Probing the exchange of CO\textsubscript{2} and O\textsubscript{2} in the shallow critical zone during weathering of marl and black shale

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Abstract. Chemical weathering of sedimentary rocks can release carbon dioxide (CO\textsubscript{2}) and consume oxygen (O\textsubscript{2}) via the oxidation of petrogenic organic carbon and sulfide minerals. These pathways govern Earth’s surface system and climate over geological timescales, but the present-day weathering fluxes and their environmental controls are only partly constrained due to a lack of in situ measurements. Here, we investigate the gaseous exchange of CO\textsubscript{2} and O\textsubscript{2} during the oxidative weathering of black shales and marls exposed in the French southern Alps. On six field trips over 1 year, we use drilled headspace chambers to measure the CO\textsubscript{2} concentrations in the shallow critical zone and quantify CO\textsubscript{2} fluxes in real time. Importantly, we develop a new approach to estimate the volume of rock that contributes CO\textsubscript{2} to a chamber, and assess effective diffusive gas exchange, by first quantifying the mass of CO\textsubscript{2} that is stored in a chamber and connected rock pores. Both rock types are characterized by similar contributing rock volumes and diffusive movement of CO\textsubscript{2}. However, CO\textsubscript{2} emissions differed between the rock types, with yields over rock outcrop surfaces (inferred from the contributing rock volume and the local weathering depths) ranging on average between 73 and 1108 t C km\textsuperscript{-2} yr\textsuperscript{-1} for black shales and between 43 and 873 t C km\textsuperscript{-2} yr\textsuperscript{-1} for marls over the study period. Having quantified diffusive processes, chamber-based O\textsubscript{2} concentration measurements are used to calculate O\textsubscript{2} fluxes. The rate of O\textsubscript{2} consumption increased with production of CO\textsubscript{2}, and with increased temperature, with an average O\textsubscript{2}:CO\textsubscript{2} molar ratio of 10:1. If O\textsubscript{2} consumption occurs by both rock organic carbon oxidation and carbonate dissolution coupled to sulfide oxidation, either an additional O\textsubscript{2} sink needs to be identified or significant export of dissolved inorganic carbon occurs from the weathering zone. Together, our findings refine the tools we have to probe CO\textsubscript{2} and O\textsubscript{2} exchange in rocks at Earth’s surface and shed new light on CO\textsubscript{2} and O\textsubscript{2} fluxes, their drivers, and the fate of rock-derived carbon.
1 Introduction

Sedimentary rocks cover ~64% of the present-day continental surface of Earth (Hartmann and Moosdorf, 2012) and contain vast amounts of carbon in carbonate minerals and organic matter (Petsch, 2014). The chemical breakdown of these rocks can act as a source of carbon dioxide (CO$_2$) to the near-surface reservoirs (hydroosphere–biosphere–pedosphere–atmosphere) and can be a sink of oxygen (O$_2$), in turn, exerting an important control on the evolution of climate and life (Berner and Berner, 2012; Berner, 1999; Sundquist and Visser, 2003). Two reactions are recognized, namely (i) the oxidation of petrogenic organic carbon (OC$_{petro}$) (Petsch, 2014) and (ii) the oxidation of sedimentary sulfide minerals that produce sulfuric acid that can, in turn, dissolve carbonate minerals (Calmels et al., 2007; Li et al., 2008; Torres et al., 2014). On a global scale, these chemical weathering pathways together emit roughly as much CO$_2$ to the atmosphere (Berner and Berner, 2012; Burke et al., 2018; Petsch, 2014; Zondervan et al., 2023) as is removed by the weathering of silicate minerals with a flux of ~90–140 Mt C yr$^{-1}$ (Gaillardet et al., 1999; Moon et al., 2014). How these fluxes play out over longer timescales remains difficult to assess (Petsch, 2014; Hilton and West, 2020); however, the decline in the atmospheric O$_2$ over the last 800,000 years (Stolper et al., 2016) has been tentatively linked to changes in global oxidative weathering fluxes (Yan et al., 2021). To improve the understanding of the changes in the Earth’s surface conditions over geological timescales, the mechanism and controls on oxidative weathering pathways need to be better constrained (Berner and Berner, 2012; Mills et al., 2021). Theoretical modeling of OC$_{petro}$ oxidation currently relies on input kinetics of the weathering reactions from laboratory experiments (Bolton et al., 2006; Petsch, 2014; Bao et al., 2017). In situ gas exchange between rocks undergoing weathering and the atmosphere can provide much needed insight.

The first field-based fluxes of weathering-derived CO$_2$ were reported by Keller and Bacon (1998) in a glacial till dominated by shales. More recently, Tune et al. (2020) found substantial CO$_2$ release and O$_2$ depletion in bedrock undergoing weathering below a forested hillslope. There, according to monitoring of gases and water chemistry, carbon release is mostly sourced from superficial soils and deep roots, with minor contributions from OC$_{petro}$ oxidation (Tune et al., 2020, 2023). At both sites, the gaseous fluxes were determined on the basis of profiles of the CO$_2$ concentration in air sampled from boreholes extending to depths of ~7 m (Keller and Bacon, 1998) and of ~16 m (Tune et al., 2020), using Fick’s law:

$$J_X = -D_X \times \frac{dC_X}{dz},$$

(1)

where $J_X$ is the molar flux (mol m$^{-2}$ s$^{-1}$) of the particular gas species X and $D_X$ its diffusivity (i.e., the capability to allow diffusion; m$^2$s$^{-1}$) in the studied vadose zone, and where $dC_X/dz$ describes the change in the concentration (mol m$^{-3}$) over depth (m). An alternative approach introduced by Soulet et al. (2018) uses gas accumulation chambers drilled into shallow weathering zones. Instead of calculating a carbon flux from a presupposed diffusion coefficient, which can introduce uncertainties (Maier and Schack-Kirchner, 2014), CO$_2$ release is measured in real time in a similar way that is commonly applied to soil surfaces (Oertel et al., 2016). This method has provided new insight into how temperature, precipitation, and topography control CO$_2$ emissions from marls (Soulet et al., 2021) and mudstones (Roylands et al., 2022). However, for weathering chambers that are installed within the rock face, three aspects remain unexplored: (i) the rock volume that contributes to the CO$_2$ accumulation measured in the chamber; (ii) how the diffusive movement of CO$_2$ in the shallow weathering zone is impacted by short-term environmental changes (e.g., in temperature and hydrology); and (iii) the quantification of O$_2$ depletion during oxidative weathering.

In this study, we investigate the weathering-driven exchange of CO$_2$ and O$_2$ by installing chambers into black shales and marls undergoing oxidation at two study sites in the steep terrains of the Draix-Bléone Observatory, France (Gaillardet et al., 2018; Draix-Bléone Observatory, 2015; Klotz et al., 2023). Building on research from an outcrop at the same observatory (Soulet et al., 2021), we find that chamber-based CO$_2$ emissions vary significantly over 1 year, linked to changes in temperature and precipitation. A new theoretical framework is developed to refine CO$_2$ flux measurements (Section 4.1) and applied to quantify the rock pore space that is probed during a measurement (Section 4.2). This allows us to normalize CO$_2$ accumulation rates based on a contributing rock volume and return estimates of fluxes emitted from the surface of rock outcrops. The resulting CO$_2$ fluxes can be accurately described as a function of temperature (Section 4.3). Using Fick’s law, measurements of the O$_2$ concentration in the chambers are then used to quantify the O$_2$ consumption in the studied rocks (Section 4.4). Together, we provide new insights into the exchange of CO$_2$ and O$_2$ in the shallow weathering zone of sedimentary rocks.

2 Material and methods

2.1 Study area

The two study sites are located in the catchments of the Brusquet (area of 1.08 km$^2$) and Moulin (0.08 km$^2$) of the Draix-Bléone Observatory (Draix-Bléone Observatory, 2015; Klotz et al., 2023), which is part of the French network of critical zone observatories (OZCAR) (Gaillardet et al., 2018). The Brusquet site is located at 44.16251° N, 6.32330° E at 847 m a.s.l. and the Moulin site at 44.14146° N, 6.36095° E at 874 m a.s.l. (Fig. 1). The catchments of the Draix-Bléone Observatory have detailed measurements of river water dis-
Figure 1. Location of the French Draix-Bléone Observatory and of the study sites for in situ CO\textsubscript{2} and O\textsubscript{2} monitoring in the Brusquet catchment (red circle) and in the Moulin catchment (blue circle), alongside the location of previous research in the Laval catchment for reference (yellow circle) (Soulet et al., 2018, 2021) and geological maps of these catchments (Janjou, 2004). Meteorological stations are present at each of the catchment outlets, with a maximum distance to the study sites of 200 m (Draix-Bléone Observatory, 2015). Catchment-specific aerial imagery (Draix-Bléone Observatory, 2015) is shown alongside transparent aerial imagery of the wider area (2018 ©IGN).

charge, river suspended load and bedload transport, and meteorological data over the last 4 decades (Cras et al., 2007; Mathys et al., 2003; Carriere et al., 2020; Draix-Bléone Observatory, 2015; Mathys and Klotz, 2008; Mallet et al., 2020; Klotz et al., 2023). Prior work has examined the occurrence of the solid phase of OC\textsubscript{petro} in the Brusquet, Moulin, and Laval catchments (Graz et al., 2011, 2012; Copard et al., 2006). The Laval catchment (0.86 km\textsuperscript{2}), which neighbors the Moulin catchment (Fig. 1), is the location of previous in situ measurements of rock-derived CO\textsubscript{2} (Soulet et al., 2018, 2021).

The Moulin catchment overlays Callovian to Oxfordian marls (Mathys et al., 2003; Graz et al., 2012; Janjou, 2004). In contrast, the lithology of the Brusquet catchment consists of a sequence of Bajocian marly limestones, Aalenian marls, and limestones to Toarcian black shales (Janjou, 2004; Copard et al., 2006; Graz et al., 2011), with the study site located on the latter (Fig. 1).

The climate is transitional between Alpine and Mediterranean, with a hot and dry summer, including short and intense rainfall events during thunderstorms (up to 150 mm h\textsuperscript{−1}), with rainfall events of lower intensity typically during spring and autumn (Soulet et al., 2021; Carriere et al., 2020; Mathys et al., 2003; Mallet et al., 2020). During winter, more than 100 d of frost can occur (Oostwoud Wijdenes and Ergenzinger, 1998; Rovera and Robert, 2006), and frost cracking from ice segregation was found to control hillslope regolith production (Ariagno et al., 2022). The mean annual rainfall is \( \sim 900 \) mm, and the mean annual air temperature is \( \sim 11 \) °C, defined by high solar radiation (\( > 2300 \) h yr\textsuperscript{−1}) (Soulet et al., 2021; Mallet et al., 2020; Mathys and Klotz, 2008).

Together, the climate and the erodible lithology of finely bedded, mechanically weak rocks result in a badland morphology, with V-shaped gullies, high physical weathering rates, and abrupt, sediment-loaded floods (Antoine et al., 1995; Cras et al., 2007; Le Bouteiller et al., 2021; Mathys et al., 2003). These features limit the development of soils and vegetation cover. In the late 19th century, following overgrazing in the wider area of the observatory, the Brus-
quet catchment was reforested (Mathys et al., 2003; Cras et al., 2007). Today ~87% of the catchment area of Brusquet is vegetated, in contrast to ~46% and ~32% of the Moulin and Laval catchments, respectively (Carriere et al., 2020; Cras et al., 2007). The sediment export fluxes of the Brusquet catchment are on average ~70, ~5700 and ~14 300 t km$^{-2}$ yr$^{-1}$ for the Moulin and Laval catchments, respectively (Mathys et al., 2003; Carriere et al., 2020).

