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Erosion sensitivity of rock organic carbon oxidation reinforces CO₂ release at high erosion rates

Victoria Alcock^{1*}, Aaron Bufe², Ella V. Walsh^{3,4}, Niels Hovius^{5,6}, Li-Hung Lin^{7,8}, Robert G. Hilton^{9,1}

¹Department of Earth Sciences, University of Oxford, Oxford, UK

²Department of Earth and Environmental Sciences, Ludwig-Maximilians-Universität München, Munich, Germany

³NORCE Norwegian Research Centre; Bjerknes Centre for Climate Research, Bergen 5007, Norway

⁴SFF Centre for Early Sapiens Behaviour (SapienCE), AHKR, University of Bergen, Bergen 5020, Norway

⁵GFZ Helmholtz Centre for Geosciences, Telegrafenberg, 14473 Potsdam, Germany

⁶Institute of Geosciences, University of Potsdam, 14476 Potsdam-Golm, Germany

⁷NTU Research Center for Future Earth, National Taiwan University, Taipei, 106, Taiwan

⁸Department of Geosciences, National Taiwan University, Taipei, 106, Taiwan

⁹School of Geography and the Environment, University of Oxford, Oxford, UK

*Corresponding author. Email: victoria.alcock@earth.ox.ac.uk

Abstract

Oxidation of rock organic carbon (OC_{petro}) can exert a substantial control on atmospheric CO_2 over geological timescales, yet the erosion sensitivity of this reaction has proved challenging to isolate and quantify due to the covariation of erosion with other environmental variables in many existing datasets. Here we apply the rhenium (Re) proxy to quantify OC_{petro} oxidation across a three order-of-magnitude erosion-rate gradient in Taiwan with minimal variation in runoff, temperature, and lithology, and use a weathering model to describe the erosion sensitivity of OC_{petro} oxidation. OC_{petro} oxidation fluxes exhibit a positive non-linear scaling with erosion, similar to sulfide mineral weathering, although with a weaker erosion rate dependence likely reflecting slower oxidation kinetics and the presence of recalcitrant graphitic carbon. Integrating OC_{petro} oxidation with existing silicate, carbonate, and sulfide weathering carbon budgets reveals erosion-dependent transitions in the net weathering carbon balance. Moderate erosion rates around 0.1 mm yr^{-1} maximise CO_2 drawdown, whereas higher erosion rates produce net CO_2 release due to oxidative weathering of OC_{petro} and sulfide minerals. Accounting for OC_{petro} oxidation increases CO_2 emissions at high erosion rates and dampens CO_2 drawdown at moderate erosion rates relative to previous characterisations that considered emissions only from sulfide oxidation. Unlike sulfide oxidation, CO_2 emissions from OC_{petro} oxidation remain a relevant carbon source on timescales of 10 Myr or more, extending the CO_2 drawdown optimum over longer timescales. Erosion and burial of biospheric organic carbon (OC_{bio}) may provide an erosion-sensitive CO_2 sink that could partially offset oxidative CO_2 release at high erosion rates.

41 Keywords

- 42 • Organic carbon
- 43 • Erosion
- 44 • Rhenium
- 45 • Oxidative weathering
- 46 • Carbon cycle

47 Highlights

- 48 • Rhenium proxy used to assess erosion sensitivity of rock organic carbon oxidation.
- 49 • Rock organic carbon oxidation scales with erosion with a square-root relationship.
- 50 • OC_{petro} oxidation is not wholly supply-limited, kinetic factors are also important.
- 51 • Rock organic carbon oxidation diminishes CO₂ drawdown potential of mountain ranges.
- 52 • Terrestrial organic carbon burial partially offsets rock organic carbon CO₂ release.

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1. Introduction

The chemical weathering of rocks at Earth's surface moderates atmospheric CO₂, and therefore global climate on million-year timescales (Urey, 1952). Since erosion increases the supply of unweathered mineral surfaces, there exists an intrinsic link between erosion, weathering, and long-term climate (e.g. Gaillardet *et al.*, 1999; West *et al.*, 2005). This link is typically viewed in terms of silicate weathering, where increased erosion could enhance sequestration of atmospheric CO₂ (e.g. Raymo and Ruddiman, 1992) provided climate conditions are favourable for weathering (West *et al.*, 2005).

However, recent studies emphasise the importance of coupled sulfide and carbonate weathering reactions in active orogens with high erosion rates (Kemeny *et al.*, 2021; Torres *et al.*, 2016). Meanwhile, some studies report a plateau or decline in silicate weathering at high erosion rates, while sulfide and carbonate weathering may remain sensitive to erosion (e.g. Bufe *et al.*, 2021; West *et al.*, 2005). Coupled sulfide-carbonate weathering is a transient CO₂ source on timescales up to the oceanic residence time of sulfate ($\sim 10^7$ years (Canfield, 2004)), conflicting with hypotheses that major orogenies drive atmospheric CO₂ sequestration and global cooling (e.g. Raymo and Ruddiman, 1992). Instead, CO₂ drawdown from weathering may be optimised at moderate erosion rates of around 0.07 mm yr⁻¹ (Bufe *et al.*, 2024). Below this optimum silicate weathering is supply-limited, suppressing CO₂ drawdown, and above it, silicate weathering becomes insensitive to erosion while CO₂ release from sulfide-carbonate weathering continues to increase (Bufe *et al.*, 2024).

The emerging view on the relationship between erosion, weathering, and long-term climate is further complicated by the consideration of oxidative weathering of rock (petrogenic) organic carbon (OC_{petro}), which constitutes an additional CO₂ source to the atmosphere (Petsch, 2014). CO₂ emissions from oxidation of OC_{petro} are estimated to be 62 - 86 Mt C year⁻¹ (Zondervan *et*

al., 2023), comparable to the input from volcanic degassing of 70 – 88 Mt C year⁻¹ (Plank and Manning, 2019) and drawdown from silicate weathering (90 – 140 Mt C year⁻¹ (Gaillardet *et al.*, 1999; Moon *et al.*, 2014). Recent work has demonstrated that CO₂ release from oxidation of OC_{petro} can negate the CO₂ sink from silicate weathering, and identifies basins with high denudation rates as hotspots for net CO₂ release through OC_{petro} oxidation (Zondervan *et al.*, 2023). There is growing evidence for a positive relationship between OC_{petro} oxidation and erosion rate (e.g. Ghazi *et al.*, 2022; Hilton *et al.*, 2014; Horan *et al.*, 2017; Ogrič *et al.*, 2023), however this behaviour has not been quantified independent of other influential variables such as runoff, temperature, or lithology. In oxic environments OC_{petro} oxidation occurs at faster rates than silicate mineral dissolution (Chang and Berner, 1999; White and Buss, 2013) and is therefore better able to keep pace with an enhanced supply of fresh materials. OC_{petro} oxidation could therefore diminish the magnitude of the CO₂ drawdown optimum observed by Bufe *et al.*, (2024), shift the optimum to lower erosion rates, and or extend the drawdown optimum behaviour beyond the timescale of the sulfide-carbonate weathering CO₂ source.

Here we use the rhenium proxy (Dalai *et al.*, 2002; Dellinger *et al.*, 2023; Grant *et al.*, 2025; Horan *et al.*, 2019) to quantify OC_{petro} oxidation across a three order-of-magnitude erosion-rate gradient in Taiwan with minimal accompanying variation in runoff, temperature, and lithology. We combine these results with published assessments of sulfide, carbonate, and silicate weathering across the same samples (Bufe *et al.*, 2021; Bufe *et al.*, 2024) to investigate the relative erosion sensitivities across the different weathering reactions. We quantify the carbon release from OC_{petro} oxidation across the erosion-rate gradient, and incorporate our estimate with existing carbon budgets (Bufe *et al.*, 2021; Bufe *et al.*, 2024) to assess the carbon cycle impact of OC_{petro} oxidation across erosion regimes and reaction timescales.

2. Materials and methods

2.1 Study setting

Taiwan is an active mountain belt formed through the late-Cenozoic convergence of the Luzon Arc of the Philippine Sea Plate and the eastern Eurasian continental margin (Teng, 1990). The Central Range forms the island's principal topography, with a northward increase in relief and erosion rates from foothills to steeper mountain slopes with frequent landslides (Dadson *et al.*, 2003; Fellin *et al.*, 2017) (Fig. 1). Taiwan has a sub-tropical climate, with most precipitation delivered during the typhoon season between June and October (Wu and Kuo, 1999).

Our wider study area spans the length of the Central Range, but is focused on its southernmost extent and the Hengchun peninsula (Fig. 1) where erosion rates have previously been constrained through an extrapolated regression between erosion rates from cosmogenic nuclides (Chen *et al.*, 2020; Fellin *et al.*, 2017) and the topographic metric normalised steepness index (ksn) (Bufe *et al.*, 2021). Estimates from this regression show a northward-increasing erosion-rate gradient of at least three orders-of-magnitude from < 0.004 to > 2 mm yr⁻¹ across the study area, with measured erosion rates ranging from 0.06 – 7.39 mm yr⁻¹ (Bufe *et al.*, 2021). In contrast to the wide range in erosion rate, average annual runoff and rainfall vary by less than three-fold within the study area (Bufe *et al.*, 2024; Dadson *et al.*, 2003; Lee *et al.*, 2006), and runoff and erosion rate are not well correlated ($R^2 = 0.31$).

Bedrock geology across the study area is dominated by passive margin sedimentary sequences with varying degrees of metamorphism. Most sample catchments are underlain by the Lushan and Pilushan formations of the Cenozoic Slate Belt, composed of argillaceous deposits metamorphosed to slate and phyllite (Lin and Chen, 2016) (Fig. 1). The southernmost catchments include sandstone and shale sequences of the Mutan formation, while catchments on the eastern margin of the Central Range are influenced by high grade metapelites and limestones of the Tananao schist (Lin and Chen, 2016). The distribution of OC_{petro} in the bedrock

is relatively uniform, with 0.2 – 0.4 weight % OC_{petro} in both the Tananao schist and Cenozoic Slate Belt (Hilton *et al.*, 2010). In the northernmost samples, where bedrock has undergone higher-grade metamorphism above ~ 450 °C, the OC_{petro} pool can be dominated by graphite (Beyssac *et al.*, 2007).

2.2 Samples

Archived water and solid samples were used from prior sampling campaigns in Taiwan (Fig. 1). Water samples (n = 215) were collected, filtered, and stored following standard procedures, with details provided by Bufe *et al.*, (2021). Most samples were collected from rivers and small streams, alongside some groundwater seepage samples. Prior ion chromatography and ICP-OES analyses of these samples has yielded an existing dataset of major anion and cation concentrations, as well as field data including pH, temperature, and dissolved oxygen (Bufe *et al.*, 2021). For subsets of the samples, previous work reported normalised steepness indices, ksn, and inferred erosion rates (n = 119), and estimated cation and anion concentrations from silicate, carbonate, and sulfide weathering as well as cyclic and hot spring inputs (n = 103) (Bufe *et al.*, 2021). We note that while the water samples represent a close-to-instantaneous measurement, the ¹⁰Be-derived erosion rates to which we compare are integrated over centennial to millennial timescales ((Fellin *et al.*, 2017), with transient perturbations possible on annual timescales (Chen *et al.*, 2020).

In addition, we used archive bedrock (n = 6) and bedload (n = 4) samples that were collected, homogenized, and stored following standard procedures (Hilton *et al.*, 2010). Previous work determined weight % of organic carbon by combustion with a CHN elemental analyser (Hilton *et al.*, 2010). Bedrock and bedload samples were used to determine the rhenium and OC_{petro} composition of the source lithologies of the sampled catchments. Following convention, bedload samples are considered here to represent the upstream catchment-averaged bedrock composition (e.g. Galy *et al.*, 2008b).

2.3 Rhenium concentration measurements

Dissolved rhenium concentrations were measured in Oxford by multi-Q-ICP-MS (Perkin Elmer NexION® 5000) broadly following established methods (e.g. Hilton *et al.*, 2021). Concentrations were determined by direct calibration against a set of seven standards matrix-matched to approximate river water composition. Samples and calibration standards were prepared in 2% HNO₃ and doped with an internal standard of 1 ppb rhodium to correct for instrumental drift and matrix variability. Measurement accuracy was assessed by repeated measurement of drinking and river water standard reference materials AQUA-1 and SLRS-6. AQUA-1 (n = 16) and SLRS-6 (n = 20) returned values of 11.3 and 14.6 ppt, each slightly above the ranges of prior characterisations 10.4 ± 0.8 (Yeghicheyan *et al.*, 2021) and 13.5 ± 0.2 ppt (Yeghicheyan *et al.*, 2019), but with overall accuracies better than 8.5%. Across the analysis, the average 2RSD of individual measurements (across 4 replicates of 18 sweeps) was 7.5%, while the average 2RSD of repeat measurements was 3.6%. All uncertainty metrics are small compared to the range of [Re]_{diss} in the samples (0.24 – 93.56 pmol L⁻¹), and sample concentrations are well above the average Re detection limit (0.06 pmol L⁻¹ or 0.01 ppt).

The rhenium concentrations of solid samples were determined broadly following established dissolution, purification, and measurement protocols (Dellinger *et al.*, 2020) (section [S.1](#)). Rhenium concentrations were analysed by multi-Q-ICP-MS following a similar procedure as for the water samples using pure Re solutions as calibration standards. Measurement accuracy and precision were assessed using reference materials AQUA-1 and SLRS-6, with similar results as for the waters. USGS reference material BCR-2 (Columbia River Basalt) was processed alongside the solid samples to establish the yield of the digestion and purification procedure. BCR-2 returned a Re concentration of 12.56 ppb, close to the range of 11.56 ± 0.74 ppb from a

prior characterisation (Jochum *et al.*, 2016), for a yield of 108%. The procedural blank was 3 pg, which is < 1% of the average mass of Re in the samples.

3. Rhenium results and proxy application

3.1 Dissolved rhenium

Dissolved rhenium, $[\text{Re}]_{\text{diss}}$, ranges from 0.29 – 96.11 pmol L^{-1} , with a median of 5.49 pmol L^{-1} ($n = 215$). This range is similar to the global range in exorheic river waters of 1.10 – 120.00 pmol L^{-1} , and the median of 5.79 pmol L^{-1} ($n = 2110$) (Ghazi *et al.*, 2024). The range of $[\text{Re}]_{\text{diss}}$ in this study overlaps and extends a range previously reported for a smaller set of Taiwan rivers of 4.7 – 25.4 pmol L^{-1} (Hilton *et al.*, 2014). Most samples (71%) had $[\text{Re}]_{\text{diss}}$ below 10 pmol L^{-1} (Fig. S2). Of those samples with $[\text{Re}]_{\text{diss}} > 10 \text{ pmol L}^{-1}$, catchments underlain by Tananao schist bedrock are overrepresented, potentially reflecting relative Re enrichment in the Tananao schist compared to other source lithologies. Samples with lower $[\text{Re}]_{\text{diss}} < 3 \text{ pmol L}^{-1}$ are mostly from the Lushan formation. $[\text{Re}]_{\text{diss}}$ shows no discernible correlation with mean annual runoff and a weak positive correlation with the regression-based inferred erosion rate (Bufe *et al.*, 2021) ($R^2 = 0.29$), however this relationship also exhibits substantial scatter.

3.2 Solid rhenium and OC_{petro}

Bedrock rhenium concentration, $[\text{Re}]_{\text{bedrock}}$, ranges from 0.07 – 2.60 ppb with an average of 0.69 ppb ($n = 9$) (Fig. 2). The highest $[\text{Re}]_{\text{bedrock}}$ was from a Lushan formation black shale, while the lowest $[\text{Re}]_{\text{bedrock}}$ was from a Pilushan formation sandstone. The $[\text{Re}]_{\text{bedrock}}$ range is consistent with the median $[\text{Re}]$ in a compilation of non-shale sedimentary rocks of 0.11 ppb ($n = 16$), but lower than median values from grey and black shales of 3.20 ($n = 754$) and 23.25 ($n = 2234$) ppb (Ghazi *et al.*, 2024). Bedload rhenium concentrations, $[\text{Re}]_{\text{bedload}}$, were more uniform than $[\text{Re}]_{\text{bedrock}}$, with a range of 0.24 – 0.71 ppb, and had a lower average concentration of 0.49 ppb ($n = 6$) (Fig. S3). This range is consistent with previous analysis of Taiwan bedloads (Hilton *et al.*,

2014) and the median river sediment [Re] of 0.38 ppb from a large compilation (n = 296) (Ghazi et al., 2024).

The mean $[OC/Re]_{\text{solids}}$ across bedrock and bedload samples was $1.39 \cdot 10^7 \text{ g g}^{-1}$ (n = 10), with a range of $0.14 - 6.03 \cdot 10^7 \text{ g g}^{-1}$ (Fig. 2). Bedload samples lay within the wider range of bedrock compositions, with an average $[OC/Re]_{\text{bedload}}$ of $0.62 \cdot 10^7 \text{ g g}^{-1}$ (n = 4), and a range of $0.41 - 0.78 \cdot 10^7 \text{ g g}^{-1}$. The measured $[OC/Re]_{\text{solids}}$ range is consistent with a prior average from Taiwan bedloads of $0.27 \cdot 10^7 \text{ g g}^{-1}$ (n = 3) (Hilton et al., 2014). The $[OC/Re]$ of the black shale sample is $0.14 \cdot 10^7 \text{ g g}^{-1}$, which is similar to measurements of Devonian and Jurassic black shales (Cohen et al., 1999; Jaffe et al., 2002; Selby and Creaser, 2003) (Fig. 2). Notably, bedrocks and bedloads in this study are enriched in both Re and OC_{petro} relative to Taiwan soils when normalised to immobile element aluminium (Hilton et al., 2014), confirming their coupled loss during weathering (Fig. S4) (Grant et al., 2025).

3.3 Source of dissolved rhenium

To quantify OC_{petro} oxidation fluxes from dissolved Re, $[Re]_{\text{diss}}$ measurements are corrected for atmospheric inputs (denoted $[X^*]$) and partitioned between the candidate geological sources; sulfides, silicates, and OC_{petro} . We follow the approach of previous studies and define endmember ratios of $[Re/Cl]_{\text{rain}}$, $[Re^*/SO_4^*]_{\text{sulfides}}$, and $[Re^*/Na^*]_{\text{silicates}}$ to estimate and subtract contributions to $[Re]_{\text{diss}}$ from rain, sulfide, and silicate sources, and attribute the remaining Re to OC_{petro} , denoted $[Re]_{OC_{\text{petro}}}$ (Dellinger et al., 2023; Horan et al., 2019) (Fig. 3A) (section S.2).

Sulfide weathering fronts are generally deeper than those of other minerals (e.g. Brantley et al., 2013), therefore samples strongly influenced by deep groundwater may be expected to reflect sulfide protolith $[Re^*/SO_4^*]$. Seepage and groundwater samples in this dataset span a two order-of-magnitude range in $[Re^*/SO_4^*]$ and define a mixing line with positive correlation ($R^2 = 0.44$), which overlaps the range spanned by Andean tributaries draining sulfur-rich rocks

(Dellinger *et al.*, 2023) (Fig. 3A). The [Re/S] range of Taiwan solids is compressed compared to the range in [Re*/SO₄*] of waters, with solid compositions restricted to the high end of the water samples. The range in [Re*/SO₄*] of waters therefore cannot be explained by mixing between our solid samples. Instead, we suggest that the seepage and groundwater samples reflect the influence of a sulfide-rich local endmember, with the lowest [Re*/SO₄*] value providing a representative endmember estimate. This sample (TWS15-92) has [Re*/SO₄*] of 9×10^{-5} pmol μmol^{-1} and a high [SO₄*], in the 92nd percentile, indicating significant sulfide weathering.

Excluding a black shale, Taiwan solid samples plot in a cluster with [Re*/Na*] from 0.002 – 0.02 pmol μmol^{-1} . The cluster overlaps with literature silicate endmembers including silicate minerals (Burton *et al.*, 1999; Burton *et al.*, 2002), Icelandic river waters (Pogge von Strandmann *et al.*, 2008), and Andes granites and monolithologic silicate bedloads and small streams (Dellinger *et al.*, 2023) (Fig. 3A). The river water sample with the lowest [Re*/Na*] (TW15HG-41) plots at the low end of the literature data, with a [Re*/Na*] of 5×10^{-4} pmol μmol^{-1} , and [Na*] in the 91st percentile. We note the presence of a hot spring sample with a more extreme [Re/Na]_{diss} of 4×10^{-5} pmol μmol^{-1} , however we take a conservative approach and use TW15HG-41 to represent the silicate endmember. Acknowledging uncertainty in the choice of endmembers for [Re*/SO₄*]_{sulfides} and [Re*/Na*]_{silicates}, we use a Monte Carlo approach to quantify uncertainty in the source partition (Dellinger *et al.*, 2023) (Fig. 3A) (section S.2.3).

