014). In open systems, processes such as magma degassing lower carbon isotopes (Cartigny 2005; Aubaud et al. 2005), while the escape of light carbon isotopes from methane increases the $\delta^{13}\text{C}$ in reduced carbon species (Cook-Kollars et al. 2014). To refine the understanding of the source of carbon, a combined analysis of mineral inclusions and trace element isotopes is essential. Additionally, geochemical tracers less affected by magma degassing, such as metal stable isotopes,

should be introduced to track the storage and recycling of sedimentary carbon in subduction zones.

4.4 Stable metal isotopes—Mg and Ca

Magnesium (Mg) and calcium (Ca) isotopes have become essential tracers for studying the recycling of sedimentary carbonates into the Earth's mantle. Magnesium and calcium are the two most abundant cations in carbonates, and variations in their isotopic composition between surface carbonates and the mantle provide important insights into subduction and deep carbon recycling processes (Li 2015; Banerjee et al. 2021).

Magnesium isotopic composition in biological carbonates, such as those from foraminifera and coccolithophores, shows significant light isotopic enrichment, with $\delta^{26}\text{Mg}$ values ranging from -6.19‰ to -1.04‰ (Wombacher et al. 2011; Liu et al. 2022). In contrast, mantle minerals typically display $\delta^{26}\text{Mg}$ values ranging from -0.25‰ to $+0.08\text{‰}$ (Yuan et al. 2023; Liu et al. 2023), which are close to zero and stable. These differences make magnesium isotopes a powerful tool for tracing the recycling of sedimentary carbonates in subduction zones. For instance, low $\delta^{26}\text{Mg}$ values in Cenozoic basalts from eastern China are linked to the recycling of sedimentary carbonates from the subduction of the Western Pacific Plate (Huang et al. 2015; Li et al. 2017). Similarly, volcanic rocks from regions such as Jeju Island (South Korea) (Kim et al. 2019), Vietnam (Hoang et al. 2018), and Tengchong (China) (Liu et al. 2017) exhibit low $\delta^{26}\text{Mg}$ values, suggesting recycled sediments play a significant role in the mantle carbon reservoir. However, low $\delta^{26}\text{Mg}$ values do not always directly indicate recycled carbonates. Carbonatized eclogites and sedimentary carbonates exhibit similar light magnesium isotopic signatures (Sun et al. 2017; Tian et al. 2018), and magmas can also become enriched in low $\delta^{26}\text{Mg}$ values due to the ilmenite crystallization (Liu and Li 2019; Wang et al. 2021), highlighting the complexity of using magnesium isotopes alone to track recycled sedimentary carbonates.

Calcium isotopes also provide valuable insights into the recycling of sedimentary carbonates. Hawaiian basalts, for example, exhibit $\delta^{44/40}\text{Ca}$ variations of approximately 0.3‰ , with correlations to Sr/Nb and $^{87}\text{Sr}/^{86}\text{Sr}$ ratios suggesting that sedimentary carbonates are subducted into the mantle plume (Huang et al. 2011). The $\delta^{44/40}\text{Ca}$ values of the mantle generally range from 0.94‰ to 1.05‰ , closely matching the silicate Earth average ($0.89\text{‰} \pm 0.22\text{‰}$), while marine carbonates exhibit lighter $\delta^{44/40}\text{Ca}$ values ($0.6\text{‰} \pm 0.02\text{‰}$), further supporting the contribution of recycled sedimentary carbonates (Kang et al. 2017; Banerjee et al. 2021). However, other processes, such as low-temperature fractionation and interactions between melts and peridotites, can also

cause significant calcium isotopic fractionation (Zhao et al. 2017; Simon 2022; Eriksen and Jacobsen 2022).

To more accurately trace recycled sedimentary carbonates, it is essential to combine calcium and magnesium isotope data with other geochemical tracers, including large-ion lithophile elements (LILEs), rare earth elements (REEs), and isotopic ratios of radioactive elements. This multi-indicator approach will provide a more comprehensive understanding of the complex processes involved in the recycling of sedimentary carbon in the mantle.