Taking a regolith bulk density of ~1.3–1.8 t m$^{-3}$ into account (Mallet et al., 2020; Mathys and Klotz, 2008; Oostwoud Wijdenes and Ergenzinger, 1998; Bechet et al., 2016; Ariagno et al., 2023), a physical erosion rate of ~0.04–0.05 mm yr$^{-1}$, ~3.2–4.4 and ~8–11 mm yr$^{-1}$ can be estimated for the Brusquet, Moulin, and Laval catchments, respectively. However, these values are catchment-scale averages, and the physical erosion can significantly vary spatially.

On steep, bare slopes, the erosion rates may be comparably high in the different catchments of the Draix-Bléone Observatory (Carriere et al., 2020; Bechet et al., 2016; Mathys et al., 2003).

The bare surfaces in the catchments are characterized by four morphologically different layers, as reviewed by Mathys and Klotz (2008): (i) near-surface loose detrital cover of locally produced clasts or colluvial material with a thickness of ~0–10 cm; (ii) the underlying upper fine regolith with a thickness of ~5–20 cm; and (iii) the lower, coarse, and compact regolith, with a thickness of ~10–20 cm; and (iv) the unweathered bedrock at the bottom (Oostwoud Wijdenes and Ergenzinger, 1998; Maquaire et al., 2002; Rovéra and Robert, 2006). The compactness and density of these layers increase, while the porosity decreases (from values of up to ~50%) (Bechet et al., 2016; Mallet et al., 2020; Garel et al., 2012), over depth towards the unweathered bedrock (Maquaire et al., 2002). The unweathered bedrock has a grain density of ~2.7 t m$^{-3}$ and a porosity of ~10%–20% (Lofi et al., 2012). The thickness of the weathering profile varies laterally, with the thickest regolith layers and detrital cover on crests, minimal development in thalwegs, and intermediate in gullies (Maquaire et al., 2002; Esteves et al., 2005).

2.2 Drilled gas accumulation chambers

To measure in situ the production of CO$_2$ and consumption of O$_2$ by oxidative reactions in the shallow critical zone of sedimentary rocks undergoing weathering, we use drilled chambers. The chambers were visited six times over the study (to capture seasonal changes in weather conditions) on 27 September 2018, 11–14 January 2019, 11–15 April 2019, 27–29 May 2019, 5–12 July 2019, and 27 September to 2 October 2019. Their design has been previously detailed (Soulet et al., 2018). In summary, a horizontal chamber is drilled directly into the exposed rock, which has been cleared of detrital cover. The shape of the drilled chambers ensures a large surface-to-volume ratio to benefit measurement of gas concentrations and potential trapping of CO$_2$. To install the chambers to a depth of ~38 cm (Fig. 2e), a mechanical drill was used, with a diameter of 2.9 cm. Rock powder left inside the chambers was blown away with a compressed-air gun. A small PVC (polyvinyl chloride) tube was inserted in the entrance of each chamber that was closed with a rubber stopper holding two glass tubes fitted with Tygon tubing. The latter allow either connection to a gas-sampling system or sealing with WeLoc® clips. To further isolate the chamber from the atmosphere, the intersection of the PVC tube and regolith was sealed with a silicone sealant (UniBond® Outdoor Window and Door Frame), which we previously tested to be free of potential contaminants for gas sampling (Roylands et al., 2022).

At both the Brusquet and Moulin study sites, we installed one array of four chambers placed in a square ($2 \times 2$) (Fig. 2) (Table 1). In each array, two chambers were placed in the same rock bed with a roughly horizontal orientation at the Brusquet site and roughly vertical at the Moulin site. The minimum distance between chambers was 70 cm. The aspect, hydrological, and geomorphic setting of the location of both arrays is similar; they were placed at the upper margin of the watersheds in steep walls of gullies on a southwest- (Brusquet) and south-facing aspect (Moulin). The chambers were drilled into bare rock faces devoid of roots and with minimal soil or vegetation cover in the vicinity to exclude a contribution by them to the CO$_2$ measurements (Fig. 2).

2.3 Rock temperature and humidity

At both sites, a chamber with the same design was installed to hold a temperature and relative humidity logger (Lascar® EL-USB-2) (Fig. 2) from 27 September 2018 onwards. A second logger was placed on the rock surface (monitoring the air directly above it), with the main body fitted inside a small PVC tube for physical protection and an aluminum foil wrapping around the tube to avoid the alteration of the temperature measurement due to the dark color of the housing of the sensor and the PVC tube.

Over the study period, technical issues with batteries of the temperature and relative humidity loggers prevented continuous data collection. To fill the gaps in the direct chamber temperature measurements, we use air temperatures from a local weather station as a proxy (Appendix A) by modifying a framework that describes soil temperatures by, amongst other variables, air temperature (Liang et al., 2014). Using the site-specific air temperatures, this approach simulates the chamber temperature well, with a root mean square error (RMSE) of 1.8 °C for the Brusquet catchment and 2.2 °C for the Moulin catchment (Appendix A).

2.4 Partial pressure of rock CO$_2$

The partial pressure of CO$_2$ ($p$CO$_2$) was measured alongside air pressure to determine the concentrations of CO$_2$ in the rock chambers with an infrared gas analyzer (IRGA; EGM-5
Figure 2. The study sites in the Brusquet catchment (a, b) and in the Moulin catchment (c, d). Identifiers of the CO$_2$ accumulation chambers (Table 1) are given next to their entrance. Furthermore, the location of temperature and relative humidity loggers in a further chamber with similar properties (circled dot) and on the rock surface (circled X) are shown. For scale, the gray cases (not used in the present study) next to the chambers are of $\sim$ 40 cm length. Panel (e) shows the design of the chambers and the sampling system adapted from Soulet et al. (2018).

Table 1. Characteristics of gas accumulation chambers drilled into weathering sedimentary rocks in the Brusquet catchment and in the Moulin catchment. For the calculation of the volume and inner surface area of the chambers, the length and insertion depth of the PVC tube and rubber stopper are used.

<table>
<thead>
<tr>
<th>Chamber identifier</th>
<th>Site</th>
<th>Same bed as installation</th>
<th>Installation date</th>
<th>Depth (cm)</th>
<th>Volume (cm$^3$)</th>
<th>Inner surface area (cm$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1 M-C-1</td>
<td>Moulin</td>
<td>3</td>
<td>24 Sep 2018</td>
<td>41.0</td>
<td>278</td>
<td>367</td>
</tr>
<tr>
<td>2 M-A-2</td>
<td>Moulin</td>
<td>4</td>
<td>24 Sep 2018</td>
<td>39.5</td>
<td>263</td>
<td>346</td>
</tr>
<tr>
<td>3 M-D-3</td>
<td>Moulin</td>
<td>1</td>
<td>24 Sep 2018</td>
<td>38.0</td>
<td>255</td>
<td>335</td>
</tr>
<tr>
<td>4 M-B-4</td>
<td>Moulin</td>
<td>2</td>
<td>24 Sep 2018</td>
<td>39.0</td>
<td>262</td>
<td>345</td>
</tr>
<tr>
<td>5 B-F-5</td>
<td>Brusquet</td>
<td>6</td>
<td>25 Sep 2018</td>
<td>38.0</td>
<td>255</td>
<td>335</td>
</tr>
<tr>
<td>6 B-G-6</td>
<td>Brusquet</td>
<td>5</td>
<td>25 Sep 2018</td>
<td>37.0</td>
<td>255</td>
<td>335</td>
</tr>
<tr>
<td>7 B-H-7</td>
<td>Brusquet</td>
<td>8</td>
<td>25 Sep 2018</td>
<td>35.0</td>
<td>229</td>
<td>299</td>
</tr>
<tr>
<td>8 B-I-8</td>
<td>Brusquet</td>
<td>7</td>
<td>25 Sep 2018</td>
<td>35.0</td>
<td>235</td>
<td>308</td>
</tr>
</tbody>
</table>

Portable Gas Analyzer, PP Systems). This is equipped with an internal pump and calibrated to $p$CO$_2$ in the range of 0 to 30 000 ppmv (parts per million by volume). First, the closed-loop sampling system is purged of CO$_2$ using an inline CO$_2$ scrub (soda lime) (Hardie et al., 2005). This is then connected to a chamber to measure the ambient $p$CO$_2$. After a short equilibration, the $p$CO$_2$ in the chamber ($p$CO$_2$ Chamber) is calculated from the CO$_2$ concentration in the combined air volume of the chamber and the sampling system by accounting for the dilution introduced from the CO$_2$-free air that was originally contained within the sampling system (Soulet et al., 2018; Roylands et al., 2022). To ensure that the determined $p$CO$_2$ Chamber is representative of the ambient $p$CO$_2$ in the rock pores around the chamber ($p$CO$_2$ Rock), measurements are only considered if the chambers were left closed overnight so that the production of CO$_2$ from oxidative weathering could reach a steady state with respect to diffusion between rock pores, chamber, and atmosphere.
2.5 CO₂ flux measurements

Real-time measurements of CO₂ release in drilled chambers have been previously described in detail (Soulet et al., 2018) and used to quantify the CO₂ flux (Soulet et al., 2021; Roylands et al., 2022). In summary, one CO₂ flux measurement consists of a series of repeated accumulations (typically eight or more) that are recorded over time after determining the pCO₂ Rock. First, the pCO₂ Chamber is lowered to a value close to the local atmosphere value using soda lime or a zeolite sieve (Sect. 2.6) (Fig. 2e). Then, CO₂ is allowed to build up, typically over ∼6 min, before the CO₂ in the chamber is again removed to a near-atmospheric value. For each repeat, the rate of CO₂ accumulation $q$ (mg C min$^{-1}$) is calculated by fitting an exponential model to the recorded pCO₂ change, following Pirk et al. (2016):

$$\frac{dm(t)}{dt} = q - \lambda(m(t) - m_0),$$  

where $\frac{dm(t)}{dt}$ is the carbon mass change (mg C) in the chamber with time (min), $m_0$ is the initial carbon mass in the chamber (mg C), and $\lambda$ (min$^{-1}$) is a constant covering the sum of all processes proportional to the carbon mass difference $(m(t) - m_0)$ and relating to the diffusion of CO₂ between rock pores, chamber, and atmosphere (Soulet et al., 2018). For this, the carbon mass ($m$; mg C) in the chamber is obtained from

$$m(t) = pCO₂_{Chamber}(t) \times V \times \frac{P}{R \times T} \times M_C \times 10^{-9},$$  

where the measured pCO₂ Chamber is in ppmv (cm$^3$ m$^{-3}$), $V$ is the combined volume (cm$^3$) of the chamber and the sampling system, $P$ is the pressure (Pa), $R$ is the universal gas constant (m$^3$ Pa K$^{-1}$ mol$^{-1}$), $T$ is the chamber temperature (K), and $M_C$ is the molar mass of carbon (g mol$^{-1}$).