We find that in 95% of samples most [Re*] is derived from OC_{petro}, with a median proportion $F_{\text{Re-OCpetro}}$ of 0.94 ($n = 212$), and $F_{\text{Re-OCpetro}} > 0.80$ for 83% of samples (Fig. 3B). These results are consistent with previous studies taking this approach in the Andes, Swiss Alps, Colorado Rockies, and the Mackenzie basin (Dellinger *et al.*, 2023; Hilton *et al.*, 2021; Horan *et al.*, 2019) and support growing consensus in favour of the Re proxy for OC_{petro} oxidation (Grant *et al.*, 2025).

3.4 OC_{petro} oxidation flux

The OC_{petro} oxidation flux, $J_{OC_{\text{petro-ox}}}$, was quantified for each sample based on equation (1) (e.g. Dellinger et al., 2023; Hilton et al., 2014; Horan et al., 2017; Märki et al., 2021). Since runoff data is limited to < 50% of the samples, we first calculate $J_{OC_{\text{petro-ox}}}$ ($\mu\text{mol L}^{-1}$) in per-Litre space, where $[Re]_{OC_{\text{petro}}}$ ($\mu\text{mol L}^{-1}$) is the concentration of OC_{petro} -derived Re (section 3.3, S.2), $[OC/Re]_{\text{solids}}$ (10^7 g g^{-1}) is the OC:Re ratio of the weathering material (section 3.2), and f_{graphite} is the proportion of OC_{petro} present as graphite and assumed resistant to oxidation (Galy et al., 2008a). Assuming a 1:1 stoichiometry, $J_{OC_{\text{petro-ox}}}$ also represents the molar CO_2 release associated with OC_{petro} oxidation.

$$J_{OC_{\text{petro-ox}}} = [Re]_{OC_{\text{petro}}} \times (1 - f_{\text{graphite}}) \times \left[\frac{OC}{Re} \right]_{\text{solids}} \quad (1)$$

We make this calculation over each sample within the Re source partition Monte Carlo simulation to propagate the simulated $[Re]_{OC_{\text{petro}}}$ and incorporate uncertainties in $[OC/Re]_{\text{solids}}$ and f_{graphite} (section S.2.3).

4. Modelling erosion sensitivity

Erosion is a first-order control on the rates of chemical weathering reactions because erosion controls the supply of fresh minerals to the weathering zone (e.g. Gaillardet *et al.*, 1999). Weathering fluxes can therefore be expected to scale with erosion rate according to the erosion sensitivity of the weathering reaction in question. However, weathering reactions are also controlled by the availability of acidic fluids and kinetic factors linked to intrinsic mineral reactivity and temperature (Lasaga *et al.*, 1994; Maher and Chamberlain, 2014). These hydrologic and kinetic effects can decouple weathering fluxes from erosion and moderate erosion sensitivity across erosion rates (Liu *et al.*, 2024; West *et al.*, 2005). Weathering rates are therefore commonly conceptualised as limited either by suppressed mineral supply at low erosion rates (“supply-limited”, with high erosion sensitivity), or by fluid availability or reaction kinetics at high erosion rates (“weathering-limited”, with low erosion sensitivity) (e.g. West *et al.*, 2005). Here, for a subset of the samples ($n = 85$), we fit a power law to quantify the average erosion sensitivity of OC_{petro} oxidation, and a multiparameter weathering model capturing the transition from supply-limited to weathering-limited behaviour to constrain any variation in erosion sensitivity of OC_{petro} across the erosion rate range.

4. 1 Power law

To determine the average erosion sensitivity of $Re_{OC_{\text{petro}}}$ and OC_{petro} oxidation fluxes we fit a power law of the form $\omega = \lambda E^\alpha$ to the weathering flux (ω) data where λ and α are constants, and α represents the average erosion sensitivity over the full erosion flux range. The best fit parameters were identified by minimising the sum of squared residuals (RSS) in $\log_{10}$ space. Best-fit parameters λ and α are reported alongside their standard errors.

4.2 Multiparameter model

To resolve potential changes in the sensitivity of $\text{Re}_{\text{OCpetro}}$ fluxes and associated OC_{petro} oxidation to erosion across the range of erosion rates, we fit a widely adopted model (e.g. West, 2012) relating chemical weathering flux, ω ($\text{t km}^{-2} \text{ year}^{-1}$), and physical erosion flux, E ($\text{t km}^{-2} \text{ year}^{-1}$) (2).

$$\omega = \chi_m E \left(1 - \exp \left(\frac{-K f(T) f(q_w) \left(\frac{z}{E} \right)^{(\sigma+1)}}{\sigma+1} \right) \right) \quad (2)$$

The variable χ_m (-) represents the fraction of weatherable Re or OC_{petro} in the bedrock. K ($\text{year}^{-(\sigma+1)}$) is a constant which characterises the fundamental properties of mineral weathering and their dependence on grain surface area. $f(T)$ and $f(q_w)$ are functions of temperature T (K) and runoff q_w (m year^{-1}), while z (t km^{-2}) is a length scale corresponding to the effective depth of the weathering zone. σ (-) describes the change in weathering rates with residence time of material in the weathering zone, where $\sigma = -1$ implies a linear decrease in rates with time, and $\sigma = 1$ implies a linear increase (West, 2012).

At low erosion rates, the exponential term is small, and the relationship is dominated by the linear scaling with erosion, $\omega \propto \chi_m E$. At high erosion rates, the exponential term dominates the relationship, and the variable σ can allow non-linear behaviour in which the function shallows ($-1 < \sigma < 0$), plateaus ($\sigma = 0$), or declines ($0 < \sigma \leq 1$). The deviation from a linear increase at high erosion rates represents the transition in weathering reactions from supply-limited towards weathering-limited behaviour.

This model has previously been used to characterise the erosion sensitivity of silicate, carbonate, and sulfide weathering fluxes in Taiwan using a subset of samples from this study (Bufe *et al.*, 2024). To maximise consistency and enable comparison between our results and results of prior work, we follow the approach of Bufo *et al.* (2024) to extend this model to our results for $\text{Re}_{\text{OCpetro}}$ and OC_{petro} oxidation fluxes. The model was decomposed into four parameters, $p_1 = \chi_m$, $p_2 = [K f(q) f(T)] / [\sigma + 1]$, $p_3 = z$, and $p_4 = \sigma + 1$. Following Bufo *et al.* (2024)

and West (2012), we use wide but physically reasonable bounds of $0 < p_1 < 0.3$; $10^{-10} < p_2 < 10^{-2}$; $10^4 < p_3 < 10^{10}$; and $0 < p_4 < 2$. Best-fit parameters were found by minimising the sum of squared residuals (RSS) between the model and the data in $\log_{10}$ space. Uncertainties were determined by identifying ‘accepted’ fits in a region of parameter space for which model fits were statistically indistinguishable from the best fit, given the noise in the data (section [S.4](#)).

5. Discussion

5.1 Erosion sensitivity of OC_{petro} oxidation

Erosion has been widely understood as a first-order control on OC_{petro} oxidation rates. In oxygenated weathering environments OC_{petro} oxidation proceeds faster than silicate mineral dissolution (Chang and Berner, 1999; White and Buss, 2013), and is better able to keep pace with increased supply of material by erosion, similar to sulfide minerals (Bolton *et al.*, 2006). This relationship is supported by observed correlation between $[Re]_{\text{diss}}$ and suspended sediment yields (Hilton *et al.*, 2014), however, it has not yet been possible to quantify the erosion sensitivity due to small sample sets (Hilton *et al.*, 2021), and co-variation of confounding variables such as runoff and lithology, which also impact weathering fluxes. Our study area exhibits a three order-of-magnitude range in erosion rate, alongside a less than three-fold range in runoff, and relatively invariant lithology. Coupled with our modelling approach, this provides a framework in which to isolate the effect of erosion on OC_{petro} oxidation and quantify the erosion sensitivity of OC_{petro} largely independent of other environmental variables.

$[Re]_{OC_{\text{petro}}}$ and associated OC_{petro} oxidation fluxes showed a nonlinear positive scaling with erosion rate (Fig. 4). The average erosion rate sensitivities (α) of $[Re]_{OC_{\text{petro}}}$ and OC_{petro} oxidation were 0.52 ± 0.06 and 0.54 ± 0.06 , close to a square root relation. For both $[Re]_{OC_{\text{petro}}}$ and OC_{petro} oxidation fluxes the best-fit of the multiparameter model was indistinguishable from the power-law fit across the erosion flux range of the samples. The similarity in the fits arises due to the lack of curvature in the $[Re]_{OC_{\text{petro}}}$ and OC_{petro} oxidation flux data (Fig. 4). The multiparameter model was unable to constrain a transition from wholly supply-limited ($\alpha = 1$) towards wholly weathering-limited ($\alpha = 0$) behaviour, and instead fits $[Re]_{OC_{\text{petro}}}$ and OC_{petro} oxidation fluxes exclusively on the transitional limb of the model while maintaining a significant erosion sensitivity. Physically, these observations suggest that OC_{petro} oxidation may never be either fully supply-limited or fully weathering-limited across the constrained range in erosion rates, and

that both erosion and kinetic or hydrologic factors are important throughout the gradient. This finding is consistent with outcrop measurements of OC_{petro} oxidation that suggest strong temperature and hydrological controls at high erosion rates (Roylands *et al.*, 2022; Soulet *et al.*, 2021), and supports the notion that OC_{petro} oxidation is climate sensitive, and a potential positive feedback in the geological carbon cycle (Stow *et al.*, 2026).

Alongside $[\text{Re}]_{\text{OC}_{\text{petro}}}$ and OC_{petro} oxidation fluxes, the power-law and multiparameter models were also run for silicate, carbonate, and sulfide weathering fluxes to reproduce the results of Bufe *et al.* (2024) (Fig. S7). Silicate weathering fluxes have a weakly negative average erosion sensitivity ($\alpha = -0.16 \pm 0.04$) with a transition from the supply limit to the weathering limit at erosion rates of $\sim 0.05 \text{ mm yr}^{-1}$. Carbonate and sulfide weathering fluxes have stronger positive erosion sensitivities of 0.31 ± 0.03 and 0.69 ± 0.05 respectively, and transition from the supply limit towards the weathering limit at erosion rates $\sim 0.1 - 0.5 \text{ mm yr}^{-1}$, with sulfide fluxes maintaining a significant erosion sensitivity between limits. These results are in close agreement with those reported in Bufe *et al.* (2024).

$[\text{Re}]_{\text{OC}_{\text{petro}}}$ and OC_{petro} oxidation fluxes behave similarly to sulfide weathering fluxes, but with a slightly weaker erosion dependence. Sulfide oxidation can respond strongly to physical erosion due to its rapid reaction kinetics (Williamson and Rimstidt, 1994). In contrast, OC_{petro} reacts more slowly than sulfides, as evidenced in lab experiments (Chang and Berner, 1999) and the relative depth of their reaction fronts in weathering profiles (Petsch *et al.*, 2000). The pool of OC_{petro} can contain recalcitrant components such as graphite which is common in Taiwan (Beyssac *et al.*, 2007; Sparkes *et al.*, 2020). Graphite is commonly assumed to be unreactive (Galy *et al.*, 2008a), however subsurface measurements from Taiwan suggest that metamorphic OC_{petro} pools could remain reactive despite potential graphitisation (Lien *et al.*, 2025b). OC_{petro} can also be hosted in mineral inclusions or aggregates which offer physical shielding from oxidation (Lien *et al.*, 2025a). Microbial processes may also influence the relative erosion

sensitivities of these reactions, with evidence for both microbially mediated OC_{petro} and sulfide oxidation in Taiwanese catchments (Hemingway *et al.*, 2018; Wang *et al.*, 2024). The specific balance of these limiting factors could impede the relative ability of OC_{petro} oxidation to keep pace with increased supply via physical erosion when compared with sulfide oxidation, and explain their contrasting erosion sensitivities.

The degree of contrast in erosion sensitivities of sulfide and OC_{petro} oxidation may be generally applicable or local to Taiwan. Taiwan OC_{petro} could have a similar effective reactivity to Himalayan OC_{petro} as both locations have prevalent graphite (Galy *et al.*, 2008a). Meanwhile other locations have little graphite (Bouchez *et al.*, 2010; Dellinger *et al.*, 2023) and their OC_{petro} pools could thereby have a greater effective reactivity and erosion sensitivity. Taiwan also has amongst the world's highest annual sediment yield, with most sediment delivered via landslides during typhoon events (Chen *et al.*, 2020), potentially limiting oxidative residence time and suppressing erosion sensitivity of OC_{petro} compared to other locations. These considerations could reconcile our results with suspended sediment yield-based estimates of OC_{petro} oxidation erosion sensitivity in compilations including the Himalaya, Southern Alps, and the Mackenzie basin that suggest higher sensitivities with $\alpha \sim 0.7$ (Hilton and West, 2020; Horan *et al.*, 2017).

5. 2 Erosion sensitivity of net weathering carbon budget

Recent work has presented a CO_2 drawdown optimum at moderate erosion rates over timescales where CO_2 drawdown from silicate weathering competes with CO_2 release from coupled sulfide-carbonate weathering (Bufe *et al.*, 2024). However, it is not yet clear how OC_{petro} oxidation influences this drawdown optimum, or whether this behaviour persists on timescales preceding or beyond the influence of coupled sulfide-carbonate weathering (Torres *et al.*, 2016).

Oxidation of OC_{petro} has also been acknowledged as a substantial CO_2 source that can counteract silicate weathering, especially in regions of high denudation (Hilton *et al.*, 2014;

Horan *et al.*, 2019; Zondervan *et al.*, 2023) ([Fig. 6A](#)). However, studies that assess carbon budgets integrating the full suite of weathering reactions including OC_{petro} oxidation (e.g. Märki *et al.*, 2021; Wang *et al.*, 2025) are comparatively rare, and the influence of erosion on complete net budgets has yet to be quantified. Since the mechanism of CO_2 sourcing by OC_{petro} oxidation is decoupled from the acid-hydrolysis reactions of silicate mineral weathering and is not subject to later oceanic equilibration, CO_2 release is instantaneous and relevant across timescales. The response of OC_{petro} oxidation to erosion could therefore be a key control on the net carbon budget of weathering across timescales. Here we integrate our model results describing the sensitivity of OC_{petro} oxidation and associated CO_2 release to erosion with the existing CO_2 budget of Bufe *et al.*, (2024) and present the first assessment of the erosion sensitivity of the net weathering carbon budget including OC_{petro} .

Following the approach of Bufe *et al.* (2024) we calculate the net CO_2 flux to the atmosphere F_{CO_2} ($t\ km^{-2}\ yr^{-1}$) for each sample and for model fits from the multiparameter weathering model. Best-fit F_{CO_2} was calculated from combination of the best fit OC_{petro} oxidation, and silicate, carbonate, and sulfide weathering fluxes. Uncertainties were estimated by propagating the suite of accepted fits of the weathering model; over 3000 iterations accepted model fits were paired at random to generate an array of accepted F_{CO_2} fits, from which median and quartiles were calculated. F_{CO_2} was calculated over three timescales in which different weathering reactions are at play according to (3-5) (Bufe *et al.*, 2021; Torres *et al.*, 2016).

$$[F_{CO_2}]_{st} = ([SO_4]^{eq} - 0.5[Cat_{carb}]^{eq} - [Cat_{sil}]^{eq}) + J_{OC_{\text{petro-ox}}} \quad (3)$$

$$[F_{CO_2}]_{mt} = (0.5([SO_4]^{eq} - [Cat_{sil}]^{eq})) + J_{OC_{\text{petro-ox}}} \quad (4)$$

$$[F_{CO_2}]_{lt} = -0.5[Cat_{sil}]^{eq} + J_{OC_{\text{petro-ox}}} \quad (5)$$

On timescales shorter than marine carbonate burial (st – short timescales $< \sim 10$ kyr (Zeebe and Westbroek, 2003)), the net carbon budget is determined by the balance of bicarbonate generation through sulfuric and carbonic acid weathering of silicate and carbonate rocks, and

OC_{petro} oxidation. Our data show a transition from CO₂ sink to CO₂ source behaviour with increasing erosion rates. The model fits predict a strong CO₂ -drawdown optimum at approximately 0.12 mm yr⁻¹, and a transition to net source behaviour at ~ 1.35 mm yr⁻¹ (Fig. 5A).

On timescales longer than marine carbonate burial but shorter than marine sulfate reduction (*mt* – medium timescales ~ 10 kyr - 10 Myr (Canfield, 2004)) alkalinity production by weathering is partially balanced by carbonate precipitation, and coupled sulfide-carbonate weathering becomes a CO₂ source. Hence, the net budget is determined by the balance of carbonic acid weathering of silicates, versus coupled sulfide-carbonate weathering and OC_{petro} oxidation. Our data show approximately net neutral behaviour at low to moderate erosion rates, and a transition to substantial CO₂ source behaviour at moderate to high erosion rates (Fig. 5B). The model fits predict the transition to CO₂ source behaviour at approximately 0.16 mm yr⁻¹, but do not resolve a convincing drawdown optimum outside error of net CO₂ neutral.

The form of the erosion-CO₂ flux function is driven by the transition of silicate weathering from the supply limit to the weathering limit at moderate erosion rates (Fig. S7), while coupled sulfide-carbonate and OC_{petro} oxidation maintain supply-limited behaviour to higher erosion rates (Fig. 4, S5). The silicate CO₂ sink flux does not increase at higher erosion rates while the oxidative source fluxes do, resulting in net source behaviour above a threshold that strengthens with increasing erosion rates. Our results show a more modest CO₂ drawdown optimum than the results of Bufe *et al.* (2024), because the inclusion of OC_{petro} translates our data and model fits closer to the CO₂ source field. Even at low erosion rates where OC_{petro} oxidation fluxes are relatively small, they are enough to shift some samples into the CO₂ source field, and many others within error of neutrality. While OC_{petro} oxidation provides an additional CO₂ source over this timescale (10⁴ -10⁷ yr), we note that the CO₂ source from coupled sulfide-carbonate weathering remains the largest flux overall (Fig. 6B).

On timescales beyond marine sulfate reduction (*lt* – long timescales > 10Myr), the net carbon budget is reduced to the balance between carbonic acid weathering of silicates, and OC_{petro} oxidation (Fig. 6A). Our model fits show a transition to CO_2 source behaviour at $\sim 0.52 \text{ mm yr}^{-1}$, with a modest drawdown optimum at $\sim 0.07 \text{ mm yr}^{-1}$ (Fig. 5C). As for medium timescales, the long-timescale functional form arises due to the sustained erosion dependence of OC_{petro} oxidation while silicate weathering reaches the weathering limit. The CO_2 drawdown optimum is more pronounced on this timescale than for the medium timescale due to the absence of the CO_2 source from sulfide-carbonate weathering, and the lesser erosion sensitivity of OC_{petro} oxidation compared to sulfide weathering.

5.3 Incorporating the OC_{bio} burial carbon sink

The erosion, export, and burial of modern particulate biospheric carbon (OC_{bio}) is a substantial CO_2 sink in the geological carbon cycle that is not captured by assessment of dissolved river chemistry and rock weathering reactions. The global terrestrial OC_{bio} burial flux has been estimated between 40 – 80 Mt C yr^{-1} , similar to the OC_{petro} oxidation flux of 62 – 88 Mt C yr^{-1} (Hilton and West, 2020; Zondervan *et al.*, 2023). The terrestrial OC_{bio} carbon sink is moderated by erosion, which governs the amount of OC_{bio} exported from soils, and the supply of sediment within which OC_{bio} is preserved (e.g. Galy *et al.*, 2015). Export and burial of OC_{bio} could therefore act as an erosion-sensitive CO_2 sink that could balance CO_2 sources from OC_{petro} oxidation and coupled sulfide-carbonate weathering at high erosion rates. Due to the lack of OC_{bio} burial constraints on specific samples in this study, we cannot directly incorporate this flux into our sample-based net CO_2 analysis. Instead, we use a global compilation of OC_{bio} and suspended sediment yields (Hilton *et al.*, 2025) to derive an erosion sensitivity of OC_{bio} burial (section S.5), and incorporate this with our modelled carbon budgets to assess the overall response of the geological carbon cycle to erosion.