5 Sedimentary carbon cycle flux

5.1 Carbon input flux from subducted marine sediments

Carbon input in subduction zones is mainly through sediment, oceanic crust, and peridotite, with sedimentary carbon contributing around two-thirds of global carbon input at trenches. Sedimentary carbon flux in subduction zones has exhibited considerable variability, as shown in Table 1 and Fig. 6. Rea and Ruff (1996) estimated that approximately 1.4 Gt of sediment enters the global trench systems annually. Combining the geochemical systematics with convergence rate and other parameters, Plank and Langmuir (1998) estimated the sedimentary carbon flux to be 11 Mt/year. Dasgupta and Hirschmann (2010) suggested that the carbon flux associated with subducting sediment represents about one-third of the total carbon flux of basaltic crust,

with values ranging from 13 to 17 Mt/year. Based on existing budgets for sediment flux in global trench systems, Clift (2017) revised the global subduction flux of sedimentary carbon to 60 Mt/year, higher than earlier estimates (Plank and Langmuir 1998; Dasgupta and Hirschmann 2010). Dutkiewicz et al. (2019) also indicated a flux of 57 Mt/year, considering global tectonic models and deep-sea carbonate deposition history. In the Sunda trench, House et al. (2019) estimated a sedimentary carbon flux of 0.5–1.1 Mt/year by integrating geochemical data with seismic models. Notably, these estimates have shown considerable discrepancies due to sampling and data limitations. Geophysical methods (Clift 2017) combined with geochemistry data (House et al. 2019) provide more consistent estimates, suggesting a global sedimentary carbon influx of approximately 60 Mt/year (Dutkiewicz et al. 2019; Plank and Manning 2019).

The spatial heterogeneity of the sedimentary carbon flux delivered to trenches is significant, as shown in Fig. 6. Each subducting slab follows unique pressure–temperature paths and sedimentation histories, which influence the efficiency of carbon subduction (Clift 2017; Plank and Manning 2019). The efficiency of carbon subduction is influenced primarily by the spatial proximity of deep-sea fans (rich in sedimentary OC) or shallow seafloor (rich in sedimentary carbonates) to the trench and the geological evolution of the subduction slab (Plank and Manning 2019), such as whether it was created by slow-spreading (promoting the development of carbonated serpentinites) or formed during the Cretaceous period (promoting the presence of carbonated oceanic crust) (Plank and Manning 2019). Areas such as Tonga, Izu-Bonin,

Table 1 Summary of global sedimentary carbon input fluxes

Method summary	Study area	Sedimentary C input flux (Mt C/year)	OC/(OC+IOC) (%)	References
Combining C content and isotope measurements with sediment thickness models	Sunda Margin	1	10–25	House et al. (2019)
Using historical data and tectonic models to quantify CO ₂ flux variations	Global	57		Dutkiewicz et al. (2019)
Developed a sediment thickness model and calculation	Global	60	20	Clift (2017)
Literature compilation and estimation	Global	13–17		Dasgupta and Hirschmann (2010)
Compilation of C content and calculation	Global	13	22	Hayes and Waldbauer (2006)
Analyzing C content in ODP samples to calculate CO ₂ flux	Costa Rica	2	9	Li and Bebout (2005)
Using DSDP/ODP data to estimate C content, porosity, and density	Global	14		Jarrard (2003)
Determined by sediment subduction rate and C content in the sediment	Main subduction	19		Hilton et al. (2002)
Derived by plate convergence rates, trench length, and sediment mass flux	Global	11		Plank and Langmuir (1998)
Estimated by calcite flux within the sediments	12 subductions	26		Rea and Ruff (1996)

