

Oasis management and topography interactively shape soil inorganic carbon dynamics in hyper-arid soils

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ABSTRACT

Soil inorganic carbon (SIC) constitutes a major component of total soil carbon in arid and hyper-arid regions and plays an important role in terrestrial carbon storage. SIC originates from both lithogenic and pedogenic carbonates formed through coupled biological-geochemical processes. However, the extent to which oasis management influences SIC dynamics remains unclear, particularly under potentially acidifying conditions associated with fertilisation and enhanced biological activity. We investigated SIC dynamics in southwestern Tunisia by sampling soils to 120 cm depth across three traditional and three modern oasis systems, along topographic gradients and under contrasting tree microsites. SIC accounted for 76% (median) of total soil C across all samples, confirming its dominance under hyper-arid conditions. SIC contents and stocks (mean: 31.3 kg m⁻²) were not significantly affected by oasis management or tree position. Instead, SIC patterns were structured by interactions among management, topography, and soil depth. Along the catena, SIC declined downslope in modern oases but increased downslope in traditional systems, highlighting context-dependent topographic organization rather than uniform management effects on SIC dynamics. Carbon isotopic signatures ($\delta^{13}\text{C}$ -SIC) and isotope mixing models indicated that 20–40% of SIC was derived from soil-respired CO₂, with higher contributions in traditional oasis systems and surface soils. The negative relationship between $\delta^{13}\text{C}$ -SIC and SOC further supports coupling between organic and inorganic carbon pools. Overall, our results indicate that SIC dynamics in oasis systems are strongly buffered by geological and hydrological controls, while management primarily influences carbonate cycling and source partitioning rather than total SIC storage.

1. Introduction

Soil inorganic carbon (SIC) plays a crucial role in the global carbon (C) cycle and in soil functioning (Sharififar et al., 2023; Batool et al., 2024). Together, SIC and soil organic carbon (SOC) contribute comparably to global soil C storage, with approximately 2255 Pg SIC and 2400 Pg SOC to 2 m depth (Monger et al., 2015). The global SIC pool is predominantly located in semiarid, arid, and hyper-arid regions, which cover ~40% of the terrestrial land surface (An et al., 2019; FAO, 2019;

Ferdush and Paul, 2021) and where SIC typically accounts for 70–80% of total soil C stocks (Filippi et al., 2020). SIC comprises primary (lithogenic) and secondary (pedogenic) carbonates: lithogenic carbonates are inherited from sedimentary parent materials such as limestone, dolomite, and marl, whereas pedogenic carbonates form in situ within soils, predominantly as calcite (CaCO₃) and, less frequently, as aragonite or Mg-rich calcite (Zamanian et al., 2016). Although drylands are increasingly recognised as important components of the global carbon cycle, their main carbon reservoir, SIC, remains understudied relative to

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SOC.

Carbon isotopic composition ($\delta^{13}\text{C}$) provides a key tool to distinguish carbonate sources. Lithogenic carbonates typically retain the isotopic signature of their parent material (around 0‰ to +3‰), whereas pedogenic carbonates reflect the $\delta^{13}\text{C}$ of soil CO_2 , which is governed by vegetation type and biological activity (Kuz'yakov and Bol, 2005). This allows inference of whether carbonate formation is derived from atmospheric CO_2 (~–8‰) or from CO_2 produced by root and microbial respiration (Nordt et al., 1998).

Pedogenic SIC formation proceeds through secondary carbonate precipitation under specific environmental controls (Mikhailova et al., 2013; Lal et al., 2018; Batool et al., 2024). In drylands, precipitation occurs when soil porewaters become supersaturated, typically as a result of evaporative concentration of solutes, CO_2 degassing, or increased availability of Ca^{2+} and Mg^{2+} (Zamanian et al., 2016). This balance is regulated by interacting physicochemical factors: higher temperatures reduce CaCO_3 solubility, while strong evaporation increases ionic strength and alkalinity until precipitation is induced. Soil pCO_2 and pH exert dual controls, with elevated CO_2 from root and microbial respiration promoting carbonate dissolution (Shi et al., 2012; Obear and Soldat, 2016). The supply of Ca^{2+} from mineral weathering, aeolian inputs, or irrigation water provides the essential cations driving these processes. Consequently, in hyper-arid environments, the combined effects of high evapotranspiration, alkalinity, and cation availability largely determine rates of pedogenic SIC accumulation (Ulgiichimeg, 2021; Zhang et al., 2022).

Land use and land use change strongly influence SIC dynamics in arid systems through irrigation, fertilisation, and cropping (An et al., 2019; Dang et al., 2022). Irrigation can increase SIC by supplying Ca^{2+} and alkalinity, but excessive water inputs may enhance leaching and carbonate loss (An et al., 2019; Dang et al., 2022). In contrast, nitrogen (N) fertilisation generally promotes SIC depletion, as nitrification-induced acidification enhances carbonate dissolution and CO_2 release, particularly in calcareous soils under high N inputs (Zamanian et al., 2018, 2024; Dang et al., 2022). Cropping further regulates SIC through rhizosphere processes and plant–soil–microbe interactions: root and microbial respiration increase pCO_2 and promote dissolution under moist conditions, whereas drying induces CO_2 degassing, higher pH, and carbonate precipitation along evaporation fronts (Zamanian et al., 2016; Liu et al., 2021; Pan et al., 2022; Shao et al., 2023; Dong et al., 2024). Consequently, the net effect of land use on SIC reflects the balance between dissolution and re-precipitation processes, governed by Ca^{2+} availability, management intensity, and the irrigation–evapotranspiration regime.