Our results show that OC_{bio} burial can substantially offset the CO_2 source from OC_{petro} oxidation at high erosion rates on long timescales (Fig. 7). The erosion sensitivity (α) of OC_{bio} burial is 0.62 ± 0.03 , greater than the sensitivity of OC_{petro} oxidation of 0.54 ± 0.06 (Fig. 4, S6A). This means that at high erosion rates OC_{bio} burial outpaces OC_{petro} oxidation, forcing a CO_2 release optimum at high erosion rates, above which OC_{bio} burial dominates and CO_2 release declines. The overall shape of the best fit long-term net carbon budget therefore appears to contain both a CO_2 drawdown and a CO_2 release optimum, controlled respectively by the threshold at which silicate weathering becomes weathering-limited while OC_{petro} oxidation continues to increase, and the threshold at which OC_{bio} burial overtakes OC_{petro} oxidation (Fig. 7). On short and medium timescales, the CO_2 sink from OC_{bio} burial is insufficient to balance the additional CO_2 source from sulfide-carbonate weathering, but does diminish the magnitude of the overall CO_2 release at high erosion rates (Fig. S9).

5. 4 Implications for the carbon budget of mountain building

Our results present challenges for hypotheses invoking mountain building as a driver of global cooling, because high erosion rates favour CO_2 release through OC_{petro} oxidation and sulfide-carbonate weathering (Fig. 5). A common resolution is that enhanced burial of terrestrial OC_{bio} at high erosion rates could counteract these CO_2 source fluxes, and result in net CO_2 drawdown or neutral behaviour (Galy *et al.*, 2015; Hilton *et al.*, 2014). Our results show that this mechanism is possible, particularly on timescales beyond the influence of sulfide-carbonate weathering. However, in mountain ranges such as Taiwan exposing sedimentary rocks rich in OC_{petro} and sulfide minerals, this offset may be effective only at relatively high OC_{bio} preservation efficiencies (Fig. 7, S7), suggesting that substantial offset of CO_2 release through OC_{bio} burial may not occur in all mountain belts.

Mountain ranges within the optimum erosion rate for CO_2 drawdown may balance CO_2 release from highly erosive settings. Upland areas with sustained moderate erosion rates within the CO_2

drawdown optimum ($\sim 0.05 - 2 \text{ mm yr}^{-1}$, Fig. 5) may represent more effective long-term CO_2 sinks than rapidly eroding collisional orogens. This mechanism could be especially effective in areas with significant exposure of mafic rocks which could have a greater erosion sensitivity than other silicates through their more rapid reaction kinetics (Macdonald *et al.*, 2019). However, the CO_2 drawdown potential at optimum erosion rates remains less than the CO_2 release potential at high erosion rates (Fig. 5), therefore requiring widespread regions favouring CO_2 sequestration to counter the CO_2 release from highly erosive regions. Downstream floodplain regions could provide conditions for further silicate weathering (Lupker *et al.*, 2012), however OC_{petro} oxidation can also be sustained over floodplain reaches (Bouchez *et al.*, 2010; Dellinger *et al.*, 2023), and during submarine transport, enabling further CO_2 release (Lien *et al.*, 2025a). Ultimately, the net impact of downstream processes on CO_2 fluxes from weathering remains poorly constrained.

Our results also imply that the net CO_2 budget of a mountain range could change over the course of an orogeny as erosion rates vary (Egholm *et al.*, 2013), and different weathering reactions are favoured. CO_2 drawdown via silicate weathering could be favoured in the early stages of mountain building before significant relief is generated, or in latter stages once tectonic forces have waned and uplift and erosion may be reduced or sedimentary cover has diminished. Meanwhile during periods of maximum uplift and erosion, oxidative CO_2 release through sulfide-carbonate weathering and OC_{petro} oxidation could be enhanced, albeit potentially offset in part through terrestrial OC_{bio} burial.

Our quantification of the erosion sensitivity of OC_{petro} extends previous work by showing how oxidation of OC_{petro} modifies the erosion-dependent carbon budget of mountain belts. Since OC_{petro} -bearing sedimentary rocks are widespread in active mountain belts inverting passive margin sedimentary sequences (Zondervan *et al.*, 2023), our findings could extend beyond Taiwan. However, broader application requires improved constraints on global and site-specific

OC_{petro} abundance and reactivity, terrestrial OC_{bio} export and preservation efficiency, and the role of downstream processes such as floodplain weathering in modifying the catchment-scale carbon budget. Further work is additionally needed to constrain CO₂ fluxes associated with dissolved organic carbon (DOC) transfers. Resolving these uncertainties could allow erosion-dependent carbon cycle models to move beyond individual case studies and towards a more general assessment of the impact of mountain building in the long-term carbon cycle.

6. Conclusions

We apply the rhenium (Re) proxy over a three-order-of-magnitude erosion-rate gradient in Taiwan, with minimal covariation in runoff, temperature, or lithology, to quantify the erosion sensitivity of OC_{petro} oxidation for the first time. Our validation of the Re proxy confirms that Re is derived from OC_{petro} , thereby supporting its continued application as a proxy for OC_{petro} oxidation.

OC_{petro} oxidation exhibits a positive erosion sensitivity described by $\omega \propto E^{\alpha}$, where $\alpha = 0.52$, approximating a square-root relationship. This contrasts with the stronger erosion sensitivity of sulfide oxidation ($\alpha = 0.69$), likely reflecting faster reaction kinetics, and the weakly negative erosion sensitivity of silicate weathering ($\alpha = -0.16$), reflecting slow reaction kinetics and the onset of weathering-limited behaviour at relatively low erosion rates (Bufe *et al.*, 2024).

These contrasting erosion sensitivities result in a changing balance of CO_2 sources and sinks across the gradient in erosion rate. While the CO_2 source from sulfide-carbonate weathering is larger than from OC_{petro} oxidation, OC_{petro} oxidation is sufficient to diminish the strength of the CO_2 drawdown optimum identified at moderate erosion rates in previous work (Bufe *et al.*, 2024). Furthermore, OC_{petro} oxidation provides an additional CO_2 source that competes with silicate weathering on timescales beyond the influence of sulfide-carbonate weathering, and extends the CO_2 drawdown optimum behaviour to these longer timescales. Terrestrial OC_{bio} burial provides an erosion sensitive CO_2 sink that extends across timescales and can partially offset CO_2 release from oxidative weathering at high erosion rates, given substantial export and preservation efficiency.

Our results highlight that the net carbon budget of weathering during mountain building is governed by the competing erosion sensitivities of both oxidative and silicate weathering. Incorporating OC_{petro} oxidation reveals that rapidly eroding mountain belts can act as CO_2

548 sources on the timescales of orogenesis, especially where OC_{petro} is sufficiently reactive, silicate
549 mineral dissolution is weathering-limited, and terrestrial OC_{bio} preservation is incomplete.
550 Improved constraints on the abundance and reactivity of OC_{petro} , erosion sensitivity of OC_{bio}
551 burial and mafic silicate mineral dissolution, and processes downstream of active mountain
552 belts are therefore critical to establishing the generalised sensitivity of Earth's long term carbon
553 cycle to erosion.

554

Figures

Figure 1 – Taiwan map (samples, geology, relief, rivers)

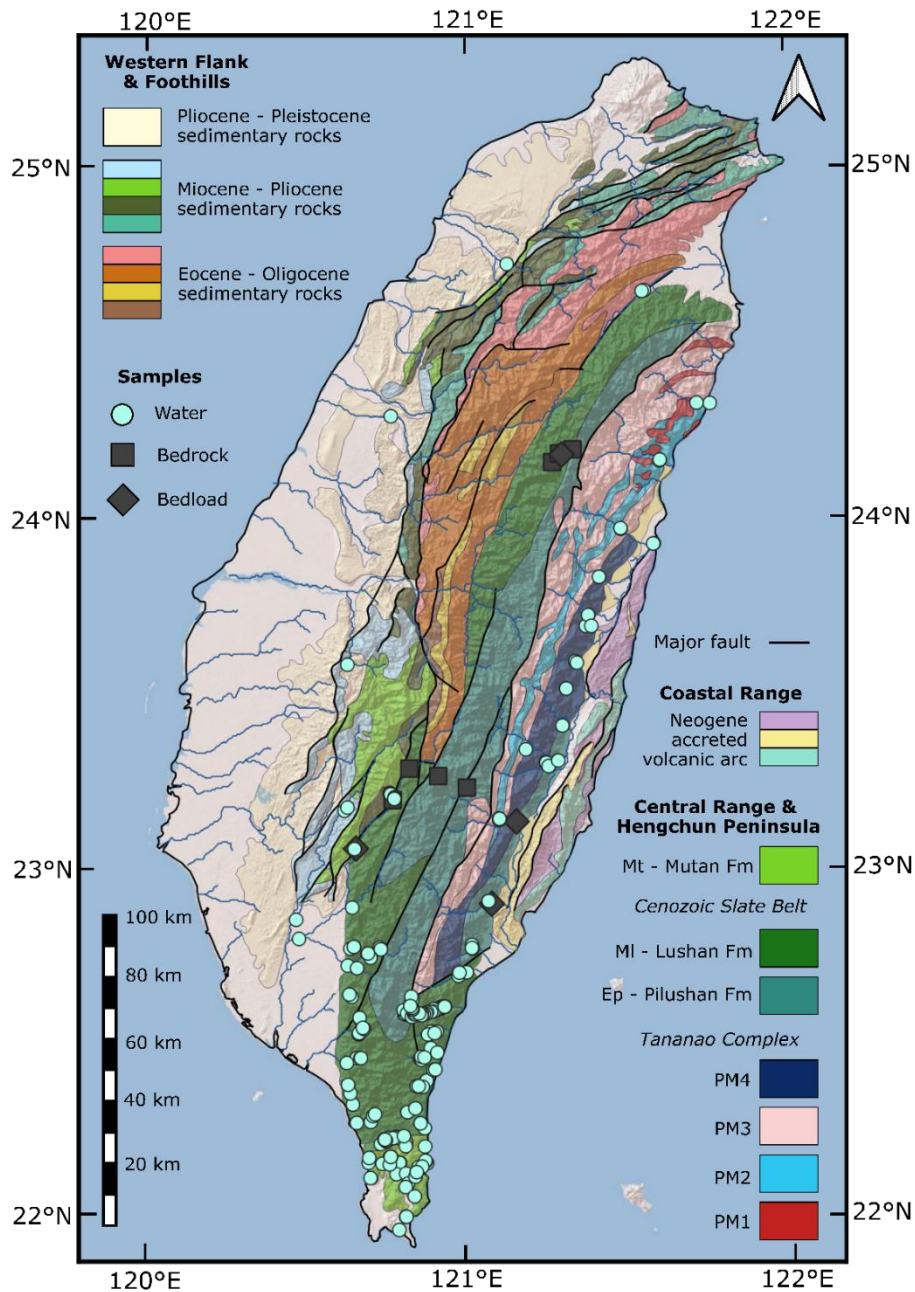


Fig. 1. Map of Taiwan with sample locations, geology, relief, and major rivers. Blue circles show the locations of water samples and dark grey squares and rhombuses show the locations of bedrock and bedload samples. The sketch geological map is adapted from Hilton *et al.*, (2010) and Lin and Chen (2016).

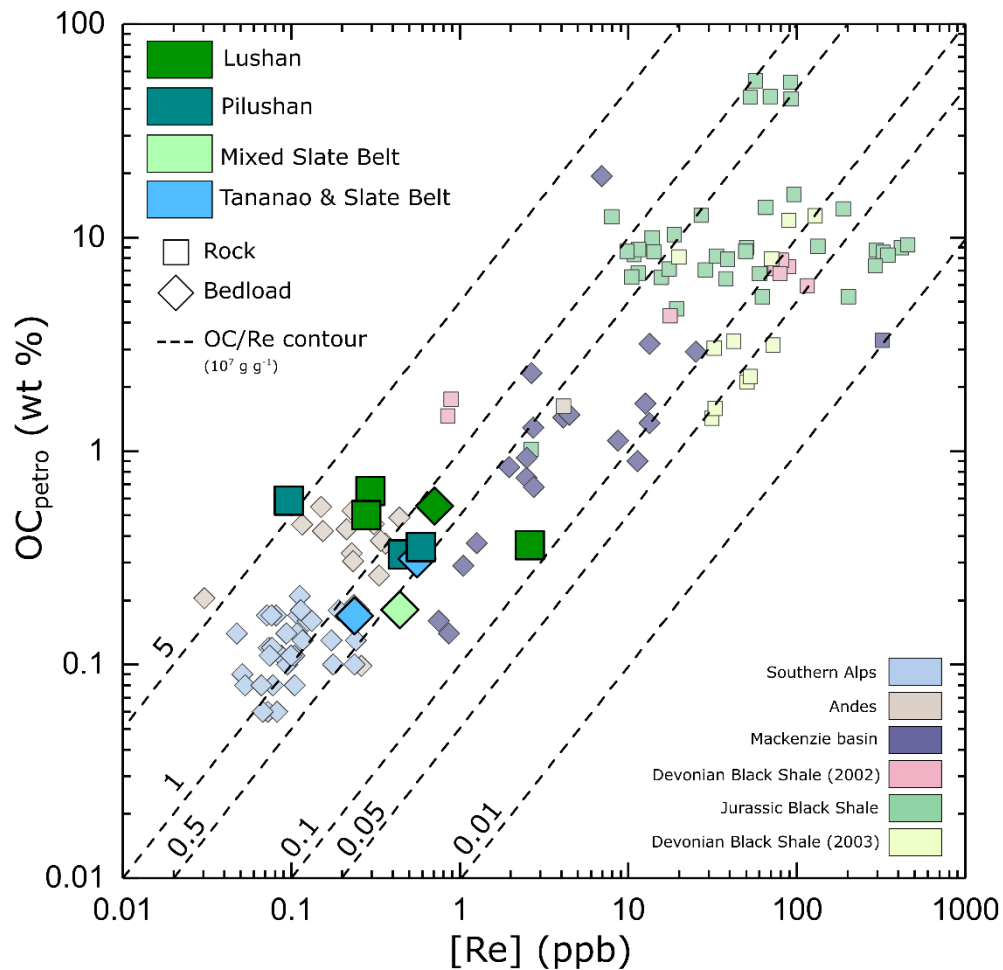
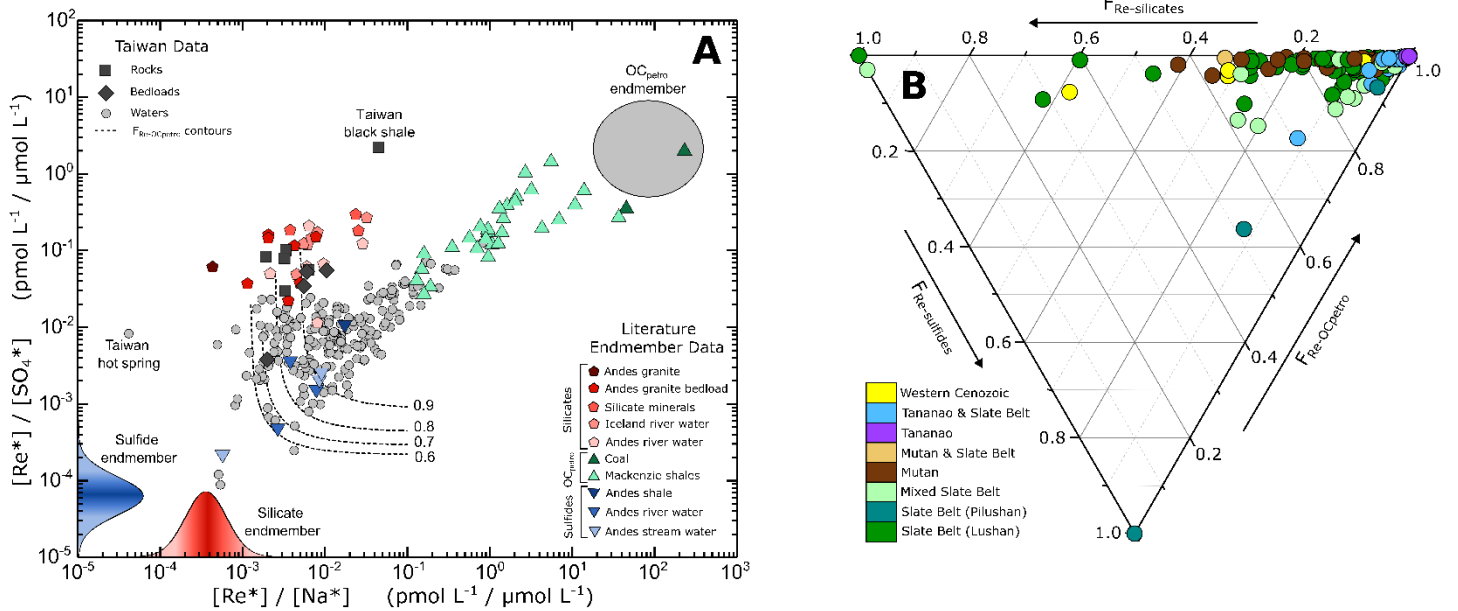
Figure 2 – Solid OC_{petro} and Re composition

Fig. 2. Rhenium and OC_{petro} composition of solids. Taiwan samples are in large symbols coloured by lithology, with literature data in small symbols. Rock samples are represented by squares and bedload samples by rhombuses. Literature data include bedloads from the Southern Alps (Horan et al., 2017), Andes (Dellinger et al., 2023), and Mackenzie basin (Horan et al., 2019), and bedrock from the Andes, Mackenzie basin, Devonian black shales (Jaffe et al., 2002; Selby and Creaser, 2003), and a Jurassic black shale (Cohen et al., 1999). Dashed lines represent contours of $[OC/Re]_{\text{solids}}$ ratios in units of 10^7 g g^{-1} .

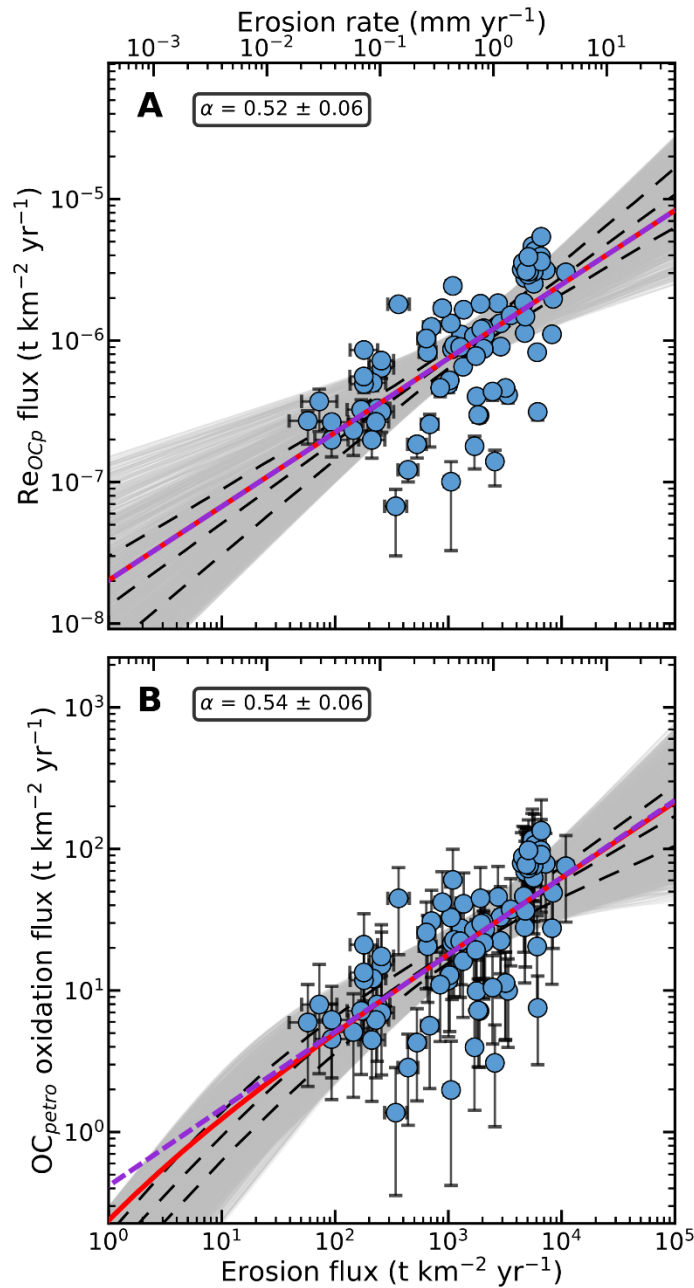
573 **Figure 3 – Source partition diagram and results**

574



575

576 **Fig. 3. A.** Rhenium source partition. $[Re^*]/[Na^*]$ and $[Re^*]/[SO_4^*]$ ratios for water samples (grey
 577 circles), and $[Re/Na]$ and $[Re/S]$ ratios for bedrocks (dark grey squares) and bedloads (dark grey
 578 rhombuses) in this study. Literature data is included to indicate approximate endmember
 579 positions including; for OC_{petro} , Mackenzie shales (Ross and Bustin, 2009) and coal (Dai et al.,
 580 2015); for sulfides, Andes sulfide-rich shales and waters from sulfide-rich catchments
 581 (Dellinger et al., 2023); for silicates, Andes granites and bedloads and waters from granitic
 582 catchments (Dellinger et al., 2023), Icelandic river waters (Pogge von Strandmann et al., 2008),
 583 and silicate minerals (Burton et al., 2002, 1999). Probability distributions indicate the
 584 distributions from which endmember values of $[Re/Na]_{silicates}$ and $[Re/SO_4]_{sulfides}$ were sampled in
 585 the Monte Carlo simulation. Dashed lines indicate contours for proportion of Re^* derived from
 586 OC_{petro} ($F_{Re-OCpetro}$). **B.** Ternary diagram showing the results of the Re source partition ($F_{Re-OCpetro}$,
 587 $F_{Re-silicates}$, and $F_{Re-sulfides}$) coloured by lithology.