DSDP Deep Sea Drilling Project, ODP Ocean Drilling Program

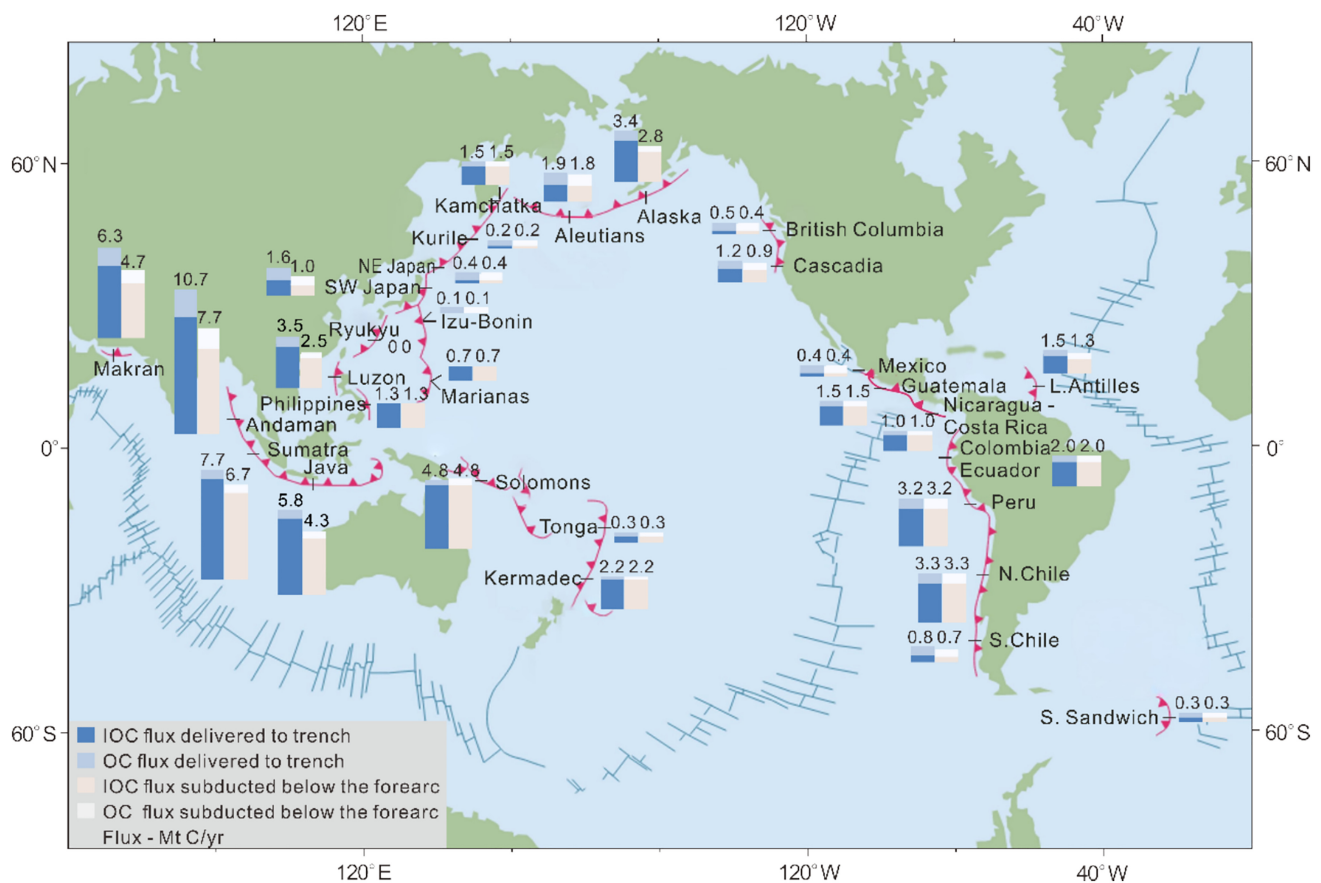


Fig. 6 Sedimentary carbon flux to trenches and subduction beneath the forearc in major subduction zones. The numbers represent the total carbon flux, including organic and inorganic carbon, in each trench (Clift 2017). Dark blue and light blue bars show inorganic and OC flux delivered to the trenches, while light red and white bars denote inorganic and OC flux subducted beneath the forearc

and the Kuril-Kamchatka trenches exhibit minimal carbonate input due to their deeper seafloor. In contrast, regions with higher biological productivity and shallower seafloors, such as the central South American subduction zone and the New Zealand trench, preserve significant amounts of carbonates (Clift 2017).

5.2 Carbon output flux

Carbon stored in Earth's interior is returned to the surface primarily through volcanism and associated degassing processes. Volcanic CO₂ emissions vary significantly across tectonic settings, with arc volcanism contributing approximately 12–74 Mt/year, mid-ocean ridge volcanism 6–60 Mt/year, and ocean island volcanism 1–3 Mt/year. Together, these variations yield an estimated total global volcanic CO₂ flux of 19–137 Mt/year (Fig. 7; Table 2). However, these estimates carry substantial uncertainty due to heterogeneous measurement techniques and incomplete spatial coverage.