In Tunisia, SIC exhibits high spatial variability due to the interplay of geological, climatic, and anthropogenic factors (Brahim et al., 2014, 2021; Jaouadi et al., 2016). Tunisian oasis agroecosystems provide an ideal system to investigate these controls under hyper-arid conditions, given their long cultivation history, the development of Anthrosols, and the contrast between traditional and intensively managed oasis systems (Table S1). Despite this, the combined effects of management and topography on SIC dynamics in such systems remain insufficiently understood.

To address this knowledge gap, we investigated how management and topography regulate SIC contents, stocks, and their contribution to total soil C. The study was conducted in the Tozeur area (Djerid region, southwestern Tunisia), where we compared three traditional oases characterized by long-term organic inputs and high date palm density with three modern oases characterized by lower plant diversity and greater reliance on synthetic fertilisation. Within each oasis, soils were sampled along a topographic gradient (upslope, midslope, downslope) and across paired microsites (under and between date palms) to 120 cm depth. We quantified SIC (including $\delta^{13}\text{C}$ for source partitioning), SOC, salinity, pH, soil texture, extractable cations and anions, and gypsum to resolve the controls on carbonate dynamics.

Building on current mechanistic understanding of SIC dynamics, we

hypothesized that (H1) intensive management, particularly N fertilisation, promotes soil acidification and enhances carbonate dissolution; (H2) topographic position controls SIC distribution along the catena through its influence on groundwater proximity, salinity, and dissolved inorganic carbon (DIC), with downslope positions favoring carbonate accumulation; and (H3) long-term management shifts carbonate pools from lithogenic to pedogenic forms via biologically driven dissolution–reprecipitation processes, reflected in more depleted $\delta^{13}\text{C}$ signatures of SIC.

2. Materials and methods

2.1. Study area and sampling design

The study was conducted in the Djerid region of southern Tunisia (Tozeur, Deguache, and Nefta; Fig. 1), a hyper-arid area characterized by gypsum-rich Quaternary deposits. According to the UNEP aridity index, the region receives 76–86 mm mean annual precipitation (Table 1) and is strongly influenced by the hypersaline Chott El Djerid (hereafter: Chott), which controls groundwater depth and salinity patterns. Oasis agriculture in the region comprises two main management systems: traditional oases, characterized by dense, diverse vegetation and multi-layered cropping, and modern oases with lower plant diversity, aligned palm planting, and greater reliance on external inputs ((Battesti, 2005); Table S1).

To capture topographic controls, we defined a catena based on relative elevation and distance to the Chott. Upslope positions (slope ~2–3%) are located 5–6 km from the Chott, midslope positions represent transitional areas, and downslope positions (slope <0.8%) occur within 0.5–1 km of the Chott margin. Within each of three areas, one traditional and one modern oasis were selected, resulting in 18 plots (3 areas $\times$ 2 oasis types $\times$ 3 topographic positions).

Soil sampling was conducted in May–June 2023. Within each plot, samples were collected from two microsites (under date palms and interspaces; ~8 m spacing) at six depth increments (0–5, 5–10, 10–30, 30–60, 60–90, and 90–120 cm), yielding 216 samples in total. Samples were sieved (<2 mm), air-dried, and stored prior to analysis. Bulk density was determined from volumetric samples collected in eight representative soil pits (0–120 cm), with measurements used to validate model-based estimates for the full dataset.

2.2. Laboratory analysis

Soil pH was measured in a 1:2.5 soil-to-water suspension, prepared by mixing soil with purified deionized water, using a direct-reading pH meter probe (inoLab pH 7110). Electrical conductivity (EC) was measured potentiometrically in a 1:5 soil-to-water extract using a Consort C862 conductometer with a cell constant of $K = 1 \text{ cm}^{-1}$ (Rhoades et al., 2018). EC values were converted to saturation extract equivalents (EC_e) using a texture-dependent factor (7 for sandy soils) and expressed as approximate salinity i.e. total dissolved solids, $\text{TDS} (\text{mg L}^{-1}) \approx \text{EC}_e (\text{dS m}^{-1}) \times 640$; divide by 58.5 g mol^{-1} to mmolc L^{-1} assuming NaCl equivalence (Rhoades, 2018). Soil texture was determined following the simplified sedimentation method of (Kettler et al., 2001) in which sand (>50 μm ; French system) is separated by sieving and the remaining fine fraction is partitioned into silt and clay based on particle settling times in water.