588 **Figure 4 – Erosion sensitivity of $Re_{OC_{petro}}$ and OC_{petro} oxidation fluxes**

589 **Fig. 4. A.** $[Re]_{OC_{petro}}$ and **B.** OC_{petro} oxidation flux data (blue circles), power law fit (dashed purple
 590 line), and best fit of the multiparameter model (red line). Uncertainties in the data points
 591 correspond to the 16th and 84th percentiles. Accepted fits (grey lines) are equivalent to a 95%
 592 confidence interval, and the quartiles (dashed black lines) correspond to the 25th, 50th, and 75th
 593 percentiles. The y axes each span four orders of magnitude. Erosion flux ($t\ km^{-2}\ yr^{-1}$) is equal to
 594 $2500 \times$ Erosion rate ($mm\ yr^{-1}$) assuming a crustal density of $2500\ kg\ m^{-3}$. The average erosion
 595 sensitivity (α) is displayed along with the standard error.

Figure 5 – Erosion sensitivity of the net weathering carbon budget

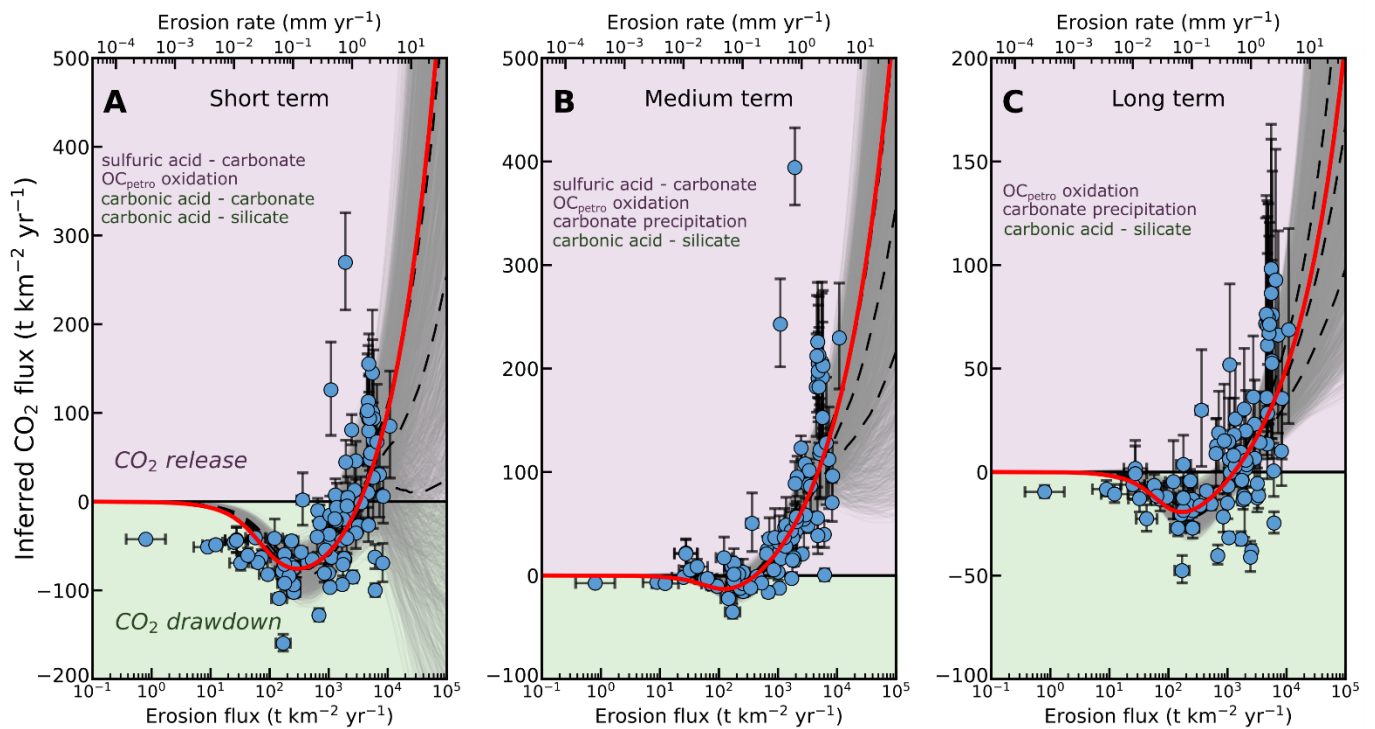
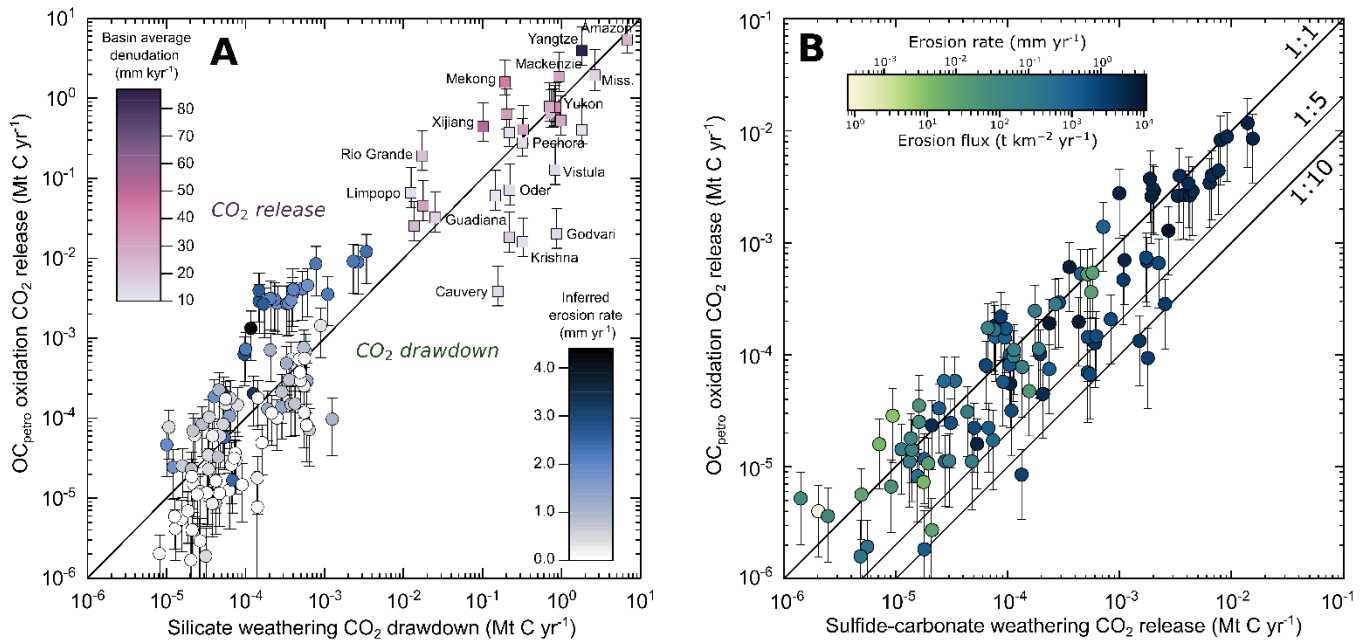


Fig. 5. Inferred net CO₂ fluxes on **A.** short timescales (< 10 kyr), **B.** medium timescales (10 kyr – 10 Myr), and **C.** long timescales (> 10 Myr), with the relevant reactions on each timescale labelled. Data points and linework follow Fig. 4.

Figure 6 – OC_{petro} oxidation -silicate weathering and OC_{petro} oxidation – sulfide-carbonate weathering CO₂ balances



605

606 **Fig. 6. A.** CO₂ release from OC_{petro} oxidation versus (long-term) CO₂ drawdown via silicate

607 weathering. Taiwan data is shown in the bottom left, coloured by inferred erosion rate (mm yr⁻¹),

608 with global data (Zondervan et al., 2023) shown in the top right, coloured by basin average

609 denudation (mm kyr⁻¹). The black line represents a 1:1 ratio. **B.** CO₂ release from OC_{petro}

610 oxidation versus (medium-term) CO₂ release from sulfide-carbonate weathering. Taiwan data is

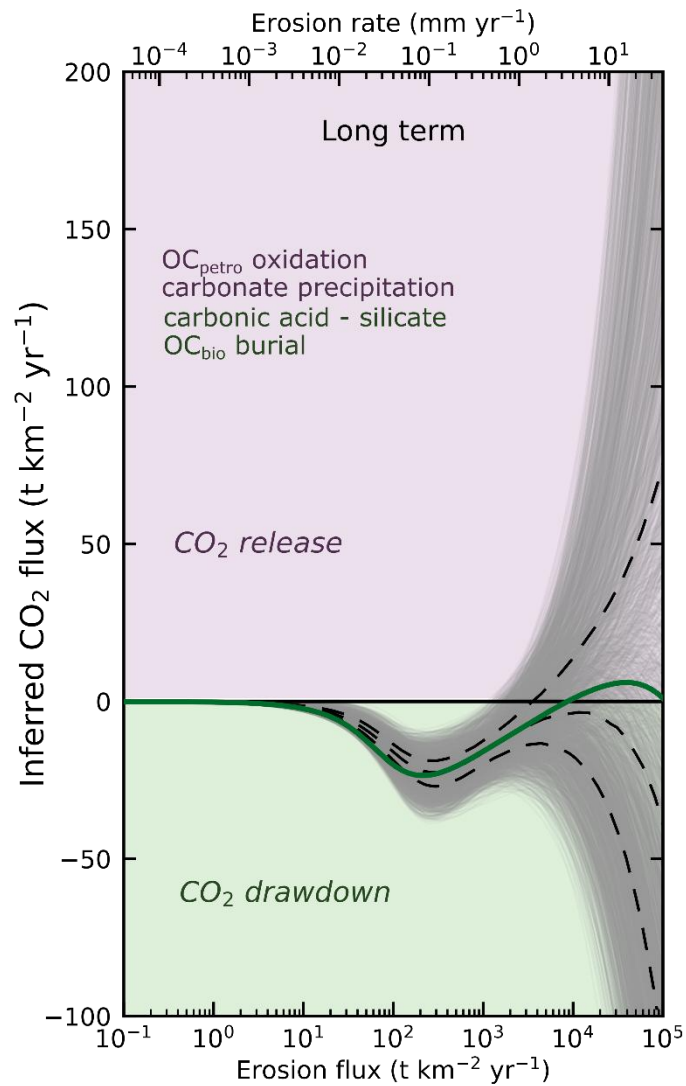
611 shown, coloured by inferred erosion rate (mm yr⁻¹). The black lines represent contours for 1:1,

612 1:5, and 1:10 ratios of OC_{petro} CO₂ release to sulfide-carbonate CO₂ release. In both plots error

613 bars represent the 16th and 84th percentiles, or a confidence interval of 68%. Note that the y axes

614 scales differ between A and B.

615 *Figure 7 – (+OC_{bio}) Erosion sensitivity of the geological carbon cycle*



616 **Fig. 7.** Long term inferred net CO₂ flux including CO₂ sequestration via OC_{bio} burial. Result from
617 combination of the best fit multiparameter model and 'best' fit OC_{bio} power law are shown in
618 green. Accepted fits are shown in grey with 25th, 50th, and 75th quartiles in dashed black lines.

619 CRediT authorship contribution statement

620 Conceptualisation: VA, AB, RH

621 Validation: VA, EW, RH

622 Formal analysis: VA

623 Investigation: VA, EW, AB, RH

624 Resources: AB, NH, RH

625 Data Curation: VA

626 Writing - Original Draft: VA

627 Writing - Review & Editing: VA, AB, EW, LL, NH, RH

628 Visualization: VA

629 Supervision: AB, RH

630 Project administration: VA, RH

631 Funding acquisition: VA, AB, NH, RH

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References

- Beyssac, O., Simoes, M., Avouac, J.P., Farley, K.A., Chen, Y.G., Chan, Y.C., Goffé, B., 2007. Late Cenozoic metamorphic evolution and exhumation of Taiwan. *Tectonics* 26. <https://doi.org/10.1029/2006TC002064>
- Bolton, E.W., Berner, R.A., Petsch, S.T., 2006. The Weathering of Sedimentary Organic Matter as a Control on Atmospheric O₂: II. Theoretical Modeling. *Am. J. Sci.* 306. <https://doi.org/10.2475/08.2006.01>
- Bouchez, J., Beyssac, O., Galy, V., Gaillardet, J., France-Lanord, C., Maurice, L., Moreira-Turcq, P., 2010. Oxidation of petrogenic organic carbon in the Amazon floodplain as a source of atmospheric CO₂. *Geology* 38, 255–258. <https://doi.org/10.1130/G30608>
- Brantley, S.L., Holleran, M.E., Jin, L., Bazilevskaya, E., 2013. Probing deep weathering in the Shale Hills Critical Zone Observatory, Pennsylvania (USA): the hypothesis of nested chemical reaction fronts in the subsurface. *Earth Surf. Process. Landf.* 38, 1280–1298. <https://doi.org/10.1002/esp.3415>
- Bufe, A., Hovius, N., Emberson, R., Rugenstein, J.K.C., Galy, A., Hassenruck-Gudipati, H.J., Chang, J.H., 2021. Co-variation of silicate, carbonate and sulfide weathering drives CO₂ release with erosion. *Nat. Geosci.* 14, 211–216. <https://doi.org/10.1038/s41561-021-00714-3>
- Bufe, A., Rugenstein, J.K.C., Hovius, N., 2024. CO₂ drawdown from weathering is maximized at moderate erosion rates. *Science* 383, 1075–1080. <https://doi.org/10.1126/SCIENCE.ADK0957>
- Burton, K.W., Gannoun, A., Birck, J.-L., Allègre, C.J., Schiano, P., Clocchiatti, R., Alard, O., 2002. The compatibility of rhenium and osmium in natural olivine and their behaviour during

- 663 mantle melting and basalt genesis. *Earth Planet. Sci. Lett.* 198, 63–76.
664 [https://doi.org/10.1016/S0012-821X\(02\)00518-6](https://doi.org/10.1016/S0012-821X(02)00518-6)
- 665 Burton, K.W., Schiano, P., Birck, J.-L., Allègre, C.J., 1999. Osmium isotope disequilibrium
666 between mantle minerals in a spinel-lherzolite. *Earth Planet. Sci. Lett.* 172, 311–322.
667 [https://doi.org/10.1016/S0012-821X\(99\)00207-1](https://doi.org/10.1016/S0012-821X(99)00207-1)
- 668 Canfield, D.E., 2004. The evolution of the Earth surface sulfur reservoir. *Am. J. Sci.* 304.
669 <https://doi.org/https://doi.org/10.2475/ajs.304.10.839>
- 670 Chang, S., Berner, R.A., 1999. Coal weathering and the geochemical carbon cycle. *Geochim.*
671 *Cosmochim. Acta* 63, 3301–3310. [https://doi.org/10.1016/S0016-7037\(99\)00252-5](https://doi.org/10.1016/S0016-7037(99)00252-5)
- 672 Chen, C.-Y., Willett, S.D., West, A.J., Dadson, S., Hovius, N., Christl, M., Shyu, J.B.H., 2020. The
673 impact of storm-triggered landslides on sediment dynamics and catchment-wide
674 denudation rates in the southern Central Range of Taiwan following the extreme rainfall
675 event of Typhoon Morakot. *Earth Surf. Process. Landf.* 45, 548–564.
676 <https://doi.org/10.1002/esp.4753>
- 677 Cohen, A.S., Coe, A.L., Bartlett, J.M., Hawkesworth, C.J., 1999. Precise Re–Os ages of organic-
678 rich mudrocks and the Os isotope composition of Jurassic seawater. *Earth Planet. Sci.*
679 *Lett.* 167, 159–173. [https://doi.org/10.1016/S0012-821X\(99\)00026-6](https://doi.org/10.1016/S0012-821X(99)00026-6)
- 680 Dadson, S.J., Hovius, N., Chen, H., Dade, W.B., Hsieh, M.-L., Willett, S.D., Hu, J.-C., Horng, M.-
681 J., Chen, M.-C., Stark, C.P., Lague, D., Lin, J.-C., 2003. Links between erosion, runoff
682 variability and seismicity in the Taiwan orogen. *Nature* 426, 648–651.
683 <https://doi.org/https://doi.org/10.1038/nature02150>
- 684 Dai, S., Seredin, V.V., Ward, C.R., Hower, J.C., Xing, Y., Zhang, W., Song, W., Wang, P., 2015.
685 Enrichment of U–Se–Mo–Re–V in coals preserved within marine carbonate
686 successions: geochemical and mineralogical data from the Late Permian Guiding

- 687 Coalfield, Guizhou, China. *Miner. Deposita* 50, 159–186.
688 <https://doi.org/10.1007/s00126-014-0528-1>
- 689 Dalai, T.K., Singh, S.K., Trivedi, J.R., Krishnaswami, S., 2002. Dissolved rhenium in the Yamuna
690 River System and the Ganga in the Himalaya: Role of black shale weathering on the
691 budgets of Re, Os, and U in rivers and CO₂ in the atmosphere. *Geochim. Cosmochim.*
692 *Acta* 66, 29–43. [https://doi.org/10.1016/S0016-7037\(01\)00747-5](https://doi.org/10.1016/S0016-7037(01)00747-5)
- 693 Dellinger, M., Hilton, R.G., Baronas, J.J., Torres, M.A., Burt, E.I., Clark, K.E., Galy, V., Ccahuana
694 Quispe, A.J., West, A.J., 2023. High rates of rock organic carbon oxidation sustained as
695 Andean sediment transits the Amazon foreland-floodplain. *Proc. Natl. Acad. Sci.* 120.
696 <https://doi.org/10.1073/PNAS.2306343120>
- 697 Dellinger, M., Hilton, R.G., Nowell, G.M., 2020. Measurements of rhenium isotopic composition
698 in low-abundance samples. *J. Anal. At. Spectrom.* 35, 377–387.
699 <https://doi.org/10.1039/c9ja00288j>
- 700 Egholm, D.L., Knudsen, M.F., Sandiford, M., 2013. Lifespan of mountain ranges scaled by
701 feedbacks between landsliding and erosion by rivers. *Nature* 498, 475–478.
702 <https://doi.org/10.1038/nature12218>
- 703 Fellin, M.G., Chen, C.Y., Willett, S.D., Christl, M., Chen, Y.G., 2017. Erosion rates across space
704 and timescales from a multi-proxy study of rivers of eastern Taiwan. *Glob. Planet.*
705 *Change* 157, 174–193. <https://doi.org/10.1016/J.GLOPLACHA.2017.07.012>
- 706 Gaillardet, J., Dupre, B., Louvat, P., Allegre, C.J., 1999. Global silicate weathering and CO
707 consumption rates deduced from the chemistry of large rivers. *Chem. Geol.* 159, 3–30.
708 [https://doi.org/10.1016/S0009-2541\(99\)00031-5](https://doi.org/10.1016/S0009-2541(99)00031-5)