In addition to direct volcanic emissions, diffuse degassing from fault systems and hydrothermal areas plays a critical

role in carbon output. These emissions are particularly significant in large silicic calderas with infrequent eruptions, where cumulative degassing over decades can reach levels comparable to the active volcanic eruptions' flux and approximately 17 Mt C/year (Werner et al. 2019). Diffuse sources (e.g., soil gas efflux, fractures, volcanic vents, and thermal springs) contribute significantly, especially in geothermal regions (Burton et al. 2013; Lee et al. 2016). The global carbon flux from hydrothermal and diffuse degassing is estimated at approximately 23 Mt C/year (Werner et al. 2019). The East African Rift (EAR) has emerged as a key region of structural carbon release. Lee et al. (2016) combined isotopic analysis with soil CO₂ flux measurements, estimating that degassing outside active volcanic centers in the EAR contributes 9–28 Mt C/year. Similarly, Hunt et al. (2017) estimated that carbon output from volcanic and geothermal sources in the EAR ranges between 1 and 9 Mt C/year.

Carbon output through both volcanic and non-volcanic pathways represents a substantial component of the global carbon cycle. Compared to sedimentary carbon subduction

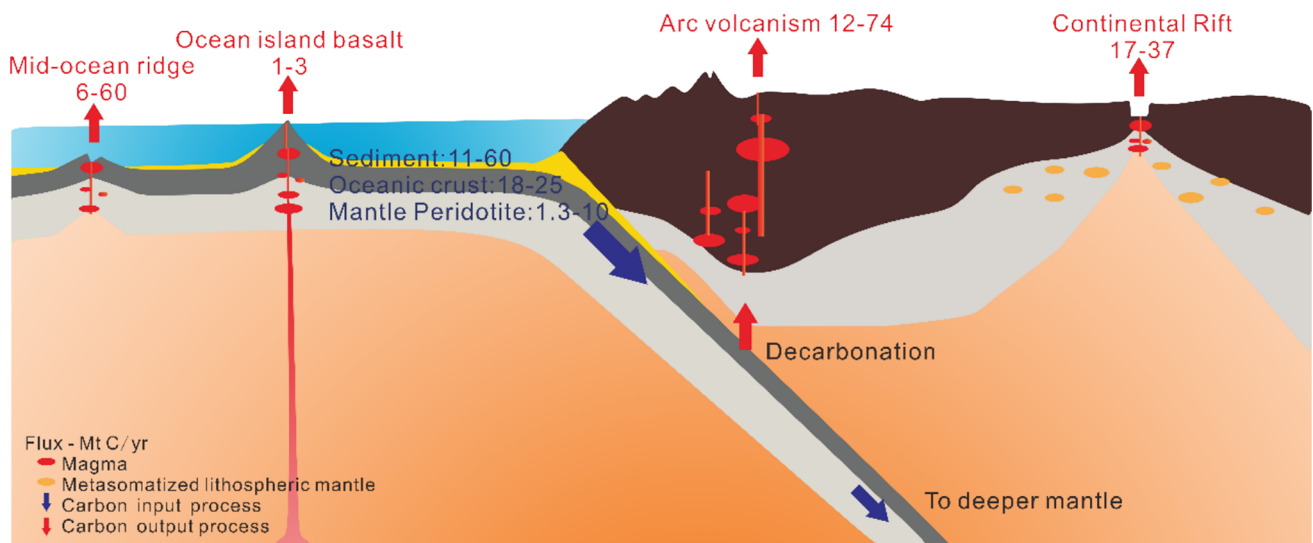


Fig. 7 Cartoon (not to scale) showing the global C cycle flux across different plate tectonic settings. The data sources are provided in Tables 1 and 2

fluxes, the net carbon balance between the Earth's surface and deep mantle remains uncertain and widely debated. Whether the Earth acts as a net source or sink of carbon over geological timescales is still unresolved. Accurate quantification of carbon output fluxes, particularly from diffuse and hydrothermal sources, alongside improved understanding of subduction processes and mantle wedge interactions, is essential for constraining the long-term global carbon budget.

6 Implications for global climate and mantle chemistry

The recycling of sedimentary carbon in subduction zones has profound implications for both the global climate system and the geochemical evolution of the mantle. Within the long-term carbon cycle, CO_2 is returned to the surface via volcanic and metamorphic degassing and through weathering of sedimentary carbonates, and is removed primarily through carbonate precipitation and burial of OC. OC buried in marine sediments acts as a net sink for atmospheric CO_2 and a long-term source of O_2 , significantly influencing atmospheric composition over geological timescales (Burdige 2007; Eguchi et al. 2019). For instance, deep-sea sediment records show that OC burial rates during glacial maxima were approximately 50% higher than during interglacials, underscoring the climate sensitivity of this carbon sink (Cartapanis et al. 2016). Furthermore, enhanced OC burial is thought to have driven the dramatic rise in atmospheric O_2 during the Great Oxidation Event (~2.5–2.2 Ga) (Lyons et al. 2014; Duncan and Dasgupta 2017), coinciding

with the Lomagundi positive $\delta^{13}\text{C}$ excursion in marine carbonates (Karhu and Holland 1996; Eguchi et al. 2019).