Soil total carbon (STC) and soil organic carbon (SOC) were determined using an elemental analyzer, with SOC measured after acid pre-treatment to remove carbonates and soil inorganic carbon (SIC) derived by difference. Soil samples (~10 g) were oven-dried at 105 °C, coarsely ground with mortar and pestle, and then finely ground in a ball mill at 25 Hz for 5 min (Retsch MM200, Hanau). For STC and total nitrogen (STN), 6–12 mg of untreated material was weighed into tin capsules and analyzed. For SOC, an aliquot (~50 mg) was treated with 2 M HCl for 60 min to remove inorganic carbonates, oven-dried again at 105 °C, and

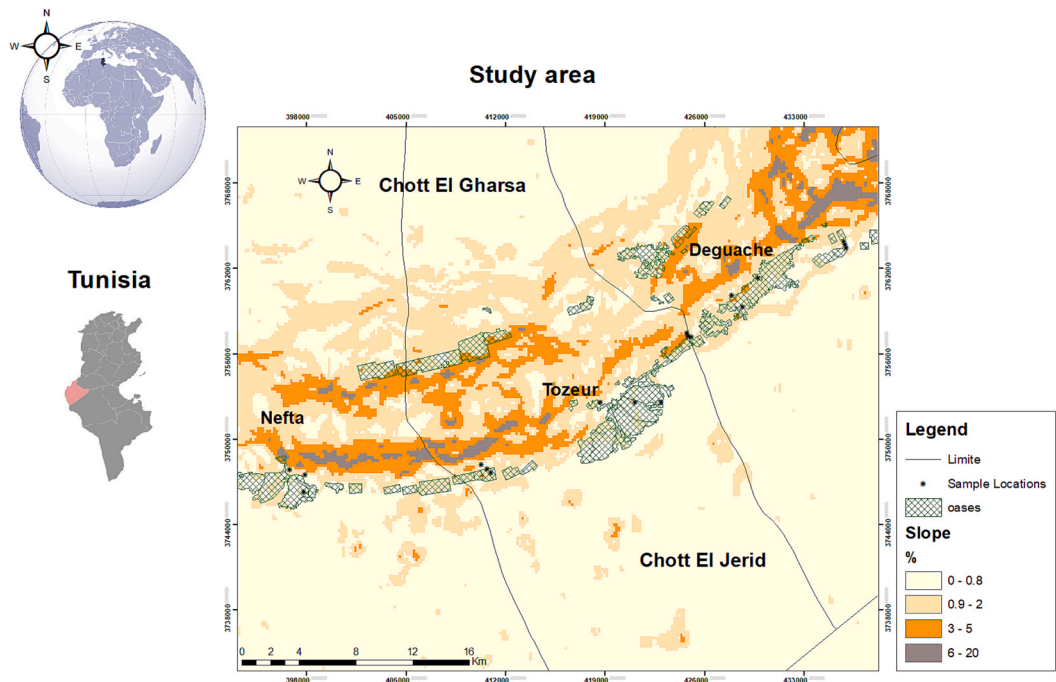


Fig. 1. Localization of study areas (modern and traditional oasis systems in Nefta, Tozeur, and Deguache) in Tunisia, with respective slope maps.

Table 1
Characterisation of modern and traditional oasis systems in the three study areas (Deguache, Tozeur, and Nefta) in Tunisia, with geographic coordinates, climate, geology, soil type and aquifer given.

Area	Deguache		Tozeur		Nefta	
Oasis management	Modern	Traditional	Modern	Traditional	Modern	Traditional
Latitude	34°0'42.08"	33°59'17.758"	33°57'9.929"	33°54'39.186"	33°51'53.996"	33°51'29.365"
Longitude	8°18'15.448"	8°14'18.143"	8°11'11.512"	8°8'37.018"	8°1'2.183"	7°53'24.878"
Elevation (m a.s.l.)	30–35	30–54	30–53	35–42	35–45	35–42
Geology	Upper Cretaceous	Upper Cretaceous	Quaternary	Quaternary	Quaternary	Quaternary
Parent Material	Quaternary gypsiferous deposits	Quaternary gypsiferous deposits	Quaternary alluvial deposits	Quaternary gypsiferous deposits	Quaternary deposits	Quaternary gypsiferous deposits
Soil type	Regosols	Regosols, Gypsisols	Regosols	Regosols, Gypsisols and Hydromorphic soils	Regosols, Gypsisols	Regosols, Gypsisols
Soil Texture	Loamy Sand	Sandy Clay Loam	Loamy Sand	Loamy Sand	Sandy Loam	Sandy Loam
Coarse Fragments (> 2 mm)	Negligible (<2%)	Negligible (<2%)	Negligible (<2%)	Negligible (<2%)	Negligible (<2%)	Negligible (<2%)
Mean annual temperature (C)	22.3	22.3	22.4	22.4	22.1	22.1
Mean annual Precipitation (mm)	85.8	85.8	81.7	81.7	76	76
Irrigation Source (Aquifer)	"Complexe Terminal" (CT)	"Complexe Terminal" (CT)	"Complexe Terminal" (CT)	"Complexe Terminal" (CT)	"Complexe Terminal" (CT)	"Complexe Terminal" (CT)