709 Galy, V., Beyssac, O., France-Lanord, C., Eglinton, T., 2008a. Recycling of graphite during
710 Himalayan erosion: A geological stabilization of carbon in the crust. *Science* 322, 943–
711 945. <https://doi.org/10.1126/science.1161408>

712 Galy, V., France-Lanord, C., Lartiges, B., 2008b. Loading and fate of particulate organic carbon
713 from the Himalaya to the Ganga-Brahmaputra delta. *Geochim. Cosmochim. Acta* 72,
714 1767–1787. <https://doi.org/DOI:%2010.1016/j.gca.2008.01.027>

715 Galy, V., Peucker-Ehrenbrink, B., Eglinton, T., 2015. Global carbon export from the terrestrial
716 biosphere controlled by erosion. *Nature* 521, 204–207.
717 <https://doi.org/10.1038/nature14400>

718 Ghazi, L., Goñi, M., Haley, B.A., Muratli, J.M., Pett-Ridge, J.C., 2022. Concentration-runoff
719 relationships of contrasting small mountainous rivers in the Pacific Northwest, USA:
720 Insights into the weathering of rhenium relative to other weathering products. *Geochim.*
721 *Cosmochim. Acta*. <https://doi.org/10.1016/J.GCA.2022.09.036>

722 Ghazi, L., Grant, K.E., Chappaz, A., Danish, M., Peucker-Ehrenbrink, B., Pett-Ridge, J.C., 2024.
723 The Global Biogeochemical Cycle of Rhenium. *Glob. Biogeochem. Cycles* 38,
724 e2024GB008254. <https://doi.org/10.1029/2024GB008254>

725 Grant, K.E., Dellinger, M., Dickson, A.J., Ogric, M., Horan, K., Petsch, S., Hilton, R.G., 2025.
726 Validating the rhenium proxy for rock organic carbon oxidation using weathering
727 profiles. *Chem. Geol.* 671, 122464. <https://doi.org/10.1016/j.chemgeo.2024.122464>

728 Hemingway, J.D., Hilton, R.G., Hovius, N., Eglinton, T.I., Haghipour, N., Wacker, L., Chen, M.-C.,
729 Galy, V.V., 2018. Microbial oxidation of lithospheric organic carbon in rapidly eroding
730 tropical mountain soils. *Science* 360, 209–212.
731 <https://doi.org/10.1126/science.aao6463>

- 732 Hilton, R.G., Gaillardet, J., Calmels, D., Birck, J.-L., 2014. Geological respiration of a mountain
733 belt revealed by the trace element rhenium. *Earth Planet. Sci. Lett.* 403, 27–36.
734 <https://doi.org/10.1016/j.epsl.2014.06.021>
- 735 Hilton, R.G., Galy, A., Hovius, N., Horng, M.-J., Chen, H., 2010. The isotopic composition of
736 particulate organic carbon in mountain rivers of Taiwan. *Geochim. Cosmochim. Acta*
737 74, 3164–3181. <https://doi.org/10.1016/j.gca.2010.03.004>
- 738 Hilton, R.G., Schwab, M.S., Galy, V., 2025. Dynamics of particulate organic carbon mobilization,
739 storage, and export across river sedimentary systems, in: *Treatise on Geochemistry*.
740 Elsevier, pp. 215–250. [https://doi.org/https://doi.org/10.1016/B978-0-323-99762-](https://doi.org/10.1016/B978-0-323-99762-1.00021-8)
741 [1.00021-8](https://doi.org/10.1016/B978-0-323-99762-1.00021-8)
- 742 Hilton, R.G., Turowski, J.M., Winnick, M., Dellinger, M., Schleppe, P., Williams, K.H., Lawrence,
743 C.R., Maher, K., West, M., Hayton, A., 2021. Concentration-Discharge Relationships of
744 Dissolved Rhenium in Alpine Catchments Reveal Its Use as a Tracer of Oxidative
745 Weathering. *Water Resour. Res.* 57. <https://doi.org/10.1029/2021WR029844>
- 746 Hilton, R.G., West, A.J., 2020. Mountains, erosion and the carbon cycle. *Nat. Rev. Earth Environ.*
747 1, 284–299. <https://doi.org/10.1038/s43017-020-0058-6>
- 748 Horan, K., Hilton, R.G., Dellinger, M., Tipper, E., Galy, V., Calmels, D., Selby, D., Gaillardet, J.,
749 Ottley, C.J., Parsons, D.R., Burton, K.W., 2019. Carbon dioxide emissions by rock
750 organic carbon oxidation and the net geochemical carbon budget of the Mackenzie
751 River Basin. *Am. J. Sci.* 319, 473–499. <https://doi.org/10.2475/06.2019.02>
- 752 Horan, K., Hilton, R.G., Selby, D., Ottley, C.J., Gröcke, D.R., Hicks, M., Burton, K.W., 2017.
753 Mountain glaciation drives rapid oxidation of rock-bound organic carbon. *Sci. Adv.* 3.
754 <https://doi.org/10.1126/sciadv.1701107>

- 755 Jaffe, L.A., Peucker-Ehrenbrink, B., Petsch, S.T., 2002. Mobility of rhenium, platinum group
756 elements and organic carbon during black shale weathering. *Earth Planet. Sci. Lett.*
757 198, 339–353. [https://doi.org/10.1016/S0012-821X\(02\)00526-5](https://doi.org/10.1016/S0012-821X(02)00526-5)
- 758 Jochum, K.P., Weis, U., Stoll, B., Wilson, S.A., Haug, G.H., Andreae, M.O., Enzweiler, J., 2016.
759 Reference Values Following ISO Guidelines for Frequently Requested Rock Reference
760 Materials. *Geoanalysts* 1751908x 3. <https://doi.org/10.1111/j.1751-908X.2015.00392.x>
- 761 Kemeny, P.C., Lopez, G.I., Dalleska, N.F., Torres, M., Burke, A., Bhatt, M.P., West, A.J.,
762 Hartmann, J., Adkins, J.F., 2021. Sulfate sulfur isotopes and major ion chemistry reveal
763 that pyrite oxidation counteracts CO₂ drawdown from silicate weathering in the
764 Langtang-Trisuli-Narayani River system, Nepal Himalaya. *Geochim. Cosmochim. Acta*
765 294, 43–69. <https://doi.org/10.1016/j.gca.2020.11.009>
- 766 Lasaga, A.C., Soler, J.M., Ganor, J., Burch, T.E., Nagy, K.L., 1994. Chemical weathering rate laws
767 and global geochemical cycles. *Geochim. Cosmochim. Acta* 58, 2361–2386.
768 [https://doi.org/10.1016/0016-7037\(94\)90016-7](https://doi.org/10.1016/0016-7037(94)90016-7)
- 769 Lee, C.-S., Huang, L.-R., Shen, H.-S., Wang, S.-T., 2006. A Climatology Model for Forecasting
770 Typhoon Rainfall in Taiwan. *Nat. Hazards* 37, 87–105. [https://doi.org/10.1007/s11069-](https://doi.org/10.1007/s11069-005-4658-8)
771 005-4658-8
- 772 Lien, W.-Y., Chen, C.-T., Lee, Y.-H., Su, C.-C., Wang, P.-L., Lin, L.-H., 2025a. Two-stage oxidation
773 of petrogenic organic carbon in a rapidly exhuming small mountainous catchment.
774 *Commun. Earth Environ.* 6, 45. <https://doi.org/10.1038/s43247-025-02015-8>
- 775 Lien, W.-Y., Chen, R., Tsai, J.-F., Chen, J.-N., Lin, L.-H., Wang, P.-L., 2025b. A novel design for
776 continuous measurement of CO₂ from deep remineralization of petrogenic organic
777 carbon. *Vadose Zone J.* 24, e70020. <https://doi.org/10.1002/vzj2.70020>
- 778 Lin, C.W., Chen, W.S., 2016. Geologic map of Taiwan.

- 779 Liu, H.-C., Li, M.-X., You, C.-F., Chen, Y.-H., Huang, K.-F., Chung, C.-H., Lin, G.-W., 2024.
780 Hydrology controls sulfuric acid-mediated weathering in an orogenic regime of
781 southwestern Taiwan. *Sci. Total Environ.* 951, 175630.
782 <https://doi.org/10.1016/j.scitotenv.2024.175630>
- 783 Lupker, M., France-Lanord, C., Galy, V., Lavé, J., Gaillardet, J., Gajurel, A.P., Guilmette, C.,
784 Rahman, M., Singh, S.K., Sinha, R., 2012. Predominant floodplain over mountain
785 weathering of Himalayan sediments (Ganga basin). *Geochim. Cosmochim. Acta* 84,
786 410–432. <https://doi.org/10.1016/j.gca.2012.02.001>
- 787 Macdonald, F.A., Swanson-Hysell, N.L., Park, Y., Lisiecki, L., Jagoutz, O., 2019. Arc-continent
788 collisions in the tropics set Earth's climate state. *Science* 364, 181–184.
789 <https://doi.org/10.1126/science.aav5300>
- 790 Maher, K., Chamberlain, C.P., 2014. Hydrologic Regulation of Chemical Weathering and the
791 Geologic Carbon Cycle. *Science* 343, 1502–1504.
792 <https://doi.org/10.1126/science.1250770>
- 793 Märki, L., Lupker, M., France-Lanord, C., Lavé, J., Gallen, S., Gajurel, A.P., Haghipour, N.,
794 Leuenberger-West, F., Eglinton, T., 2021. An unshakable carbon budget for the
795 Himalaya. *Nat. Geosci.* 14, 745–750. <https://doi.org/10.1038/s41561-021-00815-z>
- 796 Moon, S., Chamberlain, C.P., Hilley, G.E., 2014. New estimates of silicate weathering rates and
797 their uncertainties in global rivers. *Geochim. Cosmochim. Acta* 134, 257–274.
798 <https://doi.org/10.1016/J.GCA.2014.02.033>
- 799 Ogrič, M., Dellinger, M., Grant, K.E., Galy, V., Gu, X., Brantley, S.L., Hilton, R.G., 2023. Low rates
800 of rock organic carbon oxidation and anthropogenic cycling of rhenium in a slowly
801 denuding landscape. *Earth Surf. Process. Landf.* 48, 1202–1218.
802 <https://doi.org/10.1002/esp.5543>

- 803 Petsch, S.T., 2014. Weathering of organic carbon, in: *Treatise on Geochemistry*. pp. 217–238.
- 804 Petsch, S.T., Berner, R.A., Eglinton, T.I., 2000. A field study of the chemical weathering of
805 ancient sedimentary organic matter. *Org. Geochem.* 31, 475–487.
806 [https://doi.org/10.1016/S0146-6380\(00\)00014-0](https://doi.org/10.1016/S0146-6380(00)00014-0)
- 807 Plank, T., Manning, C.E., 2019. Subducting carbon. *Nature* 574, 343–352.
808 <https://doi.org/10.1038/s41586-019-1643-z>
- 809 Pogge von Strandmann, P.A.E., Burton, K.W., James, R.H., van Calsteren, P., Gislason, S.R.,
810 Sigfússon, B., 2008. The influence of weathering processes on riverine magnesium
811 isotopes in a basaltic terrain. *Earth Planet. Sci. Lett.* 276, 187–197.
812 <https://doi.org/10.1016/j.epsl.2008.09.020>
- 813 Raymo, M.E., Ruddiman, W.F., 1992. Tectonic forcing of late Cenozoic climate. *Nature* 359,
814 117–122. <https://doi.org/10.1038/359117a0>
- 815 Ross, D.J.K., Bustin, R.M., 2009. Investigating the use of sedimentary geochemical proxies for
816 paleoenvironment interpretation of thermally mature organic-rich strata: Examples
817 from the Devonian–Mississippian shales, Western Canadian Sedimentary Basin. *Chem.*
818 *Geol.* 260, 1–19. <https://doi.org/10.1016/j.chemgeo.2008.10.027>
- 819 Roylands, T., Hilton, R.G., Garnett, M.H., Soulet, G., Newton, J.-A., Peterkin, J.L., Hancock, P.,
820 2022. Capturing the short-term variability of carbon dioxide emissions from
821 sedimentary rock weathering in a remote mountainous catchment, New Zealand.
822 *Chem. Geol.* 608, 121024. <https://doi.org/10.1016/j.chemgeo.2022.121024>
- 823 Selby, D., Creaser, R.A., 2003. Re-Os geochronology of organic rich sediments: an evaluation of
824 organic matter analysis methods. *Chem. Geol.* 200, 225–240.
825 [https://doi.org/10.1016/S0009-2541\(03\)00199-2](https://doi.org/10.1016/S0009-2541(03)00199-2)

- 826 Soulet, G., Hilton, R.G., Garnett, M.H., Roylands, T., Klotz, S., Croissant, T., Dellinger, M., Le
827 Bouteiller, C., 2021. Temperature control on CO₂ emissions from the weathering of
828 sedimentary rocks. *Nat. Geosci.* 14, 665–671. [https://doi.org/10.1038/s41561-021-](https://doi.org/10.1038/s41561-021-00805-1)
829 00805-1
- 830 Sparkes, R.B., Hovius, N., Galy, A., Liu, J.T., 2020. Survival of graphitized petrogenic organic
831 carbon through multiple erosional cycles. *Earth Planet. Sci. Lett.* 531, 115992.
832 <https://doi.org/10.1016/j.epsl.2019.115992>
- 833 Stow, M., Dickson, A.J., Prytulak, J., Dellinger, M., Alcock, V., Hesselbo, S.P., Nowell, G.M.,
834 Hilton, R.G., 2026. Rhenium isotopes reveal enhanced rock organic carbon oxidation
835 over the Toarcian Oceanic Anoxic Event. *Nat. Commun.*
836 <https://doi.org/10.1038/s41467-026-71533-6>
- 837 Teng, L.S., 1990. Geotectonic evolution of late Cenozoic arc-continent collision in Taiwan.
838 *Tectonophysics* 183, 57–76. [https://doi.org/10.1016/0040-1951\(90\)90188-E](https://doi.org/10.1016/0040-1951(90)90188-E)
- 839 Torres, M.A., West, A.J., Clark, K.E., Paris, G., Bouchez, J., Ponton, C., Feakins, S.J., Galy, V.,
840 Adkins, J.F., 2016. The acid and alkalinity budgets of weathering in the Andes–Amazon
841 system: Insights into the erosional control of global biogeochemical cycles. *Earth*
842 *Planet. Sci. Lett.* 450, 381–391. <https://doi.org/10.1016/j.epsl.2016.06.012>
- 843 Urey, H.C., 1952. *The Planets: Their Origin and Development*. Oxford University Press.
- 844 Wang, P.-L., Tu, T.-H., Lin, L.-H., Chou, H.-L., Wang, Y.-J., Chen, J.-N., Wang, L.-Y., Chang, J.-M.,
845 Chu, M.-F., Hsu, Y.-C., Chang, C.-P., Wu, Y.-M., Lin, Y.-T., Ke, C.-C., 2024. Microbial
846 communities modulate chemical weathering and carbon dioxide cycling in an active
847 orogen in Taiwan. *Commun. Earth Environ.* 5, 174. [https://doi.org/10.1038/s43247-024-](https://doi.org/10.1038/s43247-024-01345-3)
848 01345-3

- 849 Wang, Y., Chen, Y., Li, S., Hedding, D.W., Chen, J., 2025. Carbon Emissions Through the
850 Weathering of Petrogenic Organic Carbon and the Net Geological Carbon Budget of the
851 Tibetan Plateau. *Geophys. Res. Lett.* 52, e2024GL110718.
852 <https://doi.org/10.1029/2024GL110718>
- 853 West, A.J., 2012. Thickness of the chemical weathering zone and implications for erosional and
854 climatic drivers of weathering and for carbon-cycle feedbacks. *Geology* 40, 811–814.
855 <https://doi.org/10.1130/G33041.1>
- 856 West, A.J., Galy, A., Bickle, M., Farley, K., 2005. Tectonic and climatic controls on silicate
857 weathering. *Earth Planet. Sci. Lett.* 235, 211–228.
858 <https://doi.org/10.1016/j.epsl.2005.03.020>
- 859 White, A.F., Buss, H.L., 2013. Natural weathering rates of silicate minerals, in: *Treatise on*
860 *Geochemistry*. Elsevier, pp. 115–155. [https://doi.org/https://doi.org/10.1016/B978-0-08-](https://doi.org/https://doi.org/10.1016/B978-0-08-095975-7.00504-0)
861 [095975-7.00504-0](https://doi.org/https://doi.org/10.1016/B978-0-08-095975-7.00504-0)
- 862 Williamson, M.A., Rimstidt, J.D., 1994. The kinetics and electrochemical rate-determining step
863 of aqueous pyrite oxidation. *Geochim. Cosmochim. Acta* 58, 5443–5454.
864 [https://doi.org/10.1016/0016-7037\(94\)90241-0](https://doi.org/10.1016/0016-7037(94)90241-0)
- 865 Wu, C.C., Kuo, Y.H., 1999. Typhoons Affecting Taiwan: Current Understanding and Future
866 Challenges. *Bull. Am. Meteorol. Soc.* 80. [https://doi.org/https://doi.org/10.1175/1520-](https://doi.org/https://doi.org/10.1175/1520-0477(1999)080<0067:TATCUA>2.0.CO;2)
867 [0477\(1999\)080<0067:TATCUA>2.0.CO;2](https://doi.org/https://doi.org/10.1175/1520-0477(1999)080<0067:TATCUA>2.0.CO;2)
- 868 Yeghicheyan, D., Aubert, D., Bouhnik-Le Coz, M., Chmeleff, J., Delpoux, S., Djouraev, I., Granier,
869 G., Lacan, F., Piro, J.-L., Rousseau, T., Cloquet, C., Marquet, A., Menniti, C., Pradoux, C.,
870 Freydier, R., da Silva-Filho, E.V., Suchorski, K., 2019. A New Interlaboratory
871 Characterisation of Silicon, Rare Earth Elements and Twenty-Two Other Trace Element

872 Concentrations in the Natural River Water Certified Reference Material SLRS-6 (NRC-
873 CNRC). *Geostand. Geoanalytical Res.* 43, 475–496. <https://doi.org/10.1111/ggr.12268>

874 Yeghicheyan, D., Grinberg, P., Alleman, L.Y., Belhadj, M., Causse, L., Chmeleff, J., Cordier, L.,
875 Djoureaev, I., Dumoulin, D., Dumont, J., Freydier, R., Mariot, H., Cloquet, C., Kumkrong,
876 P., Malet, B., Jeandel, C., Marquet, A., Riotte, J., Tharaud, M., Billon, G., Trommether, G.,
877 Séby, F., Guihou, A., Deschamps, P., Mester, Z., 2021. Collaborative determination of
878 trace element mass fractions and isotope ratios in AQUA-1 drinking water certified
879 reference material. *Anal. Bioanal. Chem.* 413, 4959–4978.
880 <https://doi.org/10.1007/s00216-021-03456-8>

881 Zeebe, R.E., Westbroek, P., 2003. A simple model for the CaCO₃ saturation state of the ocean:
882 The “Strangelove,” the “Neritan,” and the “Cretan” Ocean. *Geochem. Geophys.*
883 *Geosystems* 4. <https://doi.org/10.1029/2003GC000538>

884 Zondervan, J.R., Hilton, R.G., Dellinger, M., Clubb, F.J., Roylands, T., Ogrič, M., 2023. Rock
885 organic carbon oxidation CO₂ release offsets silicate weathering sink. *Nature* 623, 329.
886 <https://doi.org/10.1038/s41586-023-06581-9>

887

888

889 Supplementary Text

890 *S. 1 Solid sample digestion and rhenium purification*

891 The rhenium concentrations of solid samples were determined broadly following established
892 dissolution, purification, and measurement protocols (Dellinger et al., 2020). Half a gram of
893 homogenized powdered bedrock and bedload samples were digested and purified via anion
894 exchange chromatography. Samples were treated in clean PFA beakers in a mixture of 3 mL 25
895 M HF and 3 mL 16 M HNO₃ for at least 48 hours at 120 °C, dried down, and treated repeatedly
896 with a mixture of 6 mL 16 M HNO₃ and 3 mL 10M HCl for at least 48 hours at 160 °C to attack the
897 organic component. The resulting residue was dissolved in ~ 20 mL 1 M HCl and loaded onto
898 cleaned and conditioned Bio-Rad Poly-Prep® columns with 1.1 mL AG® 1-X8 resin (200-400
899 mesh, chloride form). AG® 1-X8 resin was pre-cleaned in several cycles of rinses in 18.2 MΩ cm⁻¹
900 water, 8 M HNO₃, and 5 M HCl before loading to the column. Columns were cleaned,
901 conditioned, loaded, and eluted according to the following procedure:

902 Pre-cleaning and resin load:

- 903 1) + 15 mL * M HNO₃
904 2) + 1.1 mL Bio-Rad AG® 1-X8 resin

905 Resin cleaning:

- 906 3) + 10 mL 10 M HCl
907 4) + 10 mL 1 M HCl
908 5) + 35 mL 8 M HNO₃

909 Conditioning and sample load:

- 910 6) + 5 mL 0.5 M HNO₃
911 7) + 5 mL 1 M HCl
912 8) + ~ 20 mL sample in 1 M HCl

913 Matrix elution:

- 914 9) + 10 mL 1 M HCl
915 10) + 15 mL 0.5 M HNO₃

11) + 1.5 mL 4 M HNO₃

Rhenium elution:

12) + 15 mL 4 M HNO₃

The Re cut was dried at 80 °C and refluxed in 15 M HNO₃ at 160 °C for several days to destroy any resin-derived organics, before drying and retaking in 2% HNO₃ for Q-ICP-MS analysis.