The CO₂ accumulation rate $q$ can be normalized to the internal surface area of the chamber $S_{Chamber}$ (i.e., area of exchange with the surrounding rock; m$^2$) to account for differences in the depth of the chambers, which are related to differences in volume and surface area, giving a repeat-specific flux $Q$ (mg C min$^{-1}$ m$^{-2}$):

$$Q = \frac{q}{S_{Chamber}}.$$  

Alternatively, the CO₂ accumulation can be reported as a molar-based flux $J_{CO₂}$ (mmol CO₂ min$^{-1}$ m$^{-2}$):

$$J_{CO₂} = \frac{j_{CO₂}}{S_{Chamber}} = \frac{q}{M_C \times S_{Chamber}},$$  

where $j_{CO₂}$ (mmol CO₂ min$^{-1}$) is the molar-based analogue to $q$.

Previous work has noted that the CO₂ accumulation rate during the first measurements is typically higher than subsequent repeats (Soulet et al., 2018). To calculate a CO₂ flux from these repeated accumulations (consisting of $n$ repeats), previous work excluded the first three repeats ($q_1$ to $q_3$) and took the average of a minimum of three further repeats ($q_4$ to $q_{n\geq6}$) (Roylands et al., 2022; Soulet et al., 2018, 2021). We examine this further using new data collected here, which also provide constraint on the nature of the gas exchange around the chambers.

2.6 CO₂ sampling

During a CO₂ flux measurement, the CO₂ in the chamber can be sampled by circulating it through a zeolite molecular sieve cartridge (MSC) mounted in parallel to the monitoring line (Hardie et al., 2005; Soulet et al., 2018) (Fig. 2e). The volume of carbon loaded onto a sieve is estimated by adding up the pCO₂ maxima for each trapping episode minus the final pCO₂ after trapping (near-atmospheric value), while accounting for the combined volume of the chamber and the sampling system.

The zeolite sieves were heated in the laboratory to 425 °C and purged with high-purity nitrogen gas to release the CO₂ trapped onto them prior to cryogenic purification under vacuum (Garnett and Murray, 2013). The estimated sampled volume of CO₂ from the chamber-based pCO₂ measurements ($V_{CO₂-IRGA}$; mL) can be compared with the volume recovered from the MSC in the laboratory ($V_{CO₂-MSC}$; mL), giving a sampling ratio (SR; unitless) (Roylands et al., 2022):

$$SR = \frac{V_{CO₂-MSC}}{V_{CO₂-IRGA}}.$$  

For this, all volumes of CO₂ are normalized to 0 °C and 1013 mbar. The SR thus allows us to independently check calculations of carbon mass using the pCO₂ data combined with the gas line and chamber volume measurements (Eqs. 2 and 3).

2.7 Measuring pO₂ and O₂ fluxes

While measuring pCO₂, the EGM-5 portable gas analyzer, incorporating the IRGA, also records the partial pressure of oxygen in the chamber ($pO₂_{Chamber}$; % v/v) with an electrochemical O₂ sensor. The $pO₂_{Chamber}$ cannot be used in the same way as the pCO₂ to quantify flux for two reasons: (i) the precision of the O₂ sensor of ≥ 0.1 % (v/v) is insufficient to observe real-time changes in $pO₂_{Chamber}$ due to the expected weathering fluxes; and (ii) O₂ should be consumed during oxidative weathering, and so we would require a method that replenishes oxygen, which was not done while measuring CO₂ accumulation.

An alternative method to calculate an O₂ flux is based on fick’s law (Eq. 1), using the diffusive gradient of the partial pressure of O₂ in the rock ($pO₂_{Rock}$) towards the atmospheric O₂ concentration ($pO₂_{Atm}$). To obtain $pO₂_{Rock}$, the $pO₂_{Chamber}$ measured after connecting to the chamber is corrected for the oxygen concentration in the sampling system.
The $p_{O_2}$ recorded before and during a measurement is corrected for instrument drift. The drift correction is based on measuring the $p_{O_2\text{ Atm}}$ directly before or after a chamber-based measurement and assuming an average atmospheric oxygen concentration of 20.95% ($v/v$).

To quantify the exchange of $O_2$ between the chamber, connected rock pores, and the atmosphere ($j_{O_2}$; mmol$O_2$ min$^{-1}$), we describe the process via a diffusive transfer controlled by the diffusivity of $O_2$ ($D_{O_2}$; cm$^2$ min$^{-1}$) across a spatial parameter $\omega$ (describing in combination the movement across area over depth; cm$^3$ cm$^{-2}$):

$$j_{O_2} = \frac{D_{O_2}}{\omega} \times (p_{O_2\text{ Rock}} - p_{O_2\text{ Atm}}) \times \frac{P}{R \times T} \times 10^{-3}.$$ (7)

If we assume that $\omega$ is the same for $O_2$ and $CO_2$, linking the space of $O_2$ consumption and $CO_2$ release, then $D_{O_2}$ can be related to the diffusivity of $CO_2$ ($D_{CO_2}$; cm$^2$ min$^{-1}$), based on their ideal relation in air, which is independent of temperature (Angert et al., 2015):

$$\frac{D_{CO_2}}{D_{O_2}} = 0.76 \times \frac{D_{(CO_2)}{\omega}}{D_{(O_2)}{\omega}}.$$ (8)

Differences in the effective diffusivities of gas species depend on the structure of the air-filled pore space, which is expected to have identical impacts on the gaseous movement of $O_2$ and $CO_2$ (Angert et al., 2015; Millington, 1959; Penman, 1940). Thus, if the term $\frac{D_{CO_2}}{D_{O_2}}$ (cm$^3$ min$^{-1}$) can be quantified by other means (for instance, through analysis of the $p_{CO_2\text{ Rock}}$ and $CO_2$ flux data), then we can quantify $j_{O_2}$. These themes will be discussed later (Sect. 4.1 and 4.4).

3 Results

3.1 Chamber temperature and meteorological conditions

Over the study period (27 September 2018 to 2 October 2019), similar variability in environmental conditions was recorded at the Brusquet and Moulin catchments. The temperatures of the atmosphere, chamber interiors, and at the rock surface showed daily and seasonal changes (Table 2) (Fig. 3a–f). Rainfall events were comparable in occurrence and extent, but the Brusquet catchment received less cumulative rainfall (773 mm) than the Moulin catchment (1033 mm) (Fig. 3g and h). The relative air humidity in the chambers was high and constant, with values of $\sim$93.1 ± 4.5% (± standard deviation, SD) and $\sim$91.3 ± 4.1% at the Brusquet and Moulin sites, respectively (not considering gaps in the record) (Fig. 3i and j).

3.2 $p_{CO_2}$ measurements and $CO_2$ collection

The $p_{CO_2\text{ Rock}}$ values varied between the chambers and over time. In the Brusquet catchment, the observed $p_{CO_2\text{ Rock}}$ values were on average 1490 ± 743 ppmv (± SD, if not reported otherwise, $n = 28$), and 1492 ± 633 ppmv ($n = 32$) in the Moulin catchment (Table 3).

Following the determination of $p_{CO_2\text{ Rock}}$, a total of 37 $CO_2$ flux measurements were conducted in the Brusquet catchment, of which 32 consisted of ≥ 8 repeats. In the Moulin catchment, 41 measurements were made, with 37 having ≥ 8 repeats. Every individual $CO_2$ flux measurement showed an initial decline in the accumulation rates that approached a constant value of peak $CO_2$ concentration (Fig. 4). Considering the repeats 6–8, averages of the $CO_2$ accumulation rates varied between chambers and over time at each single chamber, with the occurrence of the lowest accumulation rates in winter and highest in summer. On four visits, a chamber was measured twice a day, and the observed $CO_2$ release was higher in the afternoon than in the morning, coinciding with an increase in the chamber temperature. Overall, the observed $CO_2$ accumulation rates (averages of $q_{6}$ to $q_{8}$) were on average 15.2 ± 11.7 µg C min$^{-1}$ ($n = 32$) and 11.5 ± 8.0 µg C min$^{-1}$ ($n = 37$) in the Brusquet catchment and in the Moulin catchment, respectively. The associated values of the fitting parameter $\lambda$ (Eq. 2) were on average 0.179 ± 0.076 min$^{-1}$ ($n = 32$) and 0.140 ± 0.061 min$^{-1}$ ($n = 37$) in the Brusquet catchment and in the Moulin catchment, respectively.

At each study site, $CO_2$ was sampled from two chambers. Following recovery from the zeolite sieves in the laboratory, the sampling ratio (SR; Eq. 6) was quantified with an overall average of 1.00 ± 0.15 (Table 4).

3.3 $p_{O_2}$ measurements

The total number of usable $p_{O_2\text{ Chamber}}$ measurements was limited to 15. The difference in $p_{O_2}$ between the chambers and the atmosphere ($p_{O_2\text{ Chamber}} - p_{O_2\text{ Atm}}$) varied over time, ranging from zero within uncertainty (0.18 ± 0.36 %, $v/v$, and ± 95 % confidence interval, CI) to −1.50 ± 0.30 % ($v/v$), with the lowest $p_{O_2\text{ Chamber}}$ values in summer and higher values (lower gradient) in winter at both sites (Fig. 12a).

4 Discussion

Carbon dioxide release during oxidative weathering of sedimentary rocks exposed in steep mountain areas has been shown to vary with changes in temperature, precipitation, and local topography (Soulet et al., 2021; Roylands et al., 2022). Furthermore, previous studies on weathering profiles (Bolton et al., 2006; Petsch, 2014) and on the chemical composition of rivers (Calmels et al., 2007; Bufe et al., 2021; Hilton et al., 2021) indicated that geomorphological and hydrological factors are important controls on the release of $CO_2$ and the consumption of $O_2$ during oxidative weathering. The fluxes should also depend on the pore space characteristics of the weathering zone, such as porosity and tortuosity.
Figure 3. Environmental variables for weathering chambers in 2018 and 2019, with gray shaded areas showing fieldwork visits in the Brusquet catchment (panels a, c, e, g, and i; red) and in the Moulin catchment (panels b, d, f, h, and j; blue). Daily averages are shown by darker colors in all panels. (a, b) Hourly air temperatures from meteorological stations (Draix-Bléone Observatory, 2015). (c, d) Hourly chamber temperatures. Estimated chamber temperatures are indicated by lighter colors (Sect. 2.3) and are shown for gaps in the logger record (denoted). (e, f) Hourly rock surface temperatures. (g, h) Daily rainfall. (i, j) Relative humidity in the chambers, with gaps in the record similar to the chamber temperatures.
Table 2. Overview of the variability in the air temperature, chamber temperature, and rock surface temperature over the study period (27 September 2018 to 2 October 2019) (Fig. 3). A gap in the record of rock surface temperatures at the Moulin site (25 October 2018 to 11 January 2019) is not considered.

<table>
<thead>
<tr>
<th>Variable</th>
<th>Daily averages</th>
<th>Hourly resolution</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Average (± SD)</td>
<td>Min</td>
</tr>
<tr>
<td>Air temperature (°C)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>– Brusquet</td>
<td>10.5 ± 7.4</td>
<td>−4.9</td>
</tr>
<tr>
<td>– Moulin</td>
<td>10.5 ± 7.3</td>
<td>−4.0</td>
</tr>
<tr>
<td>Chamber temperature (°C)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>– Brusquet</td>
<td>15.7 ± 9.4</td>
<td>0.7</td>
</tr>
<tr>
<td>– Moulin</td>
<td>16.6 ± 9.7</td>
<td>0.2</td>
</tr>
<tr>
<td>Rock surface temperature (°C)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>– Brusquet</td>
<td>14.4 ± 8.7</td>
<td>−1.6</td>
</tr>
<tr>
<td>– Moulin</td>
<td>17.9 ± 9.0</td>
<td>−1.4</td>
</tr>
</tbody>
</table>

Table 3. Chamber-specific overview of $p\text{CO}_2$ Rock, with variations over time reported as 1 SD.