6–12 mg weighed into tin capsules for analysis. SIC was calculated as SIC = STC – SOC. Thereafter, STN, STC, SOC, and their isotope ratios were measured using an elemental analyzer (EA Isolink) coupled via a ConFlo IV interface to a Delta V Advantage isotope ratio mass spectrometer (Thermo Scientific, Vienna, Austria). Stable carbon isotope ratios are reported in delta notation ($\delta^{13}\text{C}$, ‰) relative to the international V-PDB standard, calculated as

$$\delta^{13}\text{C} (\text{‰ vs V - PDB}) = \left(\frac{R_{\text{sample}}}{R_{\text{standard}}} - 1 \right) \times 1000$$

with R representing the $^{13}\text{C}/^{12}\text{C}$ ratio and R_{standard} being R of Vienna-PeeDee Belemnite (V-PDB).

The C isotope composition of SIC ($\delta^{13}\text{C}_{\text{SIC}}$) was calculated using a mixing model, as follows:

$$\delta^{13}\text{C}_{\text{SIC}} = \frac{\delta^{13}\text{C}_{\text{STC}} \times C_{\text{STC}} - \delta^{13}\text{C}_{\text{SOC}} \times C_{\text{SOC}}}{C_{\text{STC}} - C_{\text{SOC}}}$$

where C_{STC} and C_{SOC} are the contents of STC and SOC and $\delta^{13}\text{C}_{\text{STC}}$ and $\delta^{13}\text{C}_{\text{SOC}}$ their respective carbon isotope signatures.

Gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) was quantified using a recently developed dissolution–ion chromatography approach. Conventional gravimetric or acetone precipitation methods were not used due to their limited specificity and sensitivity (Álvarez et al., 2022). Instead, 50 mg of finely ground, oven-dry soil was extracted with 25 mL ultrapure water (Milli-Q) at room temperature for 18 h to ensure complete dissolution of gypsum. Samples were subsequently centrifuged (10,000 rpm, 10 min), and Ca^{2+} and SO_4^{2-} concentrations in the supernatant were determined by high-performance ion chromatography (HP-IC; Dionex ICS5000, Thermo Scientific, Vienna) using a Dionex IonPac CS16 column for cations and an IonPac AS11-HC column for anions, each with respective

suppressors (CERS 500 and AERS 500). Complete dissolution is ensured by the high solubility of gypsum ($\approx 20\text{--}25\text{ g L}^{-1}$ at $25\text{ }^{\circ}\text{C}$) relative to the extraction ratio. Gypsum content was calculated assuming stoichiometric precipitation based on the limiting ion (Ca^{2+} or SO_4^{2-}), expressed on a molar basis. This assumption is appropriate under hyper-arid conditions, where gypsum precipitation proceeds to completion until one ion becomes limiting. Prior tests using replicate analyses demonstrated high reproducibility of the method, with relative standard deviations of 5–10%. Therefore, all samples were analysed as single measurements. To ensure representativeness, soils were finely ground and homogenized prior to subsampling. The relatively small sample mass (50 mg) is sufficient for quantitative dissolution while enabling efficient high-throughput processing.

In the same water extracts for gypsum determination concentrations of other cations and anions were analysed by the same ion chromatography methods, such as Na^+ , Mg^{2+} , and NH_4^+ , and NO_3^- , Cl^- and PO_4^{3-} .

2.3. Calculation of stocks and of biological source of soil inorganic carbon (SIC)

Stocks of SIC (kg m^{-2}) were calculated by multiplying the SIC concentration (g C kg^{-1}) of each horizon by its bulk density (BD; g cm^{-3}) and thickness (m) and then summing across all horizons. Given the sandy nature of the oasis soils, the coarse fragment fraction ($>2\text{ mm}$) was found to be negligible, representing less than 2% of the total soil volume; therefore, the volumetric impact on stock estimation was minimal. We estimated bulk density according of the predictive model from (Brahim et al., 2014) where:

For topsoils (0–30 cm): $\text{BD} = 0.9 (\pm 0.1) - 0.08 (\pm 0.01) * \text{OC} + 0.007 (\pm 0.001) * \text{F-Sand} + 0.007 (\pm 0.002) * \text{F-Silt} + 0.05 (\pm 0.01) * \text{pH}$. ($R^2 = 0.58$, $\text{SE} = 0.14$).

For deep soil horizons ($> 30\text{ cm}$): $\text{Db} = 1.90 (\pm 0.02) - 0.08 (\pm 0.03) * \text{OC} - 0.0031 (\pm 0.0009) * \text{Clay} - 0.0023 (\pm 0.0007) * \text{CaCO}_3$. ($R^2 = 0.30$, $\text{SE} = 0.14$).

where OC represents soil organic carbon content (%), F-Sand and F-Silt the absolute fractions of sand and silt, Clay the clay content (%), and CaCO_3 the calcium carbonate content (%).