All acids used were purified in-house by sub-boiling distillation in quartz glass stills, and reagent dilutions were made using ultra-pure 18.2 MΩ cm⁻¹ (Milli-Q) water. PFA beakers and PP columns, tubes, and pipette tips were thoroughly cleaned in acid before use.

S. 2 Extended rhenium proxy application

This section provides additional detail for the rhenium source partition and OC_{petro} oxidation proxy described in sections 3.3 and 3.4 of the main text.

S. 2. 1 Rain correction

To correct for cyclic Re, Na, and SO₄ inputs we use (S1) and (S2) to estimate a rain-corrected $[Re^*]$, $[Na^*]$, and $[SO_4^*]$. Due to the absence of evaporite deposits in the study area, $[Cl]_{rain}$ is approximated to equal $[Cl]_{diss}$ (Lin and Chen, 2016; Ho, 1986). We note that previous work on this sample set (Bufe 2021) partitions $[Cl]_{diss}$ between both cyclic and hot spring inputs, however results for $[Cl]_{cyclic}$ from this inversion are only available for 55% of the samples in this study. In Bufe *et al.*, (2021), the median $[Cl]_{hot-spring}$ contribution is 2.9% of $[Cl]_{diss}$, and only 12% of samples have a $[Cl]_{hot-spring}$ greater than 10% of the total $[Cl]_{diss}$. We therefore consider this effect to be minimal, and compute the $[Re^*]$, $[Na^*]$, and $[SO_4^*]$ needed for the geological source partition based on $[Cl]_{diss} = [Cl]_{rain}$. Values of $[Na/Cl]_{rain}$ and $[SO_4/Cl]_{rain}$ were estimated as 0.858 and 0.052 μmol μmol⁻¹ based on the composition of seawater (Millero *et al.*, 2008) following the approach of Bufe *et al.*, (2021).

$$[X]_{rain} = \left[\frac{X}{Cl} \right]_{rain} \times [Cl]_{rain} \quad (S1)$$

$$[X^*] = [X]_{diss} - [X]_{rain} \quad (S2)$$

A rainwater sample was not available from which to determine $[Re/Cl]_{rain}$, however $[Re]_{rain}$ has previously been identified as below the Q-ICP-MS detection limit of 0.05 pmol L⁻¹ (Dellinger et al., 2023). $[Re]_{seawater}$ is approximately 40 pmol L⁻¹ (Dickson et al., 2020; Anbar et al., 1992), implying a $[Re/Cl]_{seawater}$ of approximately 7 x 10⁻⁵ pmol μmol⁻¹, which would result in a very small rain correction. Since no samples were a close match to the major element composition of Taiwan rain (Wai et al., 2008), we filtered the sample set against four criteria to identify candidate samples with minimal deep groundwater influence to represent $[Re/Cl]_{rain}$; low-lying, elevation < 60 m; small catchments, < 3 km²; $[Re]_{diss} < 2$ pmol L⁻¹; and $[Re/Cl]_{diss} < 0.005$ pmol μmol⁻¹. We identified three samples (TW17-16, TW17-124, and TW17-126) which satisfied 3 or more of these criteria and took the mean of their $[Re/Cl]_{diss}$ of 0.00337 pmol μmol⁻¹ as our $[Re/Cl]_{rain}$. This value is of the same order of magnitude as ratios used by previous authors (Dellinger et al., 2023) but is two orders of magnitude greater than the seawater ratio. Our rain correction is therefore relatively large, and our attribution of Re to OC_{petro} is conservative. We are comfortable with a relatively large rain correction due to prior suggestion of anthropogenic Re sources associated with industrial activity and the proximity of the study area to petrochemical industries in west Taiwan and China. The median rain correction was -4.2% of $[Re]_{diss}$, and 75% of samples had a correction less than -15%. Neither $[Re]_{OC_{petro}}$ or normalised steepness index (ksn) are correlated with the percent magnitude of the rain correction.

961 S. 2. 2 Geological source partition

Prior work has identified three potential geological sources of Re; silicates, sulfides, and OC_p. Correlation has been observed between riverine $[Re]_{diss}$ and $[SO_4]_{diss}$, suggesting sulfide minerals in black shales could host and release Re (e.g. Colodner et al., 1993; Dalai et al., 2002; Miller et al., 2011). Previous authors also identify Re in silicate minerals (Dalai et al., 2002; Wang et al.,

2024). We note that carbonates were a significant component of the source partition in Bufe et al., (2021), however carbonates are considered an insignificant source of Re so were not included in the Re partition. Previous work estimates the $[Re/Ca]_{carbonates}$ as 4.35×10^{-5} pmol μmol^{-1} (Dalai et al., 2002). At this ratio and assuming all Ca in the samples is derived from carbonates, the median maximum possible contribution of carbonates to $[Re]_{diss}$ is $<1\%$. Similarly hot spring sample TWS15-22 has a low $[Re]_{diss}$ of 1.94 pmol L^{-1} in the bottom 10% of the sample set, therefore hot springs were excluded as a potential Re source.

Instead, OC_{petro} has consistently been identified as the dominant source of Re in rivers (e.g. Dellinger et al., 2023; Horan et al., 2019; Rout and Tripathy, 2024). This is expected owing to the association of Re and OC_{petro} in rocks (e.g. Ravizza and Turekian, 1989; Selby and Creaser, 2003), weathering profiles (Hilton et al., 2014; Jaffe et al., 2002; Peucker-Ehrenbrink and Hannigan, 2000), and marine sediments (e.g. Ravizza et al., 1991; Ravizza and Esser, 1993). To partition $[Re^*]$ between sulfides, silicates, and OC_{petro} , we therefore identify endmember ratios of $[Re^*/SO_4^*]_{sulfides}$ and $[Re^*/Na^*]_{silicates}$ (section 3.3, main text) and use (S3-5) to subtract their contributions to $[Re^*]$, attributing the remainder to OC_{petro} .

$$[Re]_{sulfides} = \left(\frac{Re^*}{SO_4^*} \right)_{sulfides} \times [SO_4^*] \quad (S3)$$

$$[Re]_{silicates} = \left(\frac{Re^*}{Na^*} \right)_{silicates} \times [Na^*] \quad (S4)$$

$$[Re]_{OC_{petro}} = [Re^*] - [Re]_{sulfides} - [Re]_{silicates} \quad (S5)$$

S. 2. 3 Monte Carlo uncertainty analysis

Acknowledging uncertainty in the choice of endmembers for $[Re^*/SO_4^*]_{sulfides}$ and $[Re^*/Na^*]_{silicates}$, we use a Monte Carlo approach to quantify uncertainty in the source partition (Dellinger et al., 2023). Over 100,000 iterations we sample $[Re^*/SO_4^*]_{sulfides}$ and $[Re^*/Na^*]_{silicates}$ from log-normal distributions spanning an order of magnitude and centered on the sample-based endmember value; for sulfides, 9×10^{-5} (range $3 - 30 \times 10^{-5}$) pmol μmol^{-1} , and for silicates, 5×10^{-4} (range $2 -$

20 x 10⁻⁴) pmol μmol⁻¹. Distribution widths were parameterized such that the range corresponds to the ± 2σ interval, thereby encompassing ~95% of the sampled values while allowing limited exploration of the tails. $[Re]_x$ (S3-5) and $F_{Re\ x}$ (S6) were computed for sulfides, silicates, and OC_{petro}, and the median value taken across all iterations. For iterations where all $[Re]_{diss}$ could be explained by contributions from silicates and sulfides, $[Re]_{OCpetro}$ was set to 0. Uncertainty bounds are reported as the 16th and 84th percentiles of the Monte Carlo results, corresponding to the 68% confidence interval, and acknowledging asymmetries in the results.

$$F_{ReOCpetro} = \frac{[Re]_{OCpetro}}{[Re^*]} \quad (S6)$$

We additionally apply a Monte Carlo uncertainty approach to our conversion of $[Re]_{OCpetro}$ to a OC_{petro} oxidation flux, $J_{OCpetro-ox}$ (section 3.4, main text). We calculate $J_{OCpetro-ox}$ (1) over each sample within the source partition Monte Carlo simulation in order to propagate the simulated $[Re]_{OCpetro}$ and incorporate uncertainties in $[OC/Re]_{solids}$ and $f_{graphite}$. Graphitisation of organic matter has been observed in Taiwan corresponding to degree of metamorphism, with relatively less ordered carbonaceous material present in the Cenozoic Slate Belt, and more ordered material in the Tananao schist (Beyssac et al., 2007). However, the absolute fraction of OC_{petro} present as graphite is unconstrained. We sample $f_{graphite}$ from a normal distribution centered on 0.4, with 95% of samples between 0.2 and 0.6. $[OC/Re]_{solids}$ was estimated based on the measured ratios from bedrock and bedload samples (section 3.2, main text). Excluding an outlier with a high $[OC/Re]_{solids}$ of 6.03 10⁷ g g⁻¹, we sampled $[OC/Re]_{solids}$ from a uniform distribution spanning the range of the samples (0.14 – 2.20 10⁷ g g⁻¹).

S. 3 Controls on $[Re]_{OCpetro}$ (multivariate regression analysis)

Erosion rate was expected to exert a first-order control on $[Re]_{OCpetro}$ and other weathering fluxes across the Taiwan dataset due to the four order-of-magnitude range in erosion rates across the study area, accompanied by little variation in runoff, temperature, and lithology (Bufe et al.,

2021; Dadson *et al.*, 2003; Fellin *et al.*, 2017; Lee *et al.*, 2006). Nonetheless, these other environmental variables are known to influence oxidative weathering processes (e.g. Hilton *et al.*, 2021; Roylands *et al.*, 2022; Soulet *et al.*, 2021) and could covary with erosion. For example, rapidly eroding mountain areas may have high runoff through orographic precipitation, and different geological units within the stratigraphy will broadly be uplifted to different elevations at different erosion rates. Some of the relationship between erosion rate and $[\text{Re}]_{\text{OCpetro}}$ and OC_{petro} oxidation fluxes (Fig. 4) could therefore potentially be explained by these other environmental controls. To evaluate this possibility we applied multivariate statistical approaches to assess the relative importance of erosion rate, mean annual runoff, temperature, pH, dissolved oxygen, and lithology in controlling $[\text{Re}]_{\text{OCpetro}}$.

Multivariate ordinary least squares (OLS) regression was applied to the suite of considered catchment properties. $[\text{Re}]_{\text{OCpetro}}$ and erosion rate were log-transformed to account for their approximate power-law relationship. Lithology was represented as a binary variable describing the presence or absence of the major geological units (Mutan, Lushan, Pilushan, Tananao) within each sample catchment based on surface outcrop (Fig. 1). The model explained 55% of the total variance in log-transformed $[\text{Re}]_{\text{OCpetro}}$, with log-transformed erosion rate exhibiting the largest and most statistically significant coefficient among the continuous variables (Fig. S5). Runoff exhibited a weakly negative coefficient with only borderline significance of ($p = 0.061$), while temperature, dissolved oxygen, and pH were not significant predictors of log-transformed $[\text{Re}]_{\text{OCpetro}}$ (Fig. S5). Of the binary lithological variables, presence of the Pilushan formation was significantly and strongly positively correlated with log-transformed $[\text{Re}]_{\text{OCpetro}}$, and presence of the Mutan formation was strongly negatively correlated, but with borderline significance ($p = 0.53$).

To account for potential covariance between predictor variables we additionally applied a multivariate ridge regression. All coefficients shrank between the OLS and ridge regressions

except log-transformed erosion rate which slightly increased, implying some of the OLS coefficients were inflated due to collinearity with erosion rate. To assess the relative importance of individual predictors we ran a lasso regression, which penalises model complexity and removes weak predictors. Runoff, dissolved oxygen, and presence of the Lushan and Mutan formation were deselected in this approach, and log-transformed erosion rate and presence of the Pilushan formation remained the largest coefficients. Inclusion of interaction terms between erosion rate and lithology did not increase the total variance explained in the models, implying that erosion sensitivity of $[\text{Re}]_{\text{OCpetro}}$ is broadly consistent across lithological domains. Principal component analysis yielded results consistent with the regression analyses, with log-transformed $[\text{Re}]_{\text{OCpetro}}$ and erosion rate aligned along PC1, while temperature, dissolved oxygen, and pH exhibit little covariation with $[\text{Re}]_{\text{OCpetro}}$ (Fig. S6). Runoff covaries with log-transformed erosion rate, but is close to orthogonal to $[\text{Re}]_{\text{OCpetro}}$, implying little covariance (Fig. S6). Lithological differences are expressed primarily along PC2, implying a secondary influence compared to erosion rate (Fig. S6). Collectively these results indicate that erosion rate exerts the dominant control on $[\text{Re}]_{\text{OCpetro}}$, with lithology representing a significant secondary influence. Despite their known influence on oxidative weathering, direct effects of runoff and temperature on $[\text{Re}]_{\text{OCpetro}}$ are either comparatively small or cannot be independently resolved from their covariance with erosion.

S. 4 Extended multiparameter modelling approach

S. 4. 1 Uncertainty quantification

Uncertainties were determined by identifying ‘accepted’ fits in a region of parameter space for which models were statistically indistinguishable from the best fit, given the noise in the data. Parameter space was sampled at random across uniform distributions over 300,000 iterations,

and the residual sum of squares (RSS) calculated for each parameter set. Parameter sets were kept as an ‘accepted fit’ if the RSS met a threshold for statistical similarity to the best fit (Bufe *et al.*, 2024) (S7), where t_{95} represents the critical value of the two-tailed Student’s t-distribution evaluated for $n - 4$ degrees of freedom, and n is the number of samples in the dataset. The median and interquartile ranges were calculated from the population of accepted fits.

$$RSS_{threshold} = RSS_{best\ fit} \left(1 + t_{95} \sqrt{\frac{1}{n}} \right) \quad (S7)$$

S. 4. 2 Runoff and temperature parameterisation

The model assumes that erosion, runoff, and temperature vary independently. Runoff varies only by a factor of 2.3 in the study area and is not correlated with erosion ($R^2 = 0.31$), while temperature varies by a factor of 2.1 and is also not correlated with erosion ($R^2 = 0.17$). The approach employed by Bufo *et al.* (2024) models temperature sensitivity as an Arrhenius relation, which is consistent with some field measurements of OC_{petro} oxidation in erosive settings (Roylands *et al.*, 2022; Soulet *et al.*, 2021). However in this study the additional parameterisation did not improve the model fits and led other parameters to become unstable, therefore we exclude explicit temperature fitting and model $f(T)$ as a constant inside p_2 . Similarly, we exclude the runoff correction employed by Bufo *et al.*, (2024) on the basis that the small range in runoff relative to erosion rate in Taiwan makes this unnecessary, and tests including the correction destabilised some parameters. Instead, we model $f(q_w)$ as a constant inside p_2 .

S. 4. 3 Flux unit conversion

Weathering fluxes were fit in units of $t\ km^{-2}\ year^{-1}$, necessitating a conversion from per-Litre space via estimated runoff (Bufo *et al.*, 2024; Lee *et al.*, 2006) and catchment drainage areas (Bufo *et al.*, 2021). This meant samples were limited to those with erosion and runoff data, and we further excluded samples for which weathering fluxes exceeded erosion fluxes, and those

with erosion rates below the limits of the cosmogenic nuclide- ksn regression. In total, 85 samples were analysed in the weathering model.

S. 5 Incorporation of OC_{bio} burial

S. 5. 1 OC_{bio} burial erosion sensitivity and uncertainty

We use a global compilation of OC_{bio} and suspended sediment yields (Hilton *et al.*, 2025) to derive an erosion sensitivity of OC_{bio} burial, and incorporate this with our weathering carbon budgets to assess the overall response of the geological carbon cycle to erosion. A power law relationship of the form $Y_{OCbio} = \lambda Y_s^a$ was fit to the data compilation using ordinary least squares regression in $\log_{10}$ space (Fig. S8A). To characterise uncertainty in the power law fit we implemented a bootstrap resampling procedure in which the data compilation was resampled with replacement and refit over 5000 iterations, generating an ensemble of fits analogous to the accepted fits of the multiparameter weathering model (section 4.2, main text; section S.5.1) (Fig. S8B). Quartiles were calculated from the fit ensemble, and a ‘best’ fit defined as the bootstrap realisation closest to the ensemble median. To convert the fits from OC_{bio} yield – suspended sediment yield space to OC_{bio} burial – erosion yield space we incorporated factors for burial and suspended sediment fractions such that $(J_{OCbio-burial} * f_{burial}) = \lambda(E * f_{suspended})^a$. Uncertainty associated with these conversions was incorporated within the bootstrap framework based on Taiwan literature estimates. $F_{suspended}$ was sampled over a uniform distribution from 0.42 – 0.98 (Dadson *et al.*, 2003), and f_{burial} was sampled over a uniform distribution from 0.70 – 1.00 (Kao *et al.*, 2014).