<table>
<thead>
<tr>
<th>Chamber identifiers</th>
<th>Site</th>
<th>$p\text{CO}_2$ Rock (ppmv)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Short</td>
<td>Long</td>
</tr>
<tr>
<td>5 B-F-5</td>
<td>Brusquet</td>
<td>861 ± 254</td>
</tr>
<tr>
<td>6 B-G-6</td>
<td>Brusquet</td>
<td>1985 ± 771</td>
</tr>
<tr>
<td>7 B-H-7</td>
<td>Brusquet</td>
<td>936 ± 340</td>
</tr>
<tr>
<td>8 B-I-8</td>
<td>Brusquet</td>
<td>1405 ± 438</td>
</tr>
<tr>
<td>Brusquet total</td>
<td></td>
<td></td>
</tr>
<tr>
<td>1 M-C-1</td>
<td>Moulin</td>
<td>1740 ± 654</td>
</tr>
<tr>
<td>2 M-A-2</td>
<td>Moulin</td>
<td>720 ± 110</td>
</tr>
<tr>
<td>3 M-D-3</td>
<td>Moulin</td>
<td>1881 ± 127</td>
</tr>
<tr>
<td>4 M-B-4</td>
<td>Moulin</td>
<td>1456 ± 576</td>
</tr>
<tr>
<td>Moulin total</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

(Boyle et al., 2006; Brantley et al., 2013; Gu et al., 2020a, b; Soulet et al., 2021). However, we lack direct observations of how the chemical and physical properties of the weathering zone affect the in situ fluxes of CO$_2$ and O$_2$. In addition, only with information on the contributing rock volume to a measured rock-derived flux can we upscale and quantify CO$_2$ and O$_2$ fluxes from the measurement site to the landscape scale.

In the following discussion, we first propose a new approach of interpreting in situ CO$_2$ flux measurements (Sect. 4.1.1) that allows us to assess the diffusion of CO$_2$ and O$_2$ in the shallow critical zone (Sect. 4.1.2). This can be used to quantify the rock volume contributing to the measured fluxes (data flow diagram in Appendix B) (Sect. 4.2). We then examine the implications of these new insights for quantifying the rock-derived CO$_2$ release (Sect. 4.3), and then determine the coinciding O$_2$ consumption (Appendix B) to investigate an overall redox budget of oxidative weathering in an erosive environment (Sect. 4.4).

4.1 Probing the gas exchange of the shallow critical zone

4.1.1 Explaining the patterns of CO$_2$ accumulation during a single flux measurement

To explain the initial decline in the CO$_2$ accumulation rates during a flux measurement that stabilizes over time (Fig. 4), we consider the known volume of a drilled chamber and distinguish it from the unknown volume of rock pore space around it. After arriving at a chamber to start a CO$_2$ flux measurement, the initial scrubbing (before the first repeat $q_1$) removes CO$_2$ from the chamber (i.e., $p\text{CO}_2$ Chamber) to a near-atmospheric level, but we assume that it does not remove the CO$_2$ from the connected pore space to a similar $p\text{CO}_2$ value. Repeated scrubbing and removal of CO$_2$ after CO$_2$ accumulations (Fig. 4) then acts to lower the CO$_2$ concentration in the rock pore space connected to the chamber ($p\text{CO}_2$ Rock). During this process, rock pores that are better connected to the chamber (i.e., having a shorter effective diffusion pathway) provide the initially stored CO$_2$ earlier than rock pores...
that are less connected. In contrast, rock pores that are connected better to the atmosphere than to the chamber are assumed to contribute only to the atmosphere. Once the pool of “excess” CO$_2$ in the pore space that is effectively connected to the chamber has been exhausted, subsequent CO$_2$ accumulation rates reach a plateau and are assumed to represent the real-time production and diffusion of CO$_2$ in the rock surrounding the chamber. This explanation requires that the air volume processed by the sampling system equals the chamber volume, whereas the gas exchange between chamber and connected rock pores happens solely via diffusion. This assumption is supported by the measured CO$_2$ sampling ratio, SR (Eq. 6), with an average SR = 1.00 ± 0.15 (Table 4), showing that we effectively trap the chamber contents. This is similar to the recovery efficiency of ∼95 % of CO$_2$ standards in the laboratory (Garnett et al., 2019).

To interpret the period in which we consider an excess of CO$_2$ (i.e. an excess over the real-time production of CO$_2$) to be diffusing into the chamber from a connected pore space, we introduce an exponential fitting model that describes the decrease in the CO$_2$ accumulation rates ($q$) over time (Fig. 5):

$$q(t) = \alpha \times \exp(-\beta \times t) + q_{\text{Plateau}},$$  

(9)

where $q_{\text{Plateau}}$ is a constant value of the plateaued CO$_2$ accumulation rate, the sum of $\alpha$ and $q_{\text{Plateau}}$ is the initial rate of accumulation (∼$q_1$; mg C min$^{-1}$) at the start of the flux measurement ($t = 0$), $\beta$ is the measurement-specific removal constant (min$^{-1}$), and the term $\alpha \times \exp(-\beta \times t)$ describes the
purging of the initially stored $p_{CO_2}$ from the rock pore space connected to the chamber over time.

To ensure reliable results from fitting the exponential model, measurements are only considered if the last three of at least eight repeats ($q_{n-2}$ to $q_n$ for $n \geq 8$) are “stabilized”, which we define as having a relative standard deviation of less than 5%. To interpret the remaining $CO_2$ flux measurements, chamber-specific averages of the removal constant $\beta$ from stabilized measurements are used to extrapolate $q_{Plateau}$ for measurements that did not stabilize.

The outputs of this analysis provide a $CO_2$ flux that represents the real-time production of $CO_2$ ($q_{Plateau}$), while quantifying the scrubbing of $CO_2$ stored initially in the connected volume of pore space around each chamber. This allows us to assess the diffusive movement of $CO_2$ in the shallow weathering zone (Sect. 4.1.2) and to estimate the contributing rock volume (Sect. 4.2) (Appendix B).

4.1.2 Assessing the diffusivity of the shallow weathering zone

After a $CO_2$ flux measurement, the chamber is resealed. The chamber interior and surrounding pore space will evolve to a steady state of the diffusive movement of $CO_2$ along a concentration gradient between the surface of the rock outcrop and the atmosphere so that $p_{CO_2}^{Chamber} = p_{CO_2}^{Rock}$. 

Figure 5. Examples of $CO_2$ flux measurements consisting of several repeated accumulation rate measurements. Measured accumulation rates ($q$; $\mu g \text{C min}^{-1}$ per chamber; $y$ axis) are shown alongside exponential fits (Eq. 9) describing their evolution over time ($x$ axis) and alongside the 95 % confidence intervals of the modeled level at which the rates plateau ($q_{Plateau}$) for chambers in the Brusquet catchment (a) and in the Moulin catchment (b). Dates and fitting parameters $\alpha$ ($\mu g \text{C min}^{-1}$) and $\beta$ (min$^{-1}$) of the single flux measurements are denoted. For a comparison of the stabilizing evolution of different $CO_2$ flux measurements, indicated by varying colors, $q(t)$ is normalized to $q_{Plateau}$ for both study sites in separate plots (c).
Thus, this steady state of a closed chamber differs from the manipulated environment of a CO$_2$ flux measurement (Fig. 4). The comparison of the two states can shed light on the gas movement and the physical properties of the rocks undergoing weathering. Here, we explore how the observed changes in $p$CO$_2$ _{Rock}$ and CO$_2$ fluxes can be explained by a framework of diffusive processes in the shallow critical zone and assess the degree to which these are modulated by weather conditions.

According to Fick’s law (Eq. 1), diffusion of gases in a porous medium is controlled by (i) the production and accumulation of CO$_2$; (ii) the volume of space (rock pores and/or chamber) and length scale over which molecules travel towards the low-$p$CO$_2$ reservoir; and (iii) the diffusivity of CO$_2$ along their path. We find a co-variation in the $p$CO$_2$ _{Rock}$ and the CO$_2$ fluxes that is similar for both sites, a relationship that can be explained by a linear regression model (Appendix C), with high $p$CO$_2$ _{Rock}$ values coinciding with high CO$_2$ accumulation rates (Fig. 6a). This indicates that the contributing volume of rock pores and the diffusivity (the remaining variables from Fick’s law) may be stable at both sites over the study period.

However, the ambient hydroclimate appears to modify the response of these variables. We consider measurements as being made during “wet” or “dry” periods, whereby wet measurements are those where the cumulative precipitation over the last 3 d was ≥ 5 mm. At a given $p$CO$_2$ _{Rock}$ value (describing the storage of CO$_2$ in the shallow critical zone), dry conditions are associated with lower CO$_2$ production when compared to wet conditions (Fig. 6a). With respect to the CO$_2$ production, previous research has shown that rock-derived CO$_2$ fluxes from drilled chambers are lower following rain events but recover subsequently over a few dry days (Soulet et al., 2021; Roylands et al., 2022). This latter observation has been linked to the degree of water saturation controlling the gas motion in the pore space and thus the supply of O$_2$ for the oxidative weathering reactions, as well as to dissolution of weathering-derived carbon and the subsequent export of dissolved inorganic carbon (DIC) (Soulet et al., 2021; Roylands et al., 2022). It is important to note that a decrease in the production of CO$_2$, associated with a lower O$_2$ supply and/or with a greater uptake of carbon into the DIC, leads also to a proportional decrease in $p$CO$_2$ _{Rock}$ values (Roylands et al., 2022; Soulet et al., 2021). To understand the changes in the relation of $p$CO$_2$ _{Rock}$ and CO$_2$ flux during wet and dry conditions, a change in the diffusivity of the contributing rock volume needs to be considered.

According to Fick’s law, a lower diffusivity at a constant contributing volume of rock results in higher $p$CO$_2$ _{Rock}$ values. Thus, wet conditions may be associated with a decrease in the diffusivity of gases in the weathering rocks. This fits a simple model describing the effective diffusivity $D_{Rock}$ (m$^2$ s$^{-1}$) of a given gas in porous media (such as rocks and soils) at a given temperature by

$$D_{Rock} = D_{Air} \times \phi_{Air-filled} \times \tau,$$

where $D_{Air}$ is the diffusion coefficient (m$^2$ s$^{-1}$) of the particular gas in air, $\tau$ is a dimensionless tortuosity factor, and $\phi_{Air-filled}$ is the air-filled porosity (v/v %) (Penman, 1940; Davidson and Trumbore, 1995). If $\phi_{Air-filled}$ decreases due to meteoric water filling the pore space, then precipitation events lower the effective diffusivity of CO$_2$ within the critical zone (Sánchez-Cañete et al., 2018). An increase in the moisture in porous media also leads to more tortuous pathways (Millington, 1959; Davidson and Trumbore, 1995), which could further lower $D_{Rock}$ under wet conditions. Analogously, rock moisture would also affect the diffusion of atmospheric O$_2$ into the rock pore space, so that this framework can explain the observed decrease in the CO$_2$ production following rain events (Roylands et al., 2022; Soulet et al., 2021).