The suitability of using model derived BD for SIC stock estimation was evaluated using direct BD measurements from (8) representatives soil pits. These validation pits covered both under tree and between tree positions, resulting in (16) validation observations (8 pits * 2 position). For these 16 observations, SIC stocks were calculated twice: one using measured BD and once using model predicted BD. The relationship between the two stock estimates was highly significant ($n = 16$ profiles, slope 0.85, $R^2 = 0.96$, $p < 0.0001$; Fig. S2), indicating that the use of model-derived BD did not substantially alter SIC stock estimates.

Moreover, we applied C isotope analysis to better understand the origin of SIC. In former studies pedogenic carbonates were assumed to exclusively derive from soil respiration (Ramnarine et al., 2012) and therefore to have a biological C isotope signature. The $\delta^{13}\text{C}$ signatures of biological CO_2 sources range from around -25 to -28‰ (Werth and Kuzyakov, 2010). This is when there is little diffusional entry of atmospheric CO_2 or downward leaching of dissolved inorganic C (DIC) deriving from it into the soil horizons where the carbonate formation happens. In desert ecosystems, pedogenic carbonates may originate from both soil-respired CO_2 ($\delta^{13}\text{C} \approx -28\text{‰}$) and atmospheric CO_2 ($\delta^{13}\text{C} \approx -8$ to -6‰). Because lithogenic carbonates and pedogenic carbonates formed from atmospheric CO_2 share similar isotopic signatures, $\delta^{13}\text{C}$ values alone cannot reliably distinguish between them. Consequently, only pedogenic carbonates derived from soil-respired (biogenic) CO_2 can be differentiated from lithogenic and atmospherically derived pedogenic carbonates based on $\delta^{13}\text{C}$. However, to link C isotopes between source CO_2 and the pedogenic carbonates formed, diffusional isotope fractionation in soils ($\Delta_{\text{diffusion}}$, ^{13}C enrichment by $+4.4\text{‰}$ at CO_2 sink location (Cerling, 1984)) and the equilibrium isotope fractionation of carbonate formation ($\Delta_{\text{CO}_2\text{-CaCO}_3}$, ^{13}C enrichment by 10.3‰ at $20\text{ }^{\circ}\text{C}$

for $\text{CO}_2\text{-CaCO}_3$) need to be accounted for (Nordt et al., 1998).

$$\text{Pedogenic } \delta^{13}\text{C}_{\text{SIC}} = \delta^{13}\text{C}_{\text{soilCO}_2} + \Delta_{\text{diffusion}} + \Delta_{\text{CO}_2\text{-CaCO}_3}$$

The reference temperature for isotope fractionation ($20\text{ }^{\circ}\text{C}$) is close to the mean annual temperature of the sites ($\sim 22\text{ }^{\circ}\text{C}$) in our studied soils at $>30\text{ cm}$ depth where carbonate formation occurs. The isotope signature of non-biological formed lithogenic carbonate (δ_A) was determined directly from our dataset. We selected the 21 samples with the highest $\delta^{13}\text{C}_{\text{SIC}}$ values, which are interpreted as the most representative of the primary lithogenic source, and calculated their median, yielding a δ_A value of $+2.8\text{‰}$ for the lithogenic endmember. This value is consistent with the established global range of $\delta^{13}\text{C}$ for unaltered marine carbonates, which served as a validation benchmark (e.g., $+2.0$ to $+4.5\text{‰}$ in modern marine cements (Oehlert and Swart, 2014); $+0.5$ to $+2.3\text{‰}$ in Quaternary deposits (Candy et al., 2012); $+2.0$ to $+3.5\text{‰}$ in Upper Cenomanian sequences (Kuhnt et al., 2017); and $+1.7$ to $+4.6\text{‰}$ in Upper Cretaceous Tethyan realms (Matsumoto et al., 2020)). The relationship between the contribution of SIC to STC (x-axis) and $\delta^{13}\text{C}$ of STC (Fig. S6) allowed extrapolation to 0%, which represents the C isotope signature of SOC (0% of STC present as SIC). The regression intercept on the y-axis produced a $\delta^{13}\text{C}_{\text{SOC}}$ of $-24.8 \pm 0.9\text{‰}$, which is close to the median directly measured $\delta^{13}\text{C}_{\text{SOC}}$ ($-24.6 \pm 0.09\text{‰}$). Using the regression-derived $\delta^{13}\text{C}_{\text{SOC}}$ value and accounting for isotope fractionation by adding 14.7‰ , we derived a biological endmember δ_B of -10.1‰ . The actually measured SIC isotope signatures (δ_M , $\delta^{13}\text{C}_{\text{SIC}}$) were assumed as the mixture between δ_A and δ_B , and the fractions f_A and f_B finally were calculated as follows:

$$f_B = \frac{\delta_M - \delta_A}{\delta_B - \delta_A}$$

2.4. Statistical analysis

Data were checked for outliers, normal distribution and variance homogeneity before further statistical analyses. Where necessary data were log (LN) transformed to conform to the principles of analysis of variance and for correlations (normal distribution of residuals). To account for the hierarchical data structure and repeated measures, we employed a systematic computational workflow (Fig. S3). We used Linear Mixed-Effects Models (LME) using the R package 'nlme' (Pinheiro et al., 1999), with an autoregressive (AR (Álvarez et al., 2022)) covariance structure to assess the effects of oasis management, topography, tree position, and soil depth on SIC, SIC contribution to STC ($\text{SIC}\% \text{TC}$), $\delta^{13}\text{C}_{\text{SIC}}$, and the additional soil physicochemical parameters detailed in Supplementary (Table S2). In these models, 'Plot' (Profile) nested within 'Oasis' was included as a random effect. For each response variable, three LME residual structures were compared: an independent residual model, a continuous autoregressive correlation structure along soil depth [CAR (1)], and a CAR(1) model combined with a varident variance structure allowing depth-specific residual variances. Model fit was evaluated using Akaike Information Criterion (AIC), and the final model was selected based on the lowest AIC. The adequacy of the selected LME was assessed using diagnostic checks on normalised residuals. Residual normality was evaluated using quantile–quantile (QQ) plots and Shapiro–Wilk tests, while homoscedasticity was assessed using residuals-versus-fitted plots, scale–location plots, and Brown–Forsythe tests across fixed-effect groups. Figures were created using the 'ggplot2' package (Wickham, 2016). Box plots were used to visualize the distribution of soil parameters across topographic positions, oasis types, soil depth, and tree position, whereas significant differences shown in the figures were derived from pairwise comparisons of estimated marginal means from the final selected linear mixed-effects models using the R package 'emmeans' (Lenth and Piskowski, 2017). Correlation heatmaps based on Pearson's correlations were used to display the relationships between soil parameters using the R package 'corrplot' (Wei and Simko, 2010). We applied principal component analysis (PCA) as multivariate statistical approach to summarize the variance in soil physicochemical

properties and to reveal underlying associations between SIC and related variables such as SOC, texture, gypsum, and soil pH. Variables were pre-selected based on conceptual relevance and to reduce redundancy among strongly collinear properties. Soil depth was included as a numerical variable using depth-layer midpoints. PCA was performed in R using `prcomp()` on complete cases after centring and scaling all variables to unit variance, equivalent to using the correlation matrix rather than the raw covariance matrix (R Core Team, 2025). PCA loadings were calculated as correlations between standardized variables and principal component scores. Because PC3 explained substantially less variance than PC1 and PC2, it was retained in Table S3 for transparency but not used as a primary axis for mechanistic interpretation. To clarify variable groupings, we additionally performed varimax-rotated factor analysis

using the same standardized variables with the R package 'psych' (Revelle, 2026).

3. Results

3.1. Effects of spatial and anthropogenic factors on soil inorganic carbon (SIC)

SIC contents ranged from 0 to 3.8% (dry mass), with mean values of 1.6% in modern oases and 1.5% in traditional oasis systems. According to the linear mixed-effects model (LME), soil depth significantly affected SIC content ($F_{5,155} = 4.71$, $p < 0.001$), whereas the main effect of oasis management was not significant. However, the interaction between