Geochemically, the recycling of sedimentary carbonates into the mantle alters the chemical and mineralogical characteristics of the lithospheric mantle through fluid–rock interaction. Subducted sediments release fluids rich in CO_2 , REEs, LILEs, and high-field-strength elements (HFSEs) (Plank and Langmuir 1998; Atlas et al. 2022), which metasomatize the overlying mantle wedge. These processes have been linked to the genesis of carbonatite-associated REE deposits (Hou et al. 2015), silica- and potassium-rich arc magmas (Skora and Blundy 2010), uranium excess in arc magmas (Avanzinelli et al. 2018), iron-rich melts (He et al. 2020a), and elevated high-platinum-group element (PGE) concentrations in carbonatite xenoliths (He et al. 2020b).

Due to the high solubility of calcite, dehydration fluids preferentially dissolve CaCO_3 at the sediment–peridotite interface, leading to the formation of clinopyroxene (Gervasoni et al. 2017). Petrology experiments have shown that even minor amounts of carbon can significantly modify the solidus of carbonated sediments. The addition of CO_2 to sedimentary rocks can either increase or decrease their melting temperatures, depending on bulk composition and the presence of alkali metals (see Sect. 3.3). Moreover, subducted sediments can increase the oxygen fugacity of the lithospheric mantle, as they typically possess higher oxygen fugacity than mid-ocean ridge or ocean island basalts (Mungall 2002). Carbonate minerals within the slab may transform into reduced species such as graphite or diamond, while simultaneously releasing oxidizing agents that elevate the redox state of the surrounding mantle (Frezzotti et al. 2011; Galvez et al. 2013).

Table 2 Summary of global carbon output fluxes

Volcanic degassing	Method	Study area	Output flux (Mt C/year)	References
Arc volcanism	Literature compilation and estimation	Global	14–24	Fischer and Aiuppa (2020)
	Based on the CO ₂ /SO ₂ ratio and SO ₂ emission rate	Global	12–16	Fischer et al. (2019)
	Direct measurement or using CO ₂ /SO ₂ ratio and SO ₂ emission rate	102 volcanoes	12	Werner et al. (2019)
	Based on the CO ₂ / ³ He ratio and ³ He emission rate	Global	26	Kagoshima et al. (2015)
	Direct measurement or using CO ₂ /SO ₂ ratio and SO ₂ emission rate	Global	74	Burton et al. (2013)
	Extrapolated from Japan Arc data to a global scale	Global	14	Shinohara (2013)
	Based on the CO ₂ /SO ₂ ratio and SO ₂ emission rate	Global	23	Fischer (2008)
	Estimated SO ₂ flux and combined it with compiled CO ₂ /S data	Global	19	Hilton et al. (2002)
	Literature compilation and estimation	Global	22	Mörner and Etiope (2002)
Mid-ocean ridge volcanism	Literature compilation and estimation	Global	16	Orcutt et al. (2019)
	CO ₂ /Ba and CO ₂ /Rb ratios	Global	6–25	Le Voyer et al. (2019)
	CO ₂ / ³ He ratio and ³ He flux	Main mid-ocean ridge	13–19	Tucker et al. (2018)
	CO ₂ /ITE (Nb, Th, Rb, Ba) ratios and ITE content	Global	18–40	Hauri et al. (2018)
	CO ₂ /Rb, CO ₂ /Ba, CO ₂ /Nb ratios and ITE content	Global	13–30	Le Voyer et al. (2017)
	CO ₂ /Ba ratio and Ba content	Global	25–44	Michael and Graham (2015)
	CO ₂ /Nb ratio and Nb content	Global	12–60	Cartigny et al. (2008)
	CO ₂ /Nb ratio and Nb content	Global	12	Hayes and Waldbauer (2006)
	CO ₂ / ³ He ratio and ³ He flux	Global	6–24	Resing et al. (2004)
Ocean island volcanism	Literature compilation and estimation	Hot springs	1	Orcutt et al. (2019)
	CO ₂ / ³ He ratio	Hawaii	1	Poland et al. (2014) and Tucker et al. (2019)
	CO ₂ , δ ¹³ C, H ₂ O, and δD data	Iceland	0–3	Hartley et al. (2014) and Barry et al. (2014)
	CO ₂ , δ ¹³ C, H ₂ O, and δD data	Pitcairn	0.02	Aubaud et al. (2006)
	CO ₂ , δ ¹³ C, H ₂ O, and δD data	Society	0.03	Aubaud et al. (2005)
Diffusive degassing	From dormant volcano degassing	Global	17	Werner et al. (2019)
	From rift volcano degassing	EAR	1–9	Hunt et al. (2017)
	Measured the CO ₂ and δ ¹³ C content in the soil	EAR	9–28	Lee et al. (2016)
	From soil and groundwater degassing	Global	32	Burton et al. (2013)
	Extrapolated from 32 volcanic lakes data to a global scale	Global	27–37	Pérez et al. (2011) and Burton et al. (2013)