To describe the diffusion of CO$_2$ during the steady state of a closed chamber, we use Fick’s law (Eq. 1) and the measured CO$_2$ flux ($q_{Plateau}$) and the concentration gradient of CO$_2$ ($p$CO$_2$ _{Rock} - p$CO$_2$ _{Atm.}) to define a measure (D$_{CO2}$, cm$^3$ min$^{-1}$) that describes the effective diffusivity of CO$_2$ towards the atmosphere over the unknown effective depth and area $\omega$ (cm$^2$):

$$D_{CO2}(\omega) = \frac{q_{Plateau}}{pCO2_{Rock} - pCO2_{Atm.}} \times \frac{R \times T}{P} \times 10^{-9} \times MC.$$

The calculated values (based on repeats 6–8) are on average 27.5 ± 12.4 cm$^3$ min$^{-1}$ ($n = 25$) and 21.8 ± 13.2 cm$^3$ min$^{-1}$ ($n = 30$) for the Brusquet catchment and the Moulin catchment, respectively.

An alternative way to assess diffusivity is to use the constant $\lambda$ (Eq. 2) describing the curvature of the repeated accumulations during a CO$_2$ flux measurement (Fig. 4) (Pirk et al., 2016). Differences between $\lambda$ and $\frac{D_{CO2}}{\omega}$ may be expected because $\lambda$ is representative of short intervals (~6 min observations), while $\frac{D_{CO2}}{\omega}$ represents a period of a few hours. We find a significant linear correlation of $\lambda$ and $\frac{D_{CO2}}{\omega}$ for all samples irrespective of the study site (Fig. 6b) (Appendix D). The similarities for both metrics affirm that the accumulation rates determined during flux measurements are representative of the longer-term CO$_2$ release towards the atmosphere.

The concordance of changes in $\lambda$ and in $\frac{D_{CO2}}{\omega}$ suggests that the rock pore space is relatively homogenous in porosity and tortuosity, since the diffusive pathways of the steady state during a stabilized flux measurement differ from that of a closed chamber (Appendix E). Minor heterogeneities may explain some scatter in the correlation of $\lambda$ and $\frac{D_{CO2}}{\omega}$, as well as short-term changes in the effective rock space contributing CO$_2$ to a chamber induced by the percolation of meteoric waters.

In more detail, we find some variability in the measures of diffusivity linked to the hydroclimatic conditions. Wet con-
Figure 6. Comparisons of the diffusivity measures $\lambda$ and $\frac{D(CO_2)}{\partial t}$, CO$_2$ accumulation rates, $pCO_2$ Rock values, and chamber temperatures. Color coding differentiates the “dry” (orange) from “wet” samples (violet), using an approximate threshold of a cumulative precipitation of 5 mm over the last 3 d. The origin of samples is indicated with open circles for the Brusquet catchment and filled triangles for the Moulin catchment. For consistency, all parameters determined during CO$_2$ flux measurements are calculated on the basis of the repeats 6–8 (error bars have 1 SD). Estimated temperatures are indicated by the accompanying error bars (RMSE).

Conditions coincide generally with somewhat lower $\frac{D(CO_2)}{\partial t}$ values for a given $\lambda$ (Fig. 6b) (Appendix D). Because the atmosphere is acting as the low-$pCO_2$ reservoir during the steady state of a closed chamber, $\frac{D(CO_2)}{\partial t}$ is likely to be more influenced by surficial processes than $\lambda$, which is affected by CO$_2$ migration pathways towards the chamber (Appendix E). For example, lower $\frac{D(CO_2)}{\partial t}$ values for a given $\lambda$ may be the result of filling of surficial cracks with water, or micro-landslides, swelling of the surface rock material, and lateral expansion following rainfall events (Bechet et al., 2015), which may hinder the migration of gas. During drier conditions, cracks may significantly increase gas exchange between the rocks and the atmosphere (Weisbrod et al., 2009; Maier and Schack-Kirchner, 2014). In the study area, desiccation cracks typically appear at steep slopes during summer, when erosion by runoff is less prevailing than in spring and

autumn, whereas a thick layer of loose detrital cover can be accumulated during winter due to frost weathering (Ariagno et al., 2022, 2023), when movement of surface materials is limited to solifluction (Bechet et al., 2016). Thus, the diffusivity of the rock surface presumably changes over time, with greater values during dry summer conditions (Fig. 6f and i).

The $\lambda$ and $\frac{D_{CO_2}}{\omega}$ values can also be explored as a function of temperature inside the chambers (Fig. 6f and i). In air, the diffusion coefficient of CO$_2$ is strongly controlled by temperature, with an increase by a factor of $\sim$1.25 at 35$^\circ$C compared to 0$^\circ$C (Massman, 1998). However, we find a much larger change in $\lambda$ and $\frac{D_{CO_2}}{\omega}$, with an average increase by a factor of $\sim$3.5 between 0$^\circ$C and 35$^\circ$C (Fig. 6f and i). This relation between temperature and diffusivity could be explained by a coinciding decrease in rock moisture. In the marls of the Laval catchment, neighboring the studied Moulin catchment, lower near-surface water contents were observed during dry summer periods, with values as low as $\sim$10% contrasting to values of up to $\sim$25% in winter (Mallet et al., 2020). However, the relation of rock moisture and temperature is not straightforward (Soulet et al., 2021), with precipitation being an important control on surface rock moisture. In addition, we observe high and constant relative air humidity in the chambers over the year (Fig. 3).

Overall, the calculated values of $V_{Rock \ pores}$ are similar for both study sites, with $365 \pm 208$ cm$^3$ (± 1 SD of the average of measurement-specific values) for the chambers in the Brusquet catchment and $322 \pm 174$ cm$^3$ for the chambers in the Moulin catchment (Fig. 7) (Table 5). However, significant variation is observed over time for each chamber (Fig. 8), while each measurement-specific value of $V_{Rock \ pores}$ is associated with a high uncertainty (Table 5). These uncertainties are not normally distributed, with an average upper relative uncertainty of 125.8 ± 140.1% (average of 95% CI ± 1 SD) and an average lower relative uncertainty of 47.8 ± 42.3% for all samples.
Table 5. Chamber-specific overview of CO$_2$ excess, $V_{\text{Rock pores}}$, and CO$_2$ accumulation rate ($q_{\text{flux measured}}$), including catchment-specific summaries. Uncertainties in the minima and maxima are representing the 95 % CI, whereas averages are reported with 1 SD.

<table>
<thead>
<tr>
<th>Chamber identifiers</th>
<th>Site</th>
<th>CO$_2$ Excess ($\mu$g C)</th>
<th>$V_{\text{Rock pores}}$ (cm$^3$)</th>
<th>CO$_2$ rate ($\mu$g C min$^{-1}$)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Average</td>
<td>n</td>
<td>Min</td>
<td>Max</td>
</tr>
<tr>
<td>5</td>
<td>B-F-5</td>
<td>Brusquet</td>
<td>222 ± 166</td>
<td>5</td>
</tr>
<tr>
<td>6</td>
<td>B-G-6</td>
<td>Brusquet</td>
<td>353 ± 267</td>
<td>12</td>
</tr>
<tr>
<td>7</td>
<td>B-H-7</td>
<td>Brusquet</td>
<td>142 ± 126</td>
<td>4</td>
</tr>
<tr>
<td>8</td>
<td>B-I-8</td>
<td>Brusquet</td>
<td>261 ± 147</td>
<td>7</td>
</tr>
<tr>
<td>Brusquet total</td>
<td></td>
<td></td>
<td>277 ± 221</td>
<td>28</td>
</tr>
<tr>
<td>1</td>
<td>M-C-1</td>
<td>Moulin</td>
<td>252 ± 150</td>
<td>10</td>
</tr>
<tr>
<td>2</td>
<td>M-A-2</td>
<td>Moulin</td>
<td>55 ± 26</td>
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<tr>
<td>3</td>
<td>M-D-3</td>
<td>Moulin</td>
<td>217 ± 50</td>
<td>3</td>
</tr>
<tr>
<td>4</td>
<td>M-B-4</td>
<td>Moulin</td>
<td>296 ± 192</td>
<td>15</td>
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<tr>
<td>Moulin total</td>
<td></td>
<td></td>
<td>245 ± 174</td>
<td>32</td>
</tr>
</tbody>
</table>

Figure 7. Catchment-specific box plots and distribution curves summarizing the volume of rock pores ($V_{\text{Rock pores}}$) connected to each chamber determined during CO$_2$ flux measurements, including four chambers at each site. Boxes indicate the 25 %–75 % range alongside the 1.5 interquartile ranges with the mean (open square) and median (line). Colors indicate the origin (Brusquet is in red; Moulin is in blue).

4.2.2 Environmental controls on the contributing rock pore volume

The variation in $V_{\text{Rock pores}}$ may be linked to changes in the diffusive processes and weather conditions (Fig. 9). Overall, higher values of $\lambda$ coincide with greater values of $V_{\text{Rock pores}}$ (all samples, with $R^2$ of a linear regression = 0.52, $p \leq 0.001$, and $n = 60$) (Fig. 9a). This is also true for the relation between $V_{\text{Rock pores}}$ and $\Delta$(CO$_2$) (Fig. 9b), which itself is positively correlated to $\lambda$ (Sect. 4.1). These relationships are similar for both sites and for wet and dry conditions. The latter observation indicates that rock moisture impacts the diffusivity of CO$_2$ and $V_{\text{Rock pores}}$ in equal measure. This, in turn, is in line with Fick’s law, with the extent of the rock pore space that contributes CO$_2$ to a chamber depending on the potential of gas to move within the rocks undergoing weathering. In this process, the degree to which changes in the diffusivity impact the contributing rock volume is driven by the effective change in length of the diffusion paths. Here, a change in $\lambda$ from 0.1 to 0.2 min$^{-1}$ is associated with a change in $V_{\text{Rock pores}}$ by a factor of $\sim 3.5$, and this roughly fits the modification of the geometry of a cylinder-shaped rock pore space around a drilled chamber when doubling its effective radius.

Furthermore, differences in $V_{\text{Rock pores}}$ coincide with changes in temperature (all samples, with $R^2$ of a linear regression = 0.47, $p \leq 0.001$, and $n = 60$) (Fig. 9c). This co-occurrence is important to consider when assessing the control of temperature on the CO$_2$ production from chemical weathering (i.e., weathering kinetics) and is discussed later (Sect. 4.3). The coincidence of positive correlations of temperature and CO$_2$ production and of temperature and the extent of $V_{\text{Rock pores}}$ (indirectly connected by $D_{\text{CO}_2}$) also suggests that changes in CO$_2$ flux are associated with changes in the contributing rock pore space (all samples, with $R^2$ of a linear regression = 0.42, $p \leq 0.001$, and $n = 60$) (Fig. 9d).

4.2.3 Upscaling chamber-based CO$_2$ fluxes

The determined $V_{\text{Rock pores}}$ can be combined with the porosity of the rocks undergoing weathering to quantify the volume of rock contributing CO$_2$ to flux measurement ($V_{\text{Rock}}$; cm$^3$):

$$V_{\text{Rock}} = \frac{V_{\text{Rock pores}}}{\phi_{\text{Air-filled}}}.$$  \hspace{1cm} (14)

This assumes that the majority of rock pores are well connected (total porosity $\approx$ connected porosity), with no significant water filling of the pore space. An effective air-filled porosity of 25 %, within a 95 % confidence interval of 15 %–35 %, is assumed, based on porosity measurements (Garel...
et al., 2012; Mallet et al., 2020) and water saturation measurements (Mallet et al., 2020) from the Draix-Bléone Observatory.