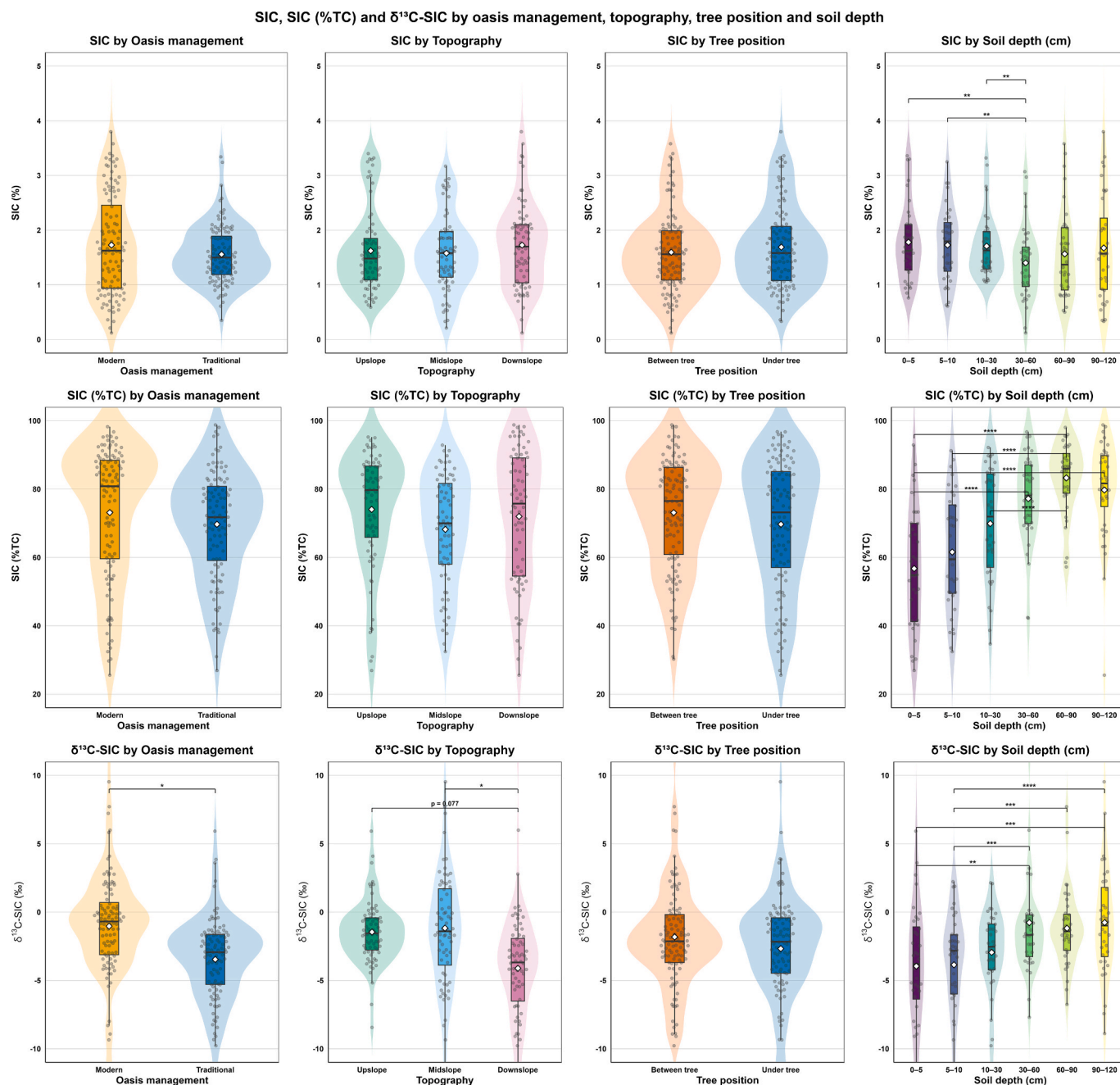


Fig. 2. Boxplots illustrating the statistical significance of environmental effects (oasis management, topography, tree position, and soil depth) on soil inorganic carbon (SIC) content, the contribution of SIC to soil total carbon (STC; SIC%STC), and the carbon isotope composition of SIC ($\delta^{13}\text{C}$ SIC). The plots display the median (line within the box), the interquartile range between the 25th and 75th percentiles (box), and whiskers extending to the most extreme values within 1.5 times the interquartile range. Values beyond this range are shown as outliers. Asterisks denote significant differences between groups.

oasis management and topography was significant ($F_{2,8} = 4.65$, $p = 0.04$), as was the interaction between topography and soil depth ($F_{10,155} = 2.17$, $p = 0.02$), indicating that the influence of topographic position on SIC content varies between oasis systems and across the soil profile (Fig. 2, Table 2).

The contribution of SIC to total soil carbon (SIC%STC) averaged 82% in modern oases and 72% in traditional systems. Soil depth had a strong effect on SIC%STC ($F_{5,155} = 21.34$, $p < 0.001$), with values increasing from approximately 54% in surface layers to 85% at depth (Fig. 2, Table 2).

SIC stocks (0–100 cm) ranged from 9.0 to 51.9 kg m⁻², with a mean of 25.3 kg m⁻². Extending the depth interval to 0–120 cm increased the mean stock to 31.3 kg m⁻² and the range to 11.3–62.4 kg m⁻², corresponding to a 24% increase in mean SIC stocks. None of the main factors (oasis management, topography, or tree position) had a significant effect on SIC stocks.

The carbon isotope composition of SIC ($\delta^{13}\text{C-SIC}$) varied significantly across fixed effects. Oasis management had a significant effect ($F_{1,4} = 10.84$, $p = 0.03$), with more negative values in traditional oases (−3.5‰) compared to modern systems (−1‰; Fig. 2). Topography also influenced $\delta^{13}\text{C-SIC}$ ($F_{2,8} = 5.40$, $p = 0.03$), with more negative values in downslope positions (−3.7‰) than in midslope and upslope positions (−1.6‰). Soil depth had a strong effect ($F_{5,155} = 6.09$, $p < 0.001$), with $\delta^{13}\text{C-SIC}$ values becoming less negative with increasing depth (from −3.9‰ in topsoil to −1.4‰ at 90 cm). Differences between tree positions were minor and not statistically significant.

In modern oases, SIC content decreased from upslope positions (mean $\pm$ SE $1.8 \pm 0.15\%$) to downslope areas ($1.4 \pm 0.16\%$), whereas

traditional oases showed the opposite pattern, with SIC increasing from upslope ($1.2 \pm 0.05\%$) to downslope positions ($1.8 \pm 0.1\%$).

Soil pH ranged from 7.4 to 8.4 and showed a significant interaction effect between topography and oasis management (Fig. 3, Table 2, Table S2). In modern oases, pH was highest at upslope positions (mean 8.2) and decreased towards downslope areas (7.8), whereas in traditional oases pH increased slightly from upslope (7.7) to downslope positions (7.9). Consequently, differences between oasis types were most pronounced at upslope positions and diminished along the catena.