S. 5. 2 Net carbon flux with OC_{bio} burial

The best fit and accepted fit ensemble from the OC_{bio} burial erosion sensitivity uncertainty analysis were propagated through the net carbon analysis alongside the best-fits and

ensembles for silicate, carbonate, sulfide, and OC_{petro} weathering as in main text section 5.2,
with adapted equations (S8-10).

$$[F_{CO_2}]_{st} = ([SO_4]^{eq} - (0.5[Cat_{carb}]^{eq} - [Cat_{sil}]^{eq})) + J_{OC_{\text{petro-oxidation}}} - J_{OC_{\text{bio-burial}}} \quad (S8)$$

$$[F_{CO_2}]_{mt} = (0.5([SO_4]^{eq} - [Cat_{sil}]^{eq})) + J_{OC_{\text{petro-oxidation}}} - J_{OC_{\text{bio-burial}}} \quad (S9)$$

$$[F_{CO_2}]_{lt} = -0.5[Cat_{sil}]^{eq} + J_{OC_{\text{petro-oxidation}}} - J_{OC_{\text{bio-burial}}} \quad (S10)$$

Supplementary Figures

Figure S1 – Correlations between IER and other environmental variables

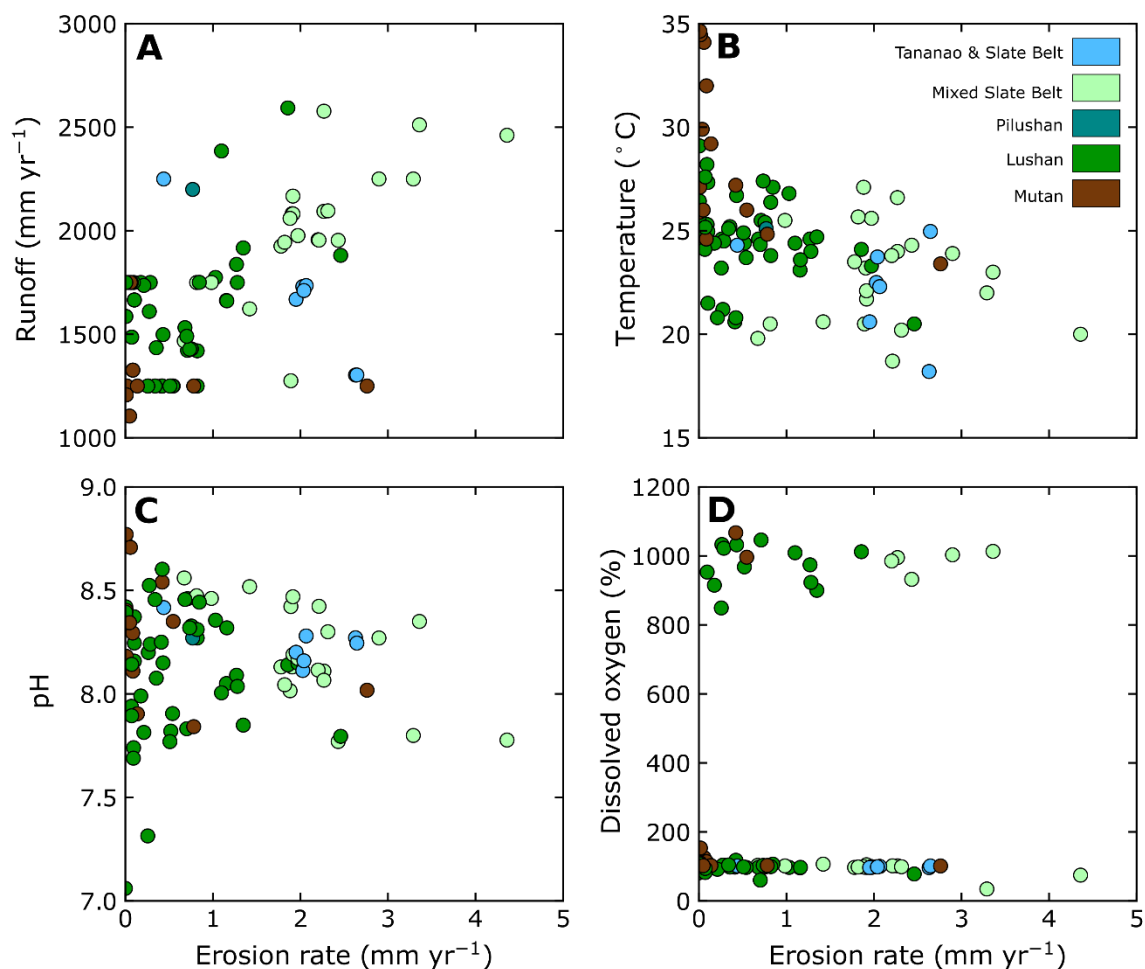
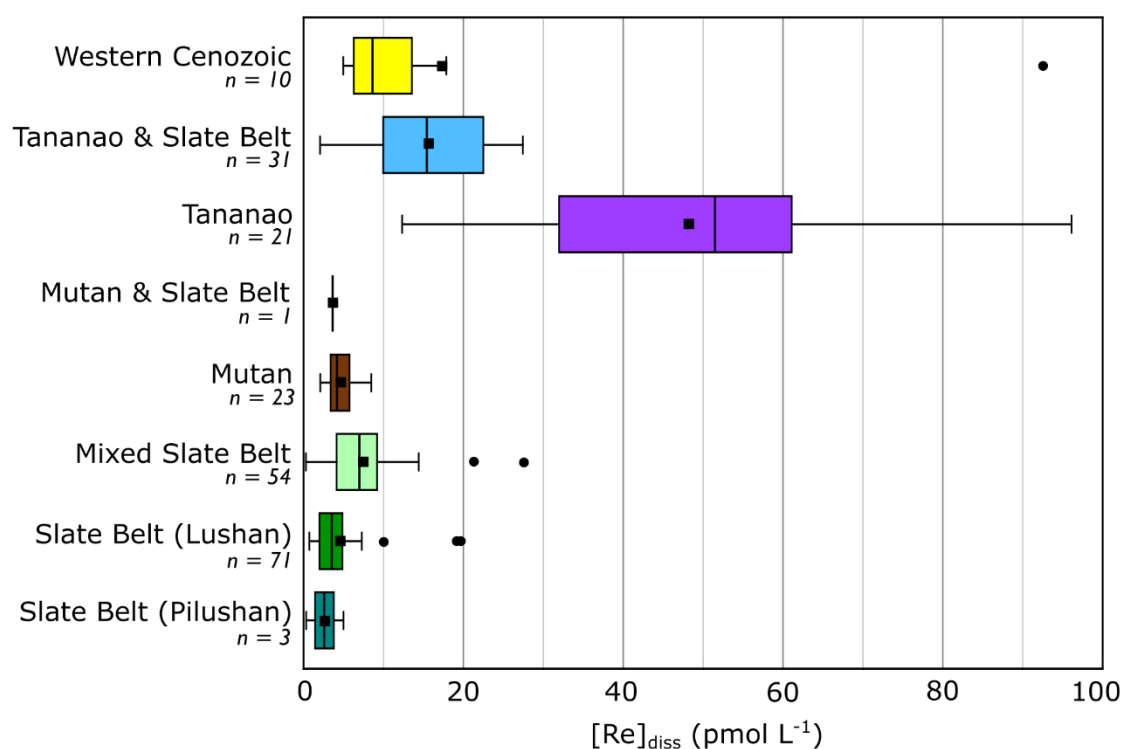


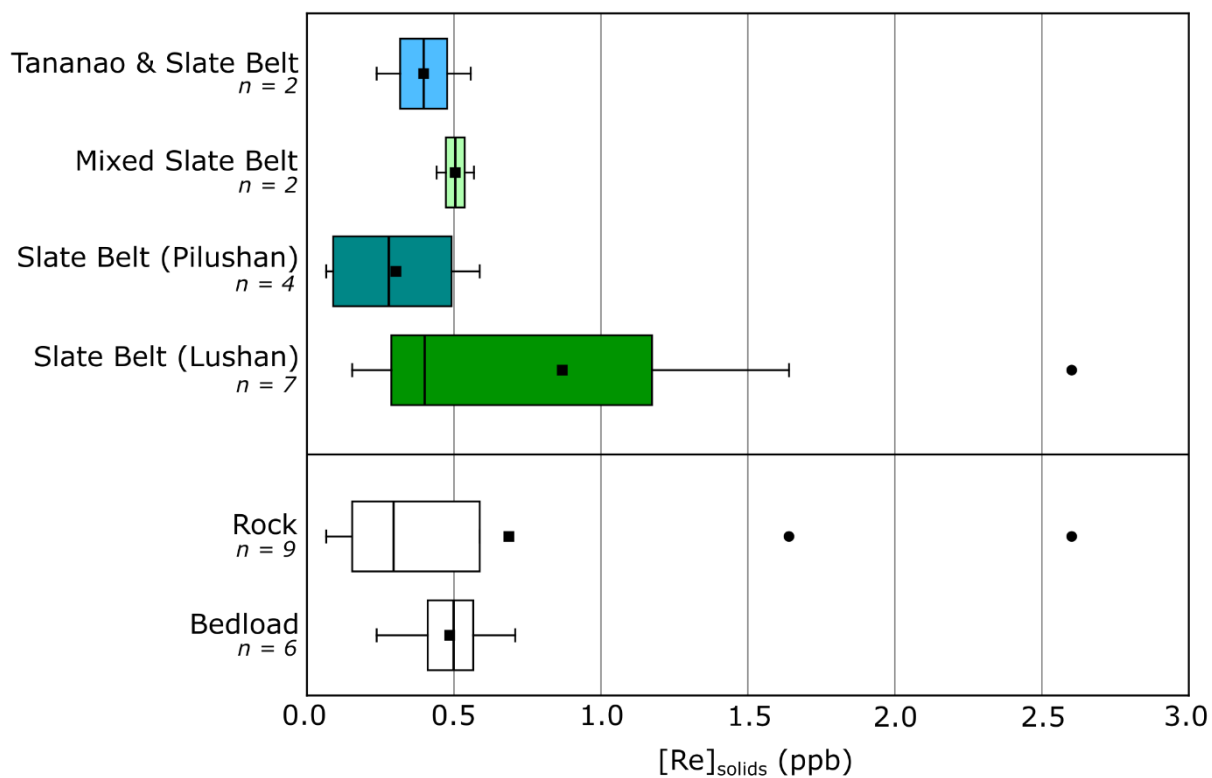
Fig. S1. Relationships between erosion rate and **A.** runoff, **B.** temperature, **C.** pH, **D.** dissolved oxygen across the study area, coloured by lithology. Erosion rate varies by 4.1 orders of magnitude, while runoff varies by 2.3 fold.

1125 *Figure S2 – $[Re]_{diss}$ boxplot by lithology*

1126 **Fig. S2.** Range in $[Re]_{diss}$ between lithologies. Boxes represent the interquartile ranges, with the
 1127 centre lines representing the medians. The means are represented by the black squares, and
 1128 outlying values by black circles. The excluding-outliers ranges of the samples are represented
 1129 by the extent of the whiskers.

1130

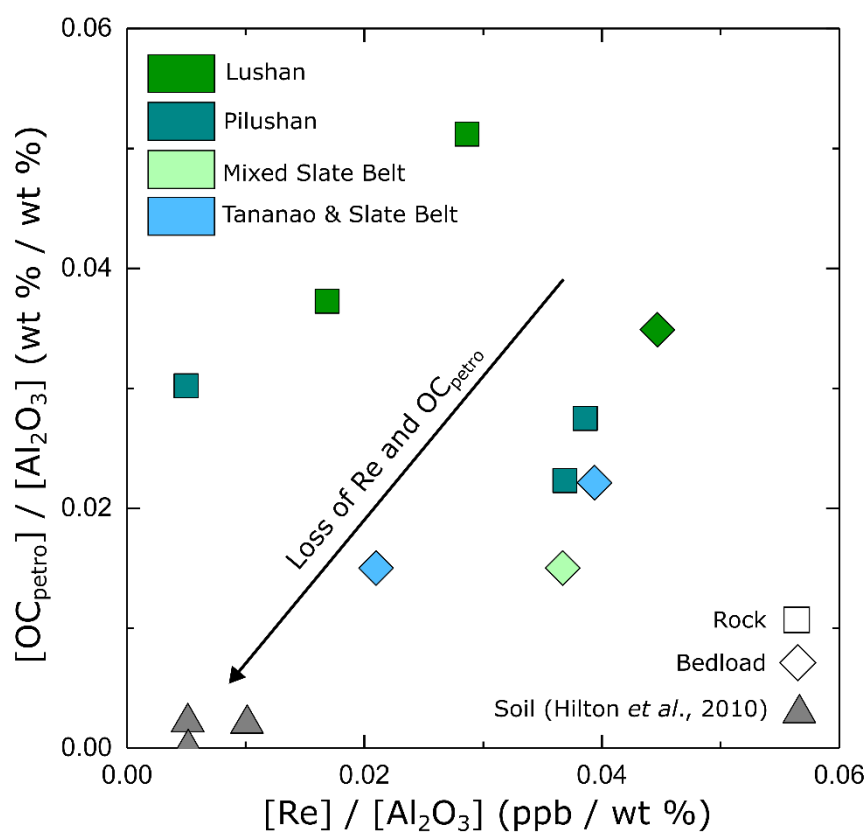
1131 *Figure S3 – $[Re]_{solids}$ boxplot by lithology and rock/bedload*



1132 **Fig. S3.** Range in $[Re]_{solids}$ categorised by lithology and sample type. Boxplot symbology follows

1133 [Fig. S2](#).

1134

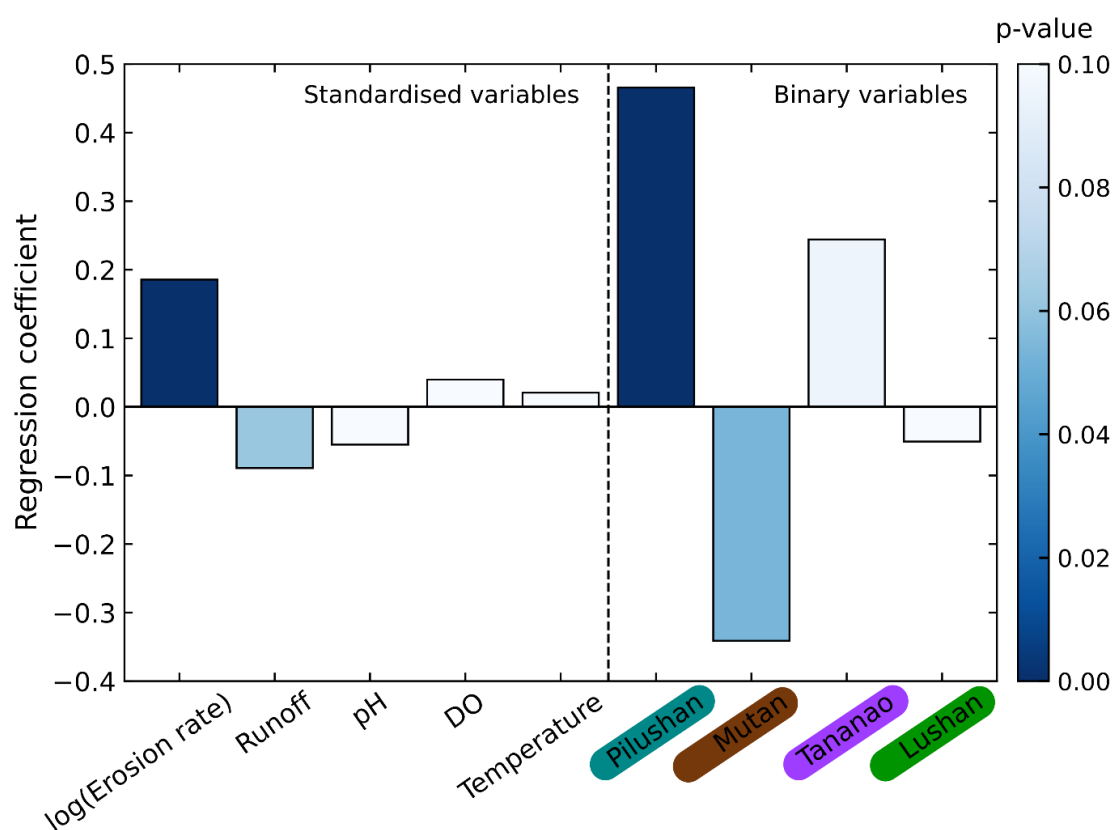
1135 *Figure S4- Coupled loss of OC_{petro} and Re during weathering*

1136

1137 **Fig. S4.** Coupled loss of OC_{petro} and Re from rocks and bedload sediments (squares and
 1138 rhombuses) to soils (triangles, Hilton *et al.*, 2010) during weathering, normalised to immobile
 1139 Al_2O_3 .

1140

1141 *Figure S5 – $[Re]_{OC_{petro}}$ multivariate OLS regression coefficients*



1142

1143 **Fig. S5.** Coefficients of multivariate OLS regression of log-transformed $[Re]_{OC_{petro}}$. Binary

1144 variables refer to the presence or absence of each lithology in the upstream catchment area.

1145 Bars are coloured by p-value where darker blue is more statistically significant and lighter blue

1146 is less significant.

1147

Figure S6 – Principal component analysis of $[Re]_{OC_{petro}}$ and environmental variables

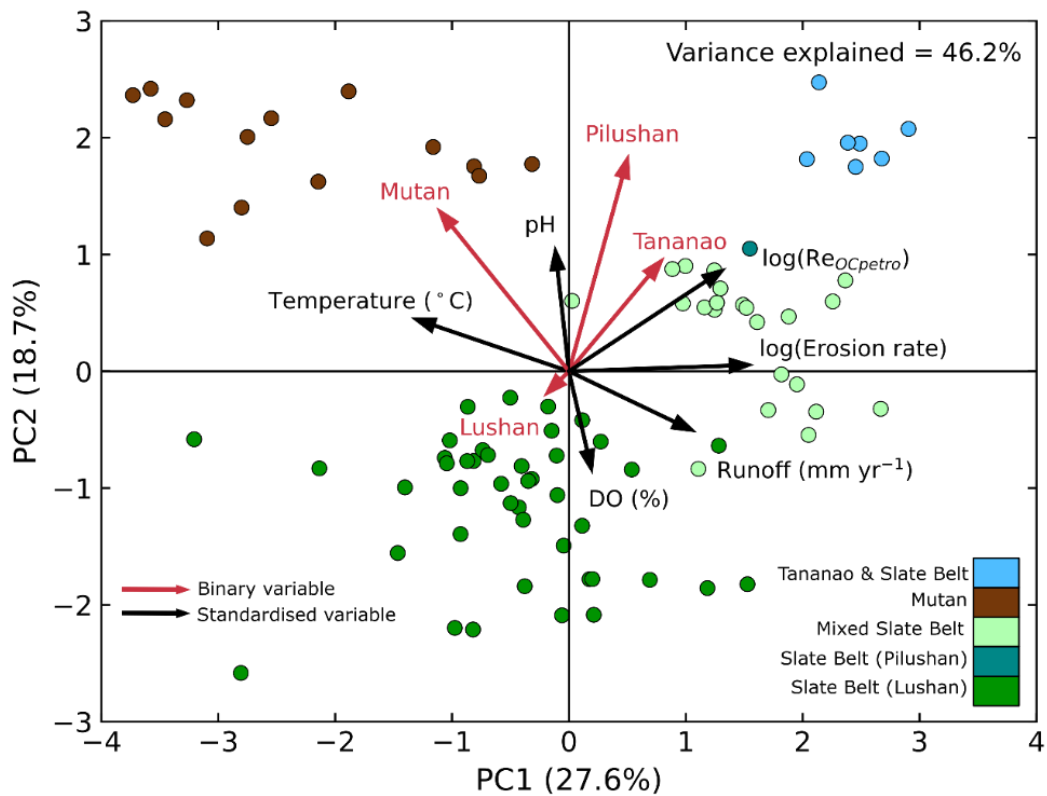
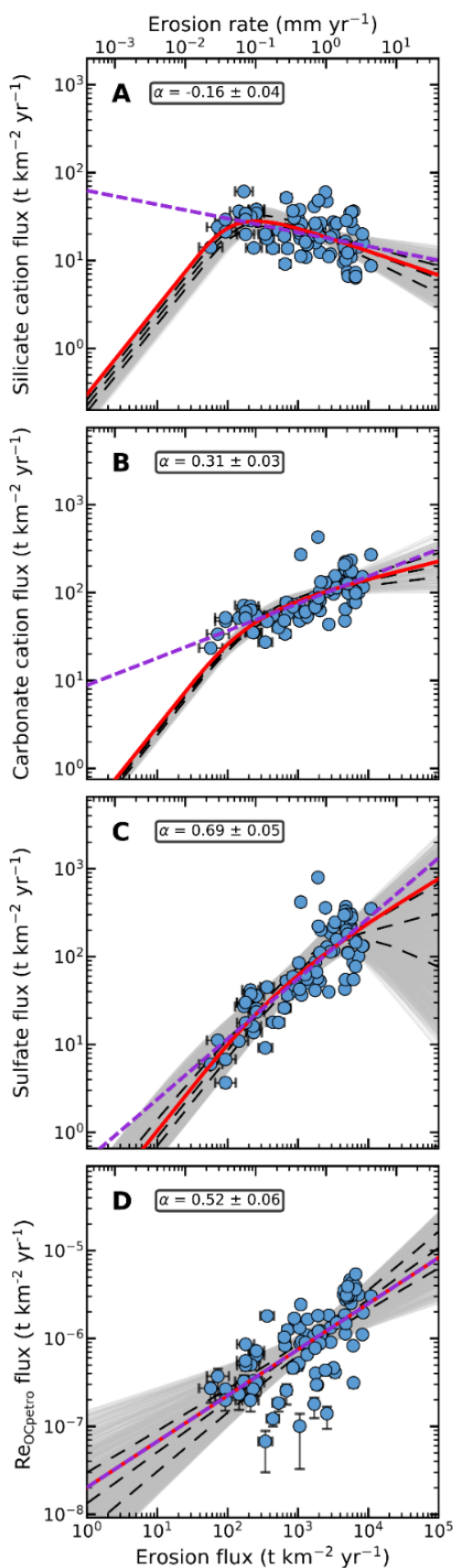
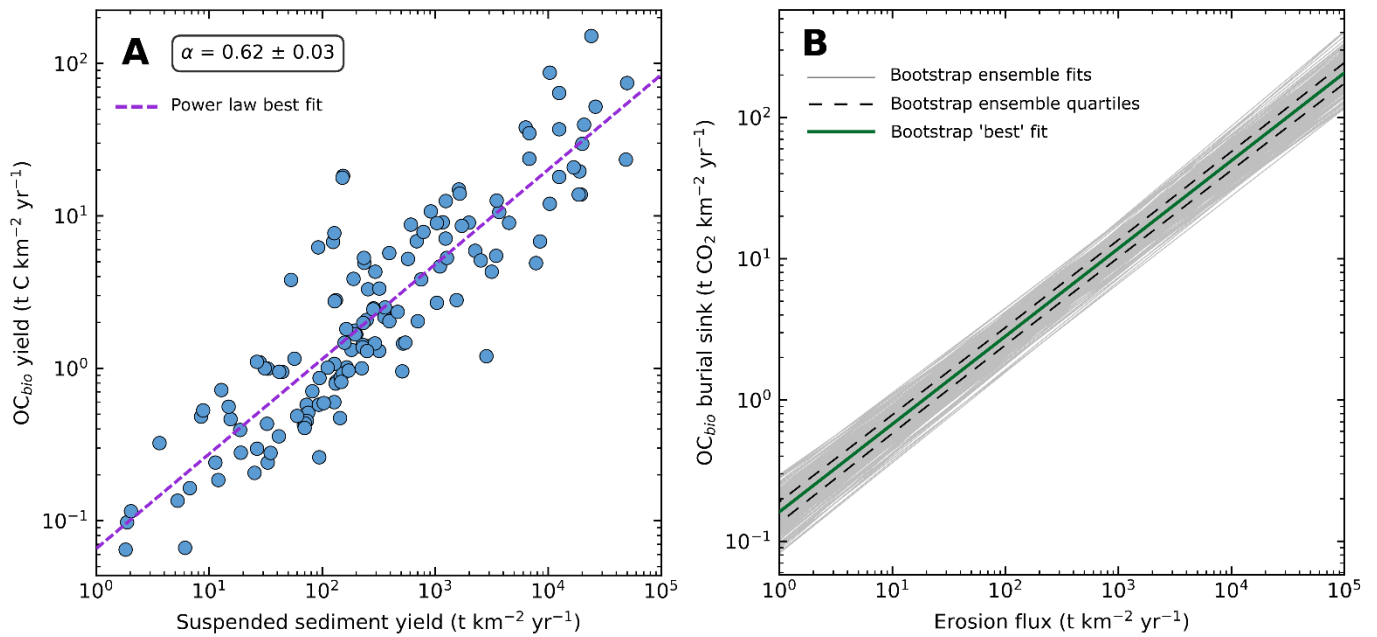


Fig. S6. PCA biplot showing the relationships between log-transformed $[Re]_{OC_{petro}}$, environmental parameters (including log-transformed erosion rate), and presence of different lithologies. Data points show positions of study samples while vectors show direction and strength of the association of each variable with the principal components. Standardised continuous environmental variables are shown in black, and presence of lithologies in red.