On average, CO$_2$ fluxes from chambers in the Brusquet catchment derive from a rock volume of $1459^{+4270}_{-2658}$ cm$^3$ (within a 95% confidence interval, based on propagating the uncertainties in the fitting procedure and of the assigned porosity), which is similar to the release of CO$_2$ in the Moulin catchment from rock volumes of $1286^{+2284}_{-696}$ cm$^3$. If we visualize these volumes as a cylindrical rock layer around the drilled chambers (Fig. 2e shows the sampling distance indicated by arrows pointing towards the chamber), then its thickness would be $\sim 2.2^{+2.6}_{-1.6}$ cm. However, the geometry of this space is unknown. Instead, considering that porosities are highest at the surface of rock outcrops in the study area (Mathys and Klotz, 2008; Loﬁ et al., 2012; Maquaire et al., 2002; Travelletti et al., 2012; Mallet et al., 2020), where unloading and climatic controls on physical weathering act most efficiently (Bechet et al., 2016; Mathys and Klotz, 2008; Bechet et al., 2015; Cras et al., 2007; Ariagno et al., 2022), the shape of the porous and permeable rock that contributes to gas exchange is likely to be more like a cone around a chamber with a radius that declines over depth.

The knowledge of the probed layer thickness can be combined with the inner surface area of the chambers to give the spatial parameter $\omega$ (Eq. 11) and to calculate the effective diffusivity of CO$_2$ in the air-filled rock pores zone. Overall, we find values of $D_{CO_2}$ ranging between $\sim 0.02$ and $\sim 0.34$ cm$^2$ min$^{-1}$ (considering the range of $\frac{D_{CO_2}}{\omega}$ at both study sites; Sect. 4.1). Interestingly, these values are similar to diffusion coefficients that were determined by laboratory experiments (despite potential differences in rock texture and pore space geometry), with O$_2$ as the tracer gas at 22.5 °C, which correspond to $D_{CO_2}$ values of 0.34 and 0.43 cm$^2$ min$^{-1}$ for limestones with porosities of 40% and 46% and of 0.04 and 0.17 cm$^2$ min$^{-1}$ for mudstones with porosities of 33% and 38%, respectively (Peng et al., 2012).

The rock volume around a chamber can be “unwrapped” to assess a surface area on an outcrop that would have the same contributing rock pore volume ($S_{Rock}$; cm$^2$). This can be done if the weathering depth $z_{Rock}$ (cm) over that CO$_2$ is thought to be produced by oxidative weathering is considered, as follows:

$$S_{Rock} = \frac{V_{Rock}}{z_{Rock}}. \quad (15)$$

The $z_{Rock}$ value can be estimated from the morphology of bare surfaces in the study area, based on the assumption that oxidative weathering of sedimentary rocks occurs roughly at the same depths where physical properties are altered (Gu et al., 2020a, b; Lebedeva and Brantley, 2020; Brantley et al., 2013). Based on previous research in a neighboring catchment (Maquaire et al., 2002; Mathys and Klotz, 2008; Oostwoud Wijdenes and Ergenzinger, 1998; Rovéra and Robert, 2006), physical alteration mostly occurs to depths of $\sim 10.0^{+20.0}_{-5.0}$ cm at slopes similar to our study sites (corresponding to the upper two layers comprised of fine regolith overlain by loose detrital cover; Sect. 2.1). Accordingly, we estimate oxidative weathering (i.e., $z_{Rock}$) to extend to similar depths at both study sites.

The corresponding values calculated for $S_{Rock}$ can be compared to the inner surface area of the chambers (Table 1). On
average, $S_{\text{Rock}}$ is smaller than the inner surface area of the drilled chambers by a factor of $2.7^{+0.02}_{-0.05} \ t \text{m}^{-3}$ (confidence interval includes the considered range of the weathering depth and the uncertainty in the contributing rock volume). This means that CO$_2$ fluxes from chambers drilled into rocks and normalized to the chamber inner surface area (Eq. 4) cannot be compared directly with topographic surface fluxes (e.g., from surface chambers), which are typically reported for soils (Oertel et al., 2016; Bond-Lamberty and Thomson, 2010). Instead, CO$_2$ fluxes from a drilled chamber need to be corrected by considering the chamber-specific $V_{\text{Rock}}$ and the weathering depth. The measured CO$_2$ fluxes (Table 5) equate to a topographic surface efflux of $541 \ t \text{C km}^{-2} \text{yr}^{-1}$ (ranging on average between 73 and 1108 t C km$^{-2}$ yr$^{-1}$) in the Brusquet catchment and of $425 \ t \text{C km}^{-2} \text{yr}^{-1}$ (between 43 and 873 t C km$^{-2}$ yr$^{-1}$) in the Moulin catchment, which is roughly similar to the global mean emissions of CO$_2$ from soils with different land cover (Oertel et al., 2016).

Carbon fluxes from oxidative weathering can be linked to a rock mass, allowing fluxes to be interpreted in terms of the overall kinetics of the oxidative weathering reactions. This is essential for theoretical carbon cycle modeling (Bolton et al., 2006; Bao et al., 2017). Following the calculation of $V_{\text{Rock}}$, the rock mass emitting CO$_2$ ($m_{\text{Rock}}; \ g$) can be estimated by using an average estimate of the density of the rock grains surrounding the chambers ($\rho_{\text{Rock}}; \ g \text{cm}^{-3} = \text{t m}^{-3}$):

$$m_{\text{Rock}} = (V_{\text{Rock}} - V_{\text{Rock pores}}) \times \rho_{\text{Rock}}.$$  \hspace{1cm} (16)

Considering a grain density of $2.7^{+0.02}_{-0.05} \ g \text{cm}^{-3}$ (Lofi et al., 2012), we find that an average rock mass of $2955^{+1029}_{-1908}$ and $2605^{+5651}_{-1577} \ g$ produces the CO$_2$ fluxes derived from chambers in the Brusquet catchment and in the Moulin catchment, respectively. To our knowledge, this allows, for the first time, an absolute report of weathering-derived CO$_2$ fluxes that are measured in real time and in situ.

The combined quantification of CO$_2$ fluxes and of the corresponding rock mass undergoing oxidative weathering...
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Figure 10. Comparison of CO₂ fluxes determined by using the average of repeated accumulation rates and by an exponential fitting model. Results from the proposed model are given on the x axis ($q_{\text{Plateau}}$; Eq. 9; error bars show 95 % CI) and are compared with the averages of the repeats 4–6 (a) and of the repeats 6–8 (b) on the y axis (error bars show 2 SD), alongside a linear regression and a 1 : 1 line for reference. Red colors indicate samples from the Brusquet catchment (stabilized in dark colors; extrapolated fitting in light colors; Sect. 4.1.1), and blue colors indicate samples from the Moulin catchment (stabilized in dark colors; extrapolated fitting in light colors).

means that there is an opportunity for future research to include investigations of the internal surface area of the studied rocks, which would allow reporting field-based CO₂ fluxes normalized to the reacting surface areas. Such normalizations are typically considered during modeling (Bao et al., 2017; Bolton et al., 2006) to acknowledge that the internal surface area can change significantly during sedimentary rock weathering (Fischer and Gaupp, 2005). Analogously to silicate weathering rates, variations in OCₚetro and carbonate weathering rates obtained from different field and laboratory conditions may be related to the conceptualization and parameterization of the reactive surface area, which needs to be considered when comparing them (Brantley et al., 2007; White and Brantley, 2003).

4.3 Implications for CO₂ flux measurements

4.3.1 Accuracy of CO₂ flux measurements

The time-dependency of carbon accumulations during the CO₂ flux measurements (Figs. 4 and 5) has provided new insights into the nature of the shallow weathering zone (Sects. 4.1 and 4.2) and has important implications for how CO₂ fluxes are quantified. In short, changes in $pC_{\text{O}_2}$ during the field operations are interpreted as the combination of (i) purging of CO₂ stored initially in the chamber and surrounding rock pores ($pC_{\text{O}_2}$ Rock) and (ii) the real-time production of CO₂ from oxidative weathering. At the start of a CO₂ flux measurement, CO₂ accumulations have important contributions from point (i), which led previous studies to use the later repeats ($q_4$ to $q_{n≥6}$) to quantify point (ii) (Soulet et al., 2018, 2021; Roylands et al., 2022). Here, longer measurements at each chamber allow us to explore this in more detail.

The time it takes for the CO₂ accumulation rates to decline and stabilize within the 95 % confidence interval of the $q_{\text{Plateau}}$ value derived from fitting CO₂ accumulation rates over time (Eq. 9) is ~90 min, corresponding to ~10 repeats (Fig. 5). This implies that treating only the first three repeats of a measurement as combined signals of purging and production of CO₂, as done previously in similar studies (Roylands et al., 2022; Soulet et al., 2021, 2018), returns a greater flux than $q_{\text{Plateau}}$. However, the overestimate is modest; the average accumulation rate of the measured repeats 4–6 is ~15 % higher than the $q_{\text{Plateau}}$ value (Fig. 10a). This is true for the entire data set, including stabilized and extrapolated measurements, and for the site-specific samples. The relative offset is constant, irrespective of the overall size of the CO₂ accumulation (Fig. 10a), which means that flux data including non-plateaued accumulations can be corrected. It also means that any link between CO₂ flux and measured environmental variables is robust (Roylands et al., 2022; Soulet et al., 2021). When taking the average of the measured repeats 6–8 (Fig. 10b), the value is ~7 % higher than $q_{\text{Plateau}}$.

4.3.2 Reporting CO₂ flux as a function of temperature

Recent research has highlighted that temperature controls the release of CO₂ from chambers drilled into the shallow weathering zone of sedimentary rocks (Soulet et al., 2021), with an exponential response:
Changes in temperature also coincide with changes in the diffusive processes in the rocks surrounding a chamber (Sect. 4.1). To differentiate changes in the diffusive framework from the CO₂ production at a given temperature (i.e., weathering kinetics), the CO₂ fluxes can be normalized to $V_{\text{Rock pores}}$, which is representative of the contributing number of rock grains undergoing oxidation (Sect. 4.2). Since the observed CO₂ fluxes range by a factor of $\sim 18.2$, while values of $V_{\text{Rock pores}}$ exhibit a lower range of a factor of $\sim 5.9$ (Table 5), this normalization does not diminish the importance of the temperature control on the CO₂ release (Fig. 11c). However, due to large measurement-specific uncertainties that are associated with the calculation of $V_{\text{Rock pores}}$, a full assessment of whether higher chamber-derived CO₂ fluxes at higher temperatures are partly a result of greater contributing rock volumes is hindered.

Despite the similarities of the topography, hydrology, erosion rates (Figs. 1 and 2), and $V_{\text{Rock pores}}$ (Figs. 7 and 8), we find site-specific differences in the bulk CO₂ production at a given temperature, which may be linked to differences in the source of carbon associated with the different rock types outcropping in the Brusquet catchment (black shales; $F_0 = 122.2 \pm 41.3 \text{ mg C m}^{-2} \text{ d}^{-1}$) and in the Moulin catchment (marls; $F_0 = 45.6 \pm 20.3 \text{ mg C m}^{-2} \text{ d}^{-1}$). To better understand these different weathering fluxes, future research is needed to assess the carbon source(s) and the response of the weathering reactions to changes in temperature at both study sites in more detail, for instance, by studying the chem-

$$F = F \times \exp(\gamma T),$$  \hspace{1cm} (17)
tical composition of the rocks (i.e., contents of OCpetro, carbonates, and sulfides) alongside the radiocarbon and stable carbon isotope composition of the released CO$_2$, analogously to previous research in the neighboring Laval catchment (Soulet et al., 2018, 2021) and in the Waiapu catchment in Aotearoa / New Zealand (Roylands et al., 2022).