EAR East African Rift, ITE incompatible trace elements

7 Conclusions and outlook

The deep carbon cycle associated with sediments can be summarized as follows:

1. Sedimentary organic carbon burial in subduction zones is governed by primary production, sediment transport, and bottom-water oxygen concentration. These factors collectively influence carbon burial efficiency and

strongly affect long-term carbon sequestration. The proportion of organic carbon subducted below the forearc varies between 10% and 25%, with significant spatial and temporal variability over geological timescales.

2. Carbon in the mantle exists primarily as accessory phases, and its stability is mainly governed by temperature, pressure, and oxygen fugacity. In the shallow mantle, carbonates as the primary carbon species can typically evolve from Ca-rich to Mg-rich with depth. At greater depths, carbon can form reduced forms (e.g., graphite, diamonds, and carbides). Although carbonates may persist in localized oxidizing domains of the deep mantle, their global distribution and stability remain uncertain.
3. Carbon in subducted sediments is released through metamorphism, fluid-driven decarbonation, partial melting, and diapirism, then migrates into the mantle within fluid or melt. Metamorphism dominates in hotter zones, while fluid-driven decarbonation prevails in cooler ones. In some cases, the combination of metamorphism and dissolution leads to higher decarbonation efficiency. Diapirism can remove up to 80% of carbon from the subducting slab in certain regions while limiting carbon transfer to the deep mantle.
4. Marbles containing microdiamond inclusions and other UHP metamorphic rocks offer direct evidence for sedimentary carbonate subduction into the deep mantle. Carbon and metal isotopes (e.g., low $\delta^{13}\text{C}$ values, low $\delta^{26}\text{Mg}$ values, low $\delta^{44}/^{40}\text{Ca}$ values) can distinguish sedimentary carbon from mantle carbon, revealing organic carbon recycling and diamond formation. However, similar signatures can result from fractionation or magma degassing; therefore, a multi-indicator analysis is essential for accurate interpretation.
5. Although sedimentary carbon contributes approximately two-thirds of the global carbon input at the trenches, with the values ranging from 57 to 60 Mt/year, its input flux in subduction zones and volcanic CO_2 emissions varies considerably due to tectonic setting differences and measurement technique limitations. As a result, deep carbon budgets remain uncertain, and further studies are needed.
6. Subduction of sedimentary carbon influences global climate by regulating atmospheric CO_2 and O_2 levels over geological timescales, and affects mantle chemistry by modifying melting temperatures, composition, and oxygen fugacity conditions.

Despite the considerable progress in research on the subduction sedimentary carbon cycle, several key issues have yet to be completely solved:

1. Role of diapirism. The contribution of diapirs to deep carbon transport remains uncertain, necessitating further geophysical imaging and high-pressure experimental studies to clarify their role in mantle carbon cycling.
2. Geochemical tracing. Isotopic signatures coupled with trace element analysis can improve the identification of subducted carbon sources, facilitating our understanding of carbon recycling in the mantle.
3. The estimates of carbon flux. Advancing sampling and measurement techniques is essential to reducing uncertainties in sedimentary carbon input and volcanic CO_2 emissions and improving global carbon budget assessments.

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Data availability Not applicable.

Declarations

Conflict of interest On behalf of all authors, the corresponding authors state that they have no conflict of interest.

Ethical approval The authors declare that they have followed all the ethical rules, including those listed on the Acta Geochimica website.

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