1158 *Figure S7 – Erosion sensitivity of silicate, carbonate, sulfide, $[Re]_{OCpetro}$ weathering fluxes***Fig. S7. A. Silicate, B. Carbonate, C. Sulfate, and D.**

$[Re]_{OCpetro}$ weathering flux data (blue circles), power law fit (dashed purple line), and best fit of the multiparameter model (red line). Uncertainties in the data points correspond to the 16th and 84th percentiles. Accepted fits (grey lines) are equivalent to a 95% confidence interval, and the quartiles (dashed black lines) correspond to the 25th, 50th, and 75th percentiles. The y axes each span four orders of magnitude. Erosion flux ($t\ km^{-2}\ yr^{-1}$) is equal to $2500 * \text{Erosion rate (mm yr}^{-1})$ assuming a crustal density of $2500\ kg\ m^{-3}$. The average erosion sensitivity (α) is displayed along with the standard error.

1173 *Figure S8 – OC_{bio} erosion sensitivity*

1174

1175 **Fig. S8. A.** Relationship between terrestrial OC_{bio} yield and suspended sediment yield.

1176 Compiled literature data (Hilton et al., 2025) is shown alongside the best fit power law, α , and

1177 the standard error in α . **B.** The erosion sensitivity of OC_{bio} burial estimated via the bootstrap

1178 resampling procedure and conversion to OC_{bio} burial flux and erosion flux space. The 'best' fit

1179 (i.e. the bootstrap realisation closest to the median) is shown in green, the bootstrap ensemble

1180 fits in grey, and their quartiles in dashed black. In each plot the y axis spans four order of

1181 magnitude as for Figs. 4 and S5.

Figure S9 – (+OC_{bio}) Erosion sensitivity of geological carbon cycle (+st, mt)

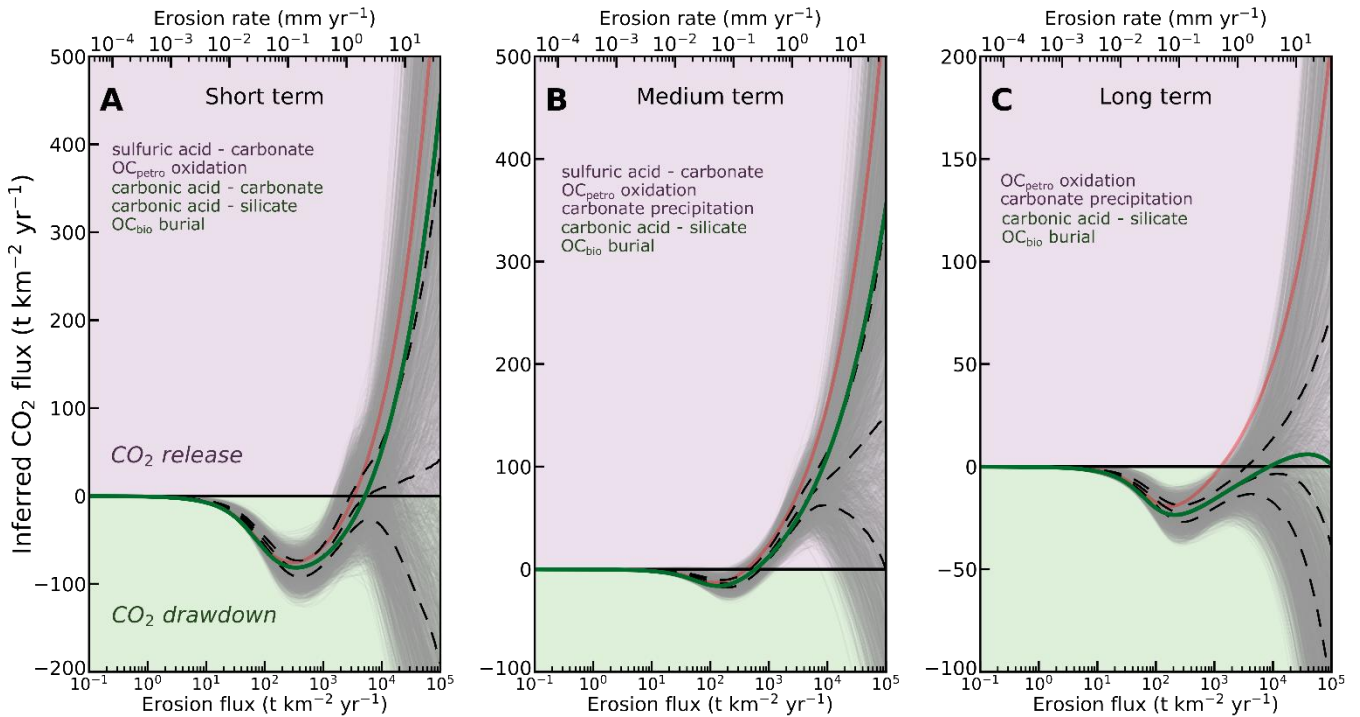


Fig. S9. Inferred net CO₂ fluxes including OC_{bio} burial on **A.** short timescales (< 10 kyr), **B.** medium timescales (10 kyr – 10 Myr), and **C.** long timescales (> 10 Myr), with the relevant reactions on each timescale labelled. The best fit of the analysis including OC_{bio} is shown in green, with the best fit of the analysis without OC_{bio} (i.e. from Fig. 5) in red. Accepted fits from the analysis including OC_{bio} (grey lines) are equivalent to a 95% confidence interval, and the quartiles (dashed black lines) correspond to the 25th, 50th, and 75th percentiles.

1202 Supplementary References

- 1203 Anbar, A.D., Creaser, R.A., Papanastassiou, D.A., Wasserburg, G.J., 1992. Rhenium in seawater:
1204 Confirmation of generally conservative behavior. *Geochim. Cosmochim. Acta* 56, 4099–
1205 4103. [https://doi.org/10.1016/0016-7037\(92\)90021-A](https://doi.org/10.1016/0016-7037(92)90021-A)
- 1206 Beyssac, O., Simoes, M., Avouac, J.P., Farley, K.A., Chen, Y.G., Chan, Y.C., Goffé, B., 2007. Late
1207 Cenozoic metamorphic evolution and exhumation of Taiwan. *Tectonics* 26.
1208 <https://doi.org/10.1029/2006TC002064>
- 1209 Bufe, A., Hovius, N., Emberson, R., Rugenstein, J.K.C., Galy, A., Hassenruck-Gudipati, H.J.,
1210 Chang, J.H., 2021. Co-variation of silicate, carbonate and sulfide weathering drives CO₂
1211 release with erosion. *Nat. Geosci.* 14, 211–216. [https://doi.org/10.1038/s41561-021-](https://doi.org/10.1038/s41561-021-00714-3)
1212 00714-3
- 1213 Bufe, A., Rugenstein, J.K.C., Hovius, N., 2024. CO₂ drawdown from weathering is maximized at
1214 moderate erosion rates. *Science* 383, 1075–1080.
1215 <https://doi.org/10.1126/SCIENCE.ADK0957>
- 1216 Colodner, D., Sachs, J., Ravizza, G., Turekian, K., Edmond, J., Boyle, E., 1993. The geochemical
1217 cycle of rhenium: a reconnaissance. *Earth Planet. Sci. Lett.* 117, 205–221.
1218 [https://doi.org/10.1016/0012-821X\(93\)90127-U](https://doi.org/10.1016/0012-821X(93)90127-U)
- 1219 Dadson, S.J., Hovius, N., Chen, H., Dade, W.B., Hsieh, M.-L., Willett, S.D., Hu, J.-C., Horng, M.-
1220 J., Chen, M.-C., Stark, C.P., Lague, D., Lin, J.-C., 2003. Links between erosion, runoff
1221 variability and seismicity in the Taiwan orogen. *Nature* 426, 648–651.
1222 <https://doi.org/https://doi.org/10.1038/nature02150>
- 1223 Dalai, T.K., Singh, S.K., Trivedi, J.R., Krishnaswami, S., 2002. Dissolved rhenium in the Yamuna
1224 River System and the Ganga in the Himalaya: Role of black shale weathering on the

- 1225 budgets of Re, Os, and U in rivers and CO₂ in the atmosphere. *Geochim. Cosmochim.*
1226 *Acta* 66, 29–43. [https://doi.org/10.1016/S0016-7037\(01\)00747-5](https://doi.org/10.1016/S0016-7037(01)00747-5)
- 1227 Dellinger, M., Hilton, R.G., Baronas, J.J., Torres, M.A., Burt, E.I., Clark, K.E., Galy, V., Ccahuana
1228 Quispe, A.J., West, A.J., 2023. High rates of rock organic carbon oxidation sustained as
1229 Andean sediment transits the Amazon foreland-floodplain. *Proc. Natl. Acad. Sci.* 120.
1230 <https://doi.org/10.1073/PNAS.2306343120>
- 1231 Dellinger, M., Hilton, R.G., Nowell, G.M., 2020. Measurements of rhenium isotopic composition
1232 in low-abundance samples. *J. Anal. At. Spectrom.* 35, 377–387.
1233 <https://doi.org/10.1039/c9ja00288j>
- 1234 Dickson, A.J., Hsieh, Y.T., Bryan, A., 2020. The rhenium isotope composition of Atlantic Ocean
1235 seawater. *Geochim. Cosmochim. Acta* 287, 221–228.
1236 <https://doi.org/10.1016/j.gca.2020.02.020>
- 1237 Fellin, M.G., Chen, C.Y., Willett, S.D., Christl, M., Chen, Y.G., 2017. Erosion rates across space
1238 and timescales from a multi-proxy study of rivers of eastern Taiwan. *Glob. Planet.*
1239 *Change* 157, 174–193. <https://doi.org/10.1016/J.GLOPLACHA.2017.07.012>
- 1240 Hilton, R.G., Gaillardet, J., Calmels, D., Birck, J.-L., 2014. Geological respiration of a mountain
1241 belt revealed by the trace element rhenium. *Earth Planet. Sci. Lett.* 403, 27–36.
1242 <https://doi.org/10.1016/j.epsl.2014.06.021>
- 1243 Hilton, R.G., Galy, A., Hovius, N., Horng, M.-J., Chen, H., 2010. The isotopic composition of
1244 particulate organic carbon in mountain rivers of Taiwan. *Geochim. Cosmochim. Acta*
1245 74, 3164–3181. <https://doi.org/10.1016/j.gca.2010.03.004>
- 1246 Hilton, R.G., Schwab, M.S., Galy, V., 2025. Dynamics of particulate organic carbon mobilization,
1247 storage, and export across river sedimentary systems, in: *Treatise on Geochemistry*.

- 1248 Elsevier, pp. 215–250. [https://doi.org/https://doi.org/10.1016/B978-0-323-99762-](https://doi.org/10.1016/B978-0-323-99762-1.00021-8)
- 1249 1.00021-8
- 1250 Hilton, R.G., Turowski, J.M., Winnick, M., Dellinger, M., Schleppe, P., Williams, K.H., Lawrence,
- 1251 C.R., Maher, K., West, M., Hayton, A., 2021. Concentration-Discharge Relationships of
- 1252 Dissolved Rhenium in Alpine Catchments Reveal Its Use as a Tracer of Oxidative
- 1253 Weathering. *Water Resour. Res.* 57. <https://doi.org/10.1029/2021WR029844>
- 1254 Ho, C.S., 1986. A synthesis of the geologic evolution of Taiwan. *Tectonophysics, Geodynamics*
- 1255 *of the Eurasia\3-Philippine Sea Plate Boundary* 125, 1–16. [https://doi.org/10.1016/0040-](https://doi.org/10.1016/0040-1951(86)90004-1)
- 1256 1951(86)90004-1
- 1257 Horan, K., Hilton, R.G., Dellinger, M., Tipper, E., Galy, V., Calmels, D., Selby, D., Gaillardet, J.,
- 1258 Ottley, C.J., Parsons, D.R., Burton, K.W., 2019. Carbon dioxide emissions by rock
- 1259 organic carbon oxidation and the net geochemical carbon budget of the Mackenzie
- 1260 River Basin. *Am. J. Sci.* 319, 473–499. <https://doi.org/10.2475/06.2019.02>
- 1261 Jaffe, L.A., Peucker-Ehrenbrink, B., Petsch, S.T., 2002. Mobility of rhenium, platinum group
- 1262 elements and organic carbon during black shale weathering. *Earth Planet. Sci. Lett.*
- 1263 198, 339–353. [https://doi.org/10.1016/S0012-821X\(02\)00526-5](https://doi.org/10.1016/S0012-821X(02)00526-5)
- 1264 Kao, S.-J., Hilton, R.G., Selvaraj, K., Dai, M., Zehetner, F., Huang, J.-C., Hsu, S.-C., Sparkes, R.,
- 1265 Liu, J.T., Lee, T.-Y., Yang, J.-Y.T., Galy, A., Xu, X., Hovius, N., 2014. Preservation of
- 1266 terrestrial organic carbon in marine sediments offshore Taiwan: mountain building and
- 1267 atmospheric carbon dioxide sequestration. *Earth Surf. Dyn.* 2, 127–139.
- 1268 <https://doi.org/10.5194/esurf-2-127-2014>
- 1269 Lee, C.-S., Huang, L.-R., Shen, H.-S., Wang, S.-T., 2006. A Climatology Model for Forecasting
- 1270 Typhoon Rainfall in Taiwan. *Nat. Hazards* 37, 87–105. [https://doi.org/10.1007/s11069-](https://doi.org/10.1007/s11069-005-4658-8)
- 1271 005-4658-8

- 1272 Lin, C.W., Chen, W.S., 2016. Geologic map of Taiwan.
- 1273 Miller, C.A., Peucker-Ehrenbrink, B., Walker, B.D., Marcantonio, F., 2011. Re-assessing the
1274 surface cycling of molybdenum and rhenium. *Geochim. Cosmochim. Acta* 75, 7146–
1275 7179. <https://doi.org/10.1016/j.gca.2011.09.005>
- 1276 Millero, F.J., Feistel, R., Wright, D.G., McDougall, T.J., 2008. The composition of Standard
1277 Seawater and the definition of the Reference-Composition Salinity Scale. *Deep Sea*
1278 *Res. Part Oceanogr. Res. Pap.* 55, 50–72. <https://doi.org/10.1016/j.dsr.2007.10.001>
- 1279 Peucker-Ehrenbrink, B., Hannigan, R.E., 2000. Effects of black shale weathering on the mobility
1280 of rhenium and platinum group elements. *Geology* 28, 475.
1281 [https://doi.org/10.1130/0091-7613\(2000\)28%3C475:EOBSWO%3E2.0.CO;2](https://doi.org/10.1130/0091-7613(2000)28%3C475:EOBSWO%3E2.0.CO;2)
- 1282 Ravizza, G., Esser, B.K., 1993. A possible link between the seawater osmium isotope record and
1283 weathering of ancient sedimentary organic matter. *Chem. Geol.* 107, 255–258.
1284 [https://doi.org/10.1016/0009-2541\(93\)90186-M](https://doi.org/10.1016/0009-2541(93)90186-M)
- 1285 Ravizza, G., Turekian, K.K., 1989. Application of the ¹⁸⁷Re-¹⁸⁷Os system to black shale
1286 geochronometry. *Geochim. Cosmochim. Acta* 53, 3257–3262.
1287 [https://doi.org/10.1016/0016-7037\(89\)90105-1](https://doi.org/10.1016/0016-7037(89)90105-1)
- 1288 Ravizza, G., Turekian, K.K., Hay, B.J., 1991. The geochemistry of rhenium and osmium in recent
1289 sediments from the Black Sea. *Geochim. Cosmochim. Acta* 55, 3741–3752.
1290 [https://doi.org/10.1016/0016-7037\(91\)90072-D](https://doi.org/10.1016/0016-7037(91)90072-D)
- 1291 Rout, R.K., Tripathy, G.R., 2024. Net effect of chemical erosion in a tropical basin on carbon
1292 cycle: Constraints from elemental and sulfur isotopic composition of the Mahanadi
1293 river water. *Chem. Geol.* 644, 121859. <https://doi.org/10.1016/j.chemgeo.2023.121859>
- 1294 Roylands, T., Hilton, R.G., Garnett, M.H., Soulet, G., Newton, J.-A., Peterkin, J.L., Hancock, P.,
1295 2022. Capturing the short-term variability of carbon dioxide emissions from

- 1296 sedimentary rock weathering in a remote mountainous catchment, New Zealand.
1297 Chem. Geol. 608, 121024. <https://doi.org/10.1016/j.chemgeo.2022.121024>
- 1298 Selby, D., Creaser, R.A., 2003. Re-Os geochronology of organic rich sediments: an evaluation of
1299 organic matter analysis methods. Chem. Geol. 200, 225–240.
1300 [https://doi.org/10.1016/S0009-2541\(03\)00199-2](https://doi.org/10.1016/S0009-2541(03)00199-2)
- 1301 Soulet, G., Hilton, R.G., Garnett, M.H., Roylands, T., Klotz, S., Croissant, T., Dellinger, M., Le
1302 Bouteiller, C., 2021. Temperature control on CO₂ emissions from the weathering of
1303 sedimentary rocks. Nat. Geosci. 14, 665–671. [https://doi.org/10.1038/s41561-021-](https://doi.org/10.1038/s41561-021-00805-1)
1304 00805-1
- 1305 Wai, K.M., Lin, N.-H., Wang, S.-H., Dokiya, Y., 2008. Rainwater chemistry at a high-altitude
1306 station, Mt. Lulin, Taiwan: Comparison with a background station, Mt. Fuji. J. Geophys.
1307 Res. Atmos. 113. <https://doi.org/https://doi.org/10.1029/2006JD008248>
- 1308 Wang, W., Dickson, A.J., Stow, M.A., Dellinger, M., Burton, K.W., Savage, P.S., Hilton, R.G.,
1309 Prytulak, J., 2024. Rhenium elemental and isotopic variations at magmatic
1310 temperatures. <https://doi.org/10.7185/geochemlet.2402>
1311