### 4.4 Linking CO$_2$ and O$_2$ fluxes

The O$_2$ consumption in the weathering zone provides a tool for investigating the kinetics of sedimentary rock weathering (Tune et al., 2020, 2023), analogous to previous research on soils (Hicks Pries et al., 2020; Angert et al., 2015; Sánchez-Cañete et al., 2018). Overall, $p$O$_2$ Chamber values in the Brusquet catchment and in the Moulin catchment were similar to or lower than the O$_2$ concentration of the atmosphere, confirming that the weathering zones are sinks of oxygen. The variation in $p$O$_2$ Chamber over time coincides with a variation in temperature within the chambers (Fig. 12a), with roughly similar relationships for both study sites. This observation fits to the recently proposed importance of temperature controlling oxidative weathering kinetics (Soulet et al., 2021). However, due to a limited number of samples and a large measurement uncertainty, neither chamber-specific differences nor the impact of precipitation on $p$O$_2$ Chamber can be evaluated accurately.

The observed $p$O$_2$ Chamber values can be used to calculate a diffusive flux of O$_2$ between the chamber, connected rock pores, and the atmosphere (Eqs. 7 and 8). This is based on the insights into the diffusive processes of the chambers and the connected rock space that come from the CO$_2$ measurements (Sects. 2.7 and 4.1) (Appendix B). Previous work on porous media has established that the effective diffusivities of CO$_2$ and O$_2$ are impacted in a similar way by the structure of the air-filled pore space (Millington, 1959; Penman, 1940; Angert et al., 2015), which allows determination of the effective diffusivity of O$_2$ from that of CO$_2$ in the shallow weathering zone.

The calculated O$_2$ fluxes are representative of the same rock volume that is releasing CO$_2$, which also means that a report of absolute fluxes is possible (Sect. 4.2). Altogether, the chamber-specific O$_2$ exchange rate in the Brusquet catchment and the Moulin catchment range between zero within uncertainty (0.42$^{+0.58}_{-0.88}$ μmol O$_2$ min$^{-1}$) to a maximum consumption of O$_2$ of $-16.33^{+7.72}_{-5.75}$ μmol O$_2$ min$^{-1}$, with increasing O$_2$ consumption with increasing temperature (Fig. 12b). The O$_2$ fluxes have a greater relative uncertainty compared to the $p$O$_2$ gradient because they include the measurement-specific diffusivity.

The O$_2$ flux into the chambers and their connected rock pores can be compared with the CO$_2$ flux from this space (Fig. 12c). At 20°C, an O$_2$ consumption rate of $\sim -8.7$ μmol O$_2$ min$^{-1}$ coincides with an average CO$_2$ accumulation rate of $\sim 1.1$ μmol CO$_2$ min$^{-1}$ in the Brusquet catchment and of $\sim 0.6$ μmol CO$_2$ min$^{-1}$ in the Moulin catchment. This is an average ratio of $\sim 1$ mol O$_2$ : 0.1 mol CO$_2$. This field-based molar ratio of O$_2$ consumption and CO$_2$ release is significantly lower than the theoretical ratio of weathering reactions describing the oxidation of sedimentary rocks. For example, the oxidation of OCpetro is theoretically described by a ratio of 1 mol O$_2$ : 1 mol CO$_2$ (Petsch, 2014). In addition, the oxidation of pyrite minerals coupled to the dissolution of carbonates is theoretically characterized by a ratio of up to 1.875 mol O$_2$ : 1 mol CO$_2$ if the CO$_2$ release occurs in situ (Torres et al., 2014; Soulet et al., 2021). To investigate this discrepancy, here we discuss several mechanisms that could influence the consumption of O$_2$ and the release of CO$_2$ in the shallow weathering zone.

In addition to OCpetro and pyrite minerals, other minerals, such as illite, chlorite, and ankerite, can be a sink of oxygen during the weathering of sedimentary rocks (Brantley et al., 2013; Sullivan et al., 2016). However, in settings where pyrite minerals are present, the chemical weathering of these other ferrous-iron-bearing minerals progresses relatively slowly and advances only more rapidly following the complete oxidation of pyrite minerals (Gu et al., 2020a, b). Accordingly, pyrite minerals should be the dominating inorganic O$_2$ sink at the two study sites.

It has been previously suggested that the oxidation of OCpetro progresses in a stepwise manner, with the formation of oxygenated compounds of organic matter prior to the release of CO$_2$ (Chang and Berner, 1999), typically resulting in an increase in the relative oxygen content of OCpetro during chemical weathering (Petsch, 2014; Tamamura et al., 2015; Longbottom and Hockaday, 2019). If the oxidation of OCpetro progresses more rapidly than the separate process of CO$_2$ release from oxygenated OCpetro in the Draix-Bléone Observatory, then this could partly explain the lower ratio of CO$_2$ released compared to the O$_2$ uptake. However, for rocks exposed in rapidly eroding terrains in the Draix-Bléone Observatory, previous studies did not find a significant effect of weathering on the chemical composition of OCpetro, despite a decrease in the quantities of OCpetro and pyrite minerals (Graz et al., 2011; Copard et al., 2006). Thus, it seems unlikely that the oxidation of OCpetro at both study sites deviates notably from the theoretical stoichiometry mentioned above.

Furthermore, if the sulfuric acid derived from the oxidation of pyrite minerals interacts with silicate minerals (Bufe et al., 2021; Blattmann et al., 2019), then this would lead to O$_2$ consumption but no CO$_2$ release. In addition, sulfuric acid could interact with OCpetro and be neutralized, yet the vast majority of OCpetro is typically made of kerogen, which is resistant to acid hydrolysis, and only minor amounts of more labile organic matter can be prone to this type of degradation (Killops and Killops, 2005; Petsch, 2014; Seifert et al., 2011; Włodarczyk et al., 2018).

Another explanation may involve the lateral transport of CO$_2$ as part of the dissolved load, lowering the gaseous release of carbon. The oxidation of OCpetro and pyrite minerals
coupled to the dissolution of carbonates occur in a humid weathering zone (Fig. 3), where CO$_2$ may be incorporated into the dissolved inorganic carbon pool (Torres et al., 2014; Soulet et al., 2021; Roylands et al., 2022; Bao et al., 2017). A recent study has quantified the export of DIC using the molar ratio of O$_2$ and CO$_2$ fluxes for sedimentary rocks undergoing weathering below a forested hillslope (Tune et al., 2023, 2020). There, carbon is sourced from soils, roots, and OC$_{petro}$, with an absence of an inorganic carbon source and of pyrite minerals. If a part of the CO$_2$ from oxidative weathering is exported as DIC in the Brusquet and the Moulin catchments, then this would raise the observed O$_2$ consumption to CO$_2$ release ratio and would have to do so by a factor in the range of $\sim 4$ and $\sim 15$. This is worthy of future research towards understanding O$_2$ consumption in the weathering zone. Here, a more accurate quantification is hindered because the proportions of inorganic carbon release versus organic carbon release (which derive from weathering reactions with a different stoichiometry as described above) are unknown. This again calls for future work to assess the carbon sources in more detail.

Overall, we have developed the tools needed to quantify the production or consumption, storage, and movement of CO$_2$ and O$_2$ in the near-surface of rocks undergoing weathering (Appendix B). If combined, for example, with surface chambers (for gaseous exchange) and boreholes extending below the oxidation front (profiling gaseous and dissolved processes) (Tune et al., 2020; Tokunaga et al., 2016), alongside radiocarbon and stable carbon isotope analyses (for partitioning the weathering reactions) (Soulet et al., 2021; Roylands et al., 2022; Keller and Bacon, 1998; Tune et al., 2023), the cycling of carbon and oxygen in the total critical zone and its environmental controls could be investigated comprehensively. Using drilled chambers benefits investigations using the gradient method for profiles of gas concentrations because, with the approach suggested here, the diffusivity measures can be determined in situ and in real time, which are otherwise typically estimated (Tune et al., 2020; Maier and Schack-Kirchner, 2014; Keller and Bacon, 1998; Tokunaga et al., 2016). Furthermore, by assessing the release and movement of nitrous oxide from the subsurface (Wan et al., 2021), an overall greenhouse gas budget could be developed for sedimentary rocks undergoing weathering. This would be especially valuable for sites with a thin soil cover, which typically dominate more widespread terrains at lower slopes (Milodowski et al., 2015; Heimsath et al., 2012). There, the additional, modern carbon pool complicates the disentangling of biogeochemical processes and the corresponding source-specific CO$_2$ and O$_2$ fluxes (Tune et al., 2020; Keller and Bacon, 1998; Longbottom and Hockaday, 2019; Hemingway et al., 2018; Copard et al., 2006; Tune et al., 2023).

5 Conclusions

This study has further developed and assessed methods for in situ constraints on the release of CO$_2$ and the consumption of O$_2$ during oxidative weathering of exposed sedimentary rocks. Our new method framework allows for both accurate quantification of weathering fluxes over hourly to daily timescales, while also constraining diffusive processes in the shallow weathering zone. At two sites of the Draix-Bléone Observatory (France), accumulation chambers were installed by drilling holes directly into rocks undergoing weathering in the Brusquet catchment (black shales) and in the Moulin catchment (marls). At each site, using an array of four chambers, measurements of $\rho$CO$_2$$_{Rock}$ and CO$_2$ fluxes were carried out alongside $p$O$_2$$_{Rock}$ measurements during six field trips over 1 year.
We find that during a single visit to a chamber, the accumulation rates decline over a few measurement cycles before reaching a stable CO$_2$ accumulation rate. This pattern is consistent across the field trips and can be described by an exponential model. To explain these observations, we provide a new framework which considers the measured CO$_2$ accumulation as a combination of the real-time CO$_2$ production during weathering plus the release of excess CO$_2$ built up in pore space surrounding the chambers. By doing so, we can assess the rock pore volume and rock mass that produce CO$_2$. For the first time, this allows an absolute report of rock-derived CO$_2$ fluxes measured in situ and in real time, providing input data for future studies modeling the chemical weathering of sedimentary rocks. The assessment of contributing rock pore space allows us also to normalize the fluxes to an outcrop surface area, enabling comparison of the weathering fluxes at the study sites to other rock types and soils across different terrains and climates. Furthermore, by studying the accumulation of CO$_2$ in a chamber and the connected rock pore space over time, the diffusivity of gases in the shallow weathering zone and its environmental controls are investigated, including an absolute, in situ determination of the diffusion coefficients.

In addition to these insights into the CO$_2$ release, $pO_2$ values for the studied rocks are presented and used together with the quantification of the diffusive processes in the weathering zone to calculate O$_2$ fluxes. It is shown that the consumption of O$_2$ co-varies with changes in the emission of CO$_2$ over time, which are driven by changes in temperature. However, the O$_2$ fluxes indicate significantly greater oxidative weathering rates compared to the CO$_2$ fluxes. We suggest that this discrepancy may result from the (i) export of inorganic carbon by the dissolved load of percolating waters lowering the effective release of gaseous CO$_2$ and (ii) silicate weathering by sulfuric acid as a sink of O$_2$.

A site-specific difference in the magnitude of CO$_2$ emissions at the two study sites cannot be explained by differences in the lithological properties influencing the diffusion of gas within the rock space surrounding the chambers, as both study sites have similar characteristics, which is evidenced by diffusivity measures changing similarly alongside temperature and precipitation. This finding suggests that differences in the source of carbon are the main reason for the observed CO$_2$ flux differences, providing an opportunity for future research to investigate the control of the chemical composition of the rocks (i.e., contents of OC$_{petro}$, carbonates and sulfides) on the CO$_2$ flux size.

Together, this study provides a new framework to quantify the release of CO$_2$ and the consumption of O$_2$ during oxidative weathering of sedimentary rocks, aiding our efforts to upscale fluxes and constrain how they respond to changes in climate and the geomorphic and hydrological setting.

Appendix A: Modeling chamber temperatures

To fill the gaps in the direct chamber temperature measurements, we use air temperatures from a local weather station as a proxy by modifying a framework that describes soil temperatures by, amongst other variables, air temperature (Liang et al., 2014). The approach combines measured air temperatures with a Fourier-fitted function that describes the daily average temperature inside the rock chambers by weighting averaged air temperatures by the fractional duration of daylight ($L$) at the latitude of the Draix-Bléone Observatory. In more detail, we estimate the current chamber temperature $T_{Chamber}$ ($^\circ$C) at an hourly resolution as follows:

$$T_{Chamber} = T_{mean} \times \text{coeff}_A + (T_{Air-6 h} - T_{mean}) \times \text{coeff}_B, \quad (A1)$$

where $T_{Air-6 h}$ is the hourly air temperature ($^\circ$C) from nearby meteorological stations (Draix-Bléone Observatory, 2015) delayed by 6 h, $\text{coeff}_A$ and $\text{coeff}_B$ are site-specific fitting coefficients, and $T_{mean}$ ($^\circ$C) is the long-term trend of rock temperature described by

$$T_{mean} = \text{coeff}_{C1} + \text{coeff}_{C2} \times \cos(\text{coeff}_{C3} \times T_{Air, 7 d} \times L) + \text{coeff}_{C4} \times \sin(\text{coeff}_{C3} \times T_{Air, 7 d} \times L), \quad (A2)$$

where $\text{coeff}_{C1}$ to $\text{coeff}_{C4}$ are site-specific fitting coefficients (Table A1) in a first-order Fourier-model, and $T_{Air, 7 d}$ ($^\circ$C) is the 7 d average of the past air temperatures at hourly resolution. Using the site-specific air temperatures, this approach simulates $T_{Chamber}$ well, with a root mean square error (RMSE) of 1.8 $^\circ$C for the Brusquet catchment and 2.2 $^\circ$C for the Moulin catchment (Fig. A1).
Figure A1. (a) Comparison of chamber temperatures measured and estimated by a modeling framework based on air temperature and the fractional duration of daylight at the latitude of the Draix-Bléone Observatory at hourly resolution for the Brusquet catchment (red) and the Moulin catchment (blue), which agree with a 1:1 relation (dashed line). (b) Normally distributed residuals between the measured temperatures and the modeling framework.

Table A1. Overview of site-specific fitting coefficients used for modeling chamber temperatures based on air temperature and fractional duration of daylight (Eqs. A1 and A2) and details of the goodness of the fitting model, based on comparisons to measured chamber temperatures with hourly and daily resolution.

<table>
<thead>
<tr>
<th>Site</th>
<th>$R^2$</th>
<th>$p$ value</th>
<th>$n$</th>
<th>RMSE (hourly) ($°C$)</th>
<th>RMSE (daily) ($°C$)</th>
<th>coeff$_A$</th>
<th>coeff$_B$</th>
<th>coeff$_C_1$</th>
<th>coeff$_C_2$</th>
<th>coeff$_C_3$</th>
<th>coeff$_C_4$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Brusquet</td>
<td>0.96</td>
<td>$&lt; 0.001$</td>
<td>7392</td>
<td>1.80</td>
<td>1.58</td>
<td>1.147</td>
<td>0.361</td>
<td>0.408</td>
<td>2.364</td>
<td>0.080</td>
<td>30.106</td>
</tr>
<tr>
<td>Moulin</td>
<td>0.94</td>
<td>$&lt; 0.001$</td>
<td>5664</td>
<td>2.21</td>
<td>1.99</td>
<td>1.222</td>
<td>0.298</td>
<td>0.362</td>
<td>2.074</td>
<td>0.093</td>
<td>26.407</td>
</tr>
</tbody>
</table>
Appendix B: Data flow diagram for chamber-based CO$_2$ and O$_2$ flux measurements

Figure B1. Central data flow for the new approach developed in this study to quantify the diffusive exchange of CO$_2$ and O$_2$ during shallow rock weathering, based on real-time measurements using drilled headspace chambers.
Appendix C: Linear regression of $p\text{CO}_2\text{Rock}$ and \(\text{CO}_2\) accumulation rates

Table C1. Details of linear regressions comparing measurement-specific values of $p\text{CO}_2\text{Rock}$ and \(\text{CO}_2\) accumulation (based on repeats 6–8) (with $p\text{CO}_2$ (ppmv) = $a_1 \times \text{CO}_2$ rate (µg C min$^{-1}$) + $a_2$), including the standard errors in the fitting parameters. Differentiations into dry and wet samples are based on a threshold of a cumulative precipitation of 5 mm over 3 d prior to the measurement. Both sites have similar linear regressions with overlapping standard errors.

<table>
<thead>
<tr>
<th>Data set</th>
<th>$a_1$ ±</th>
<th>$a_2$</th>
<th>R$^2$</th>
<th>p value</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>All</td>
<td>55.1 ± 5.1</td>
<td>848.7 ± 83.5</td>
<td>0.69</td>
<td>&lt; 0.001</td>
<td>55</td>
</tr>
<tr>
<td>– Brusquet</td>
<td>55.0 ± 6.3</td>
<td>741.6 ± 120.5</td>
<td>0.77</td>
<td>&lt; 0.001</td>
<td>25</td>
</tr>
<tr>
<td>– Moulin</td>
<td>61.1 ± 8.5</td>
<td>872.5 ± 117.0</td>
<td>0.64</td>
<td>&lt; 0.001</td>
<td>30</td>
</tr>
<tr>
<td>All – dry</td>
<td>59.1 ± 6.4</td>
<td>664.3 ± 122.2</td>
<td>0.75</td>
<td>&lt; 0.001</td>
<td>31</td>
</tr>
<tr>
<td>All – wet</td>
<td>61.0 ± 8.3</td>
<td>948.7 ± 101.0</td>
<td>0.71</td>
<td>&lt; 0.001</td>
<td>24</td>
</tr>
</tbody>
</table>

Appendix D: Linear regression of diffusivity measures

Table D1. Details of linear regressions comparing the diffusivity measures $\frac{\Delta\text{CO}_2}{\Delta t}$ and $\lambda$ (with $\frac{\Delta\text{CO}_2}{\Delta t}$ = $b_1 \times \lambda$ (min$^{-1}$) + $b_2$), including the standard errors in the fitting parameters. Differentiations into dry and wet samples are based on a threshold of a cumulative precipitation of 5 mm over 3 d prior to the measurement. The regressions are based on using the average values from the repeats 6–8 of the flux measurements, which are typically close to stabilization of the \(\text{CO}_2\) accumulation rate because a significant number of the measurements was limited to eight repeats. Both sites have comparable linear regressions with overlapping standard errors.

<table>
<thead>
<tr>
<th>Data set</th>
<th>$b_1$ ±</th>
<th>$b_2$</th>
<th>R$^2$</th>
<th>p value</th>
<th>n</th>
</tr>
</thead>
<tbody>
<tr>
<td>All</td>
<td>163.4 ± 13.2</td>
<td>−1.13 ± 2.25</td>
<td>0.74</td>
<td>&lt; 0.001</td>
<td>55</td>
</tr>
<tr>
<td>– Brusquet</td>
<td>162.1 ± 13.9</td>
<td>−1.54 ± 2.67</td>
<td>0.86</td>
<td>&lt; 0.001</td>
<td>25</td>
</tr>
<tr>
<td>– Moulin</td>
<td>170.4 ± 24.0</td>
<td>−1.57 ± 3.61</td>
<td>0.64</td>
<td>&lt; 0.001</td>
<td>30</td>
</tr>
<tr>
<td>All – dry</td>
<td>151.4 ± 18.5</td>
<td>2.25 ± 3.58</td>
<td>0.71</td>
<td>&lt; 0.001</td>
<td>31</td>
</tr>
<tr>
<td>All – wet</td>
<td>157.2 ± 22.9</td>
<td>−1.94 ± 3.13</td>
<td>0.68</td>
<td>&lt; 0.001</td>
<td>24</td>
</tr>
</tbody>
</table>

Appendix E: Different diffusion pathways of a closed versus manipulated chamber

The diffusion pathways of a closed chamber differ from that of a manipulated chamber. During a flux measurement, \(\text{CO}_2\), which is released from a rock grain undergoing oxidative weathering into the pore space, moves via diffusion along a concentration gradient, which is initiated by repeatedly lowering the $p\text{CO}_2\text{Chamber}$ to a near-atmospheric level (Soulet et al., 2018) from the rock pores towards the manipulated chamber if they are connected more effectively to the chamber than to the rock–atmosphere boundary. In contrast, without a sampling system acting as the receiving reservoir, rock-derived \(\text{CO}_2\) travels along the concentration gradient towards the atmosphere. In the latter scenario, rock pores that contribute \(\text{CO}_2\) during a flux measurement to a chamber either contribute the \(\text{CO}_2\) via diffusion directly to the atmosphere (without a pathway through the chamber) or through the chamber towards the atmosphere.

Appendix F: Repeated \(\text{CO}_2\) flux measurements on the same date

On four visits, a chamber was measured twice a day, and the observed \(\text{CO}_2\) release was higher in the afternoon than in the morning, coinciding with an increase in the chamber temperature. For chamber 5 at the Brusquet site, $q_{\text{Plateau}}$ increased from 1.1 to 3.8 µg C min$^{-1}$ and from 2.6 to 4.7 µg C min$^{-1}$, coinciding with temperate increases from $−0.9$ to $1.0 \degree\text{C}$ and from $2.0$ to $5.0 \degree\text{C}$, respectively. For chamber 4 at the Moulin site, $q_{\text{Plateau}}$ increased from 14.6 to 18.0 µg C min$^{-1}$ and from 16.3 to 21.4 µg C min$^{-1}$, coinciding with temperate increases from 22.5 to 26.5 \degree\text{C} and from 22.0 to 27.0 \degree\text{C}, respectively.

Data availability. All data supporting the findings of this study can be found in the tables in the Supplement.

Supplement. The supplement related to this article is available online at: https://doi.org/10.5194/esurf-12-271-2024-supplement.

Author contributions. The research was conceptualized by TR, with help from RGH and ELM. The main methods were designed by TR, with help from RGH, GS, MHG, and ELM. Field-based and laboratory geochemical measurements were led by TR, with contributions from MHG, RGH, SK, MD, and FN, and climate data and aerial imagery were provided by SK and CLB. TR led the formal analysis, investigation, data visualization, and writing of the original draft under the supervision of RGH and ELM. All authors contributed to subsequent review and editing, led by TR, RGH, ELM, and GS. Funding was acquired by RGH, TR, and MHG.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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