In contrast, gypsum content and $\delta^{13}\text{C-SIC}$ did not follow this inverse pattern between oasis types (Fig. 4, Table S2). Gypsum concentrations were substantially higher in traditional oases and increased towards downslope positions (>600 mg g⁻¹), whereas modern oases showed consistently low gypsum contents (<100 mg g⁻¹) across the topographic gradient. The $\delta^{13}\text{C-SIC}$ values became more negative from upslope (0 to −2‰) to downslope positions (−5‰ to −9‰) in both oasis types, with overall more negative signatures in traditional systems.

3.2. Relationships between soil inorganic carbon (SIC) and potential physicochemical drivers

Principal component analysis (PCA; Fig. 4, Table S3) was conducted using 15 soil variables, with highly collinear variables excluded. The first three principal components (PC1–PC3) together explained 53% of the total variance. PC1 reflected depth-related gradients, with negative loadings for SIC%STC (−0.75), $\delta^{13}\text{C-SIC}$ (−0.60), soil depth (−0.57), and pH (−0.40), and positive loadings for SOC (0.86), PO_4^{3-} (0.81), NH_4^+ (0.65), and NO_3^- (0.58). PC2 indicated an association of SIC content with salinity and gypsum-related variables, with strong positive loadings for Na^+ (0.92), Cl^- (0.73), Mg^{2+} (0.58), gypsum (0.49), and SIC (0.40). PC3 explained substantially less variance than PC1 and PC2, and was retained in Table S3 for transparency but was not used as a primary axis for mechanistic interpretation.

To clarify variable groupings and avoid overinterpretation of lower-order PCA axes, we additionally applied varimax-rotated factor analysis (Fig. 4C). The rotated factor structure provided a clearer separation of controls. MR1 was mainly associated with SOC and nutrient availability, with strong positive loadings for SOC, PO_4^{3-} , NH_4^+ , and NO_3^- , and negative loadings for SIC%STC, $\delta^{13}\text{C-SIC}$, and soil depth. MR2 represented a salinity-related factor, with strong loadings for Na^+ and Cl^- . MR3 was mainly associated with gypsum and soil depth. Importantly, SIC showed only weak loading on MR2 and negligible loading on MR3, whereas gypsum loaded strongly on MR3. This indicates that SIC and gypsum were not strongly coupled but were structured by partly different environmental controls.

Relationships identified in the PCA were further examined using Pearson correlations (Fig. S5). Overall, correlation coefficients for SIC were low ($R < 0.25$). SIC content nevertheless showed weak but significant positive correlations with soil pH, salinity (Cl^- , Na^+), Mg^{2+} , and sand content, and negative correlations with silt, whereas no significant relationships were observed with gypsum, clay, or SOC.

The contribution of SIC to total soil carbon (SIC%STC) was positively correlated with SIC content ($R = 0.42$), soil pH (0.30), gypsum (0.13), and salinity (0.14), and negatively correlated with SOC (−0.80) and available nutrients (NO_3^- , PO_4^{3-} , NH_4^+ ; $R = -0.48$ to -0.25). SIC%STC showed no relationship with soil texture.

The $\delta^{13}\text{C-SIC}$ values were positively correlated with SIC%STC ($R = 0.27$) and soil pH (0.21), and negatively correlated with SOC (−0.41), nutrients (NO_3^- , SO_4^{2-} , PO_4^{3-} , NH_4^+ , Mg^{2+} , Ca^{2+} ; $R = -0.44$ to -0.18), salinity (−0.26), gypsum (−0.19), and clay (−0.18). The negative relationship between SOC and $\delta^{13}\text{C-SIC}$ (Fig. S5) was consistent across oasis types, with a steeper slope in modern systems (−0.47) than in traditional systems (−0.34).

Table 2

Linear Mixed-Effects Model (LME) analysis incorporating a first-order autoregressive (AR) covariance structure to assess the main effects of oasis management, topography, tree position and soil depth as well as their interactions on soil inorganic C (SIC) content, SIC contribution to soil total C (STC; SIC%STC), the stable C isotope composition of SIC ($\delta^{13}\text{C-SIC}$) and the biological source contribution to SIC. The model accounts for the hierarchical data structure by treating plots nested within oasis systems as random effects. P values are provided, with significance levels of ***, $p < 0.001$, **, $p < 0.01$, *, $p < 0.05$, and †, $p < 0.1$ (trend).

Effect	SIC	SIC%STC	$\delta^{13}\text{C-SIC}$	Biological source
Oasis management (Om)	$F(1,4) = 0.08$, $p = 0.78$	$F(1,4) = 0.38$, $p = 0.57$	$F(1,4) = 10.84$, $p = 0.03$ *	$F(1,4) = 7.10$, $p = 0.05$ †
Topography (T)	$F(2,8) = 0.17$, $p = 0.84$	$F(2,8) = 2.37$, $p = 0.15$	$F(2,8) = 5.40$, $p = 0.03$ *	$F(2,8) = 5.80$, $p = 0.02$ *
Tree position (TP)	$F(1,14) = 2.92$, $p = 0.10$	$F(1,14) = 1.15$, $p = 0.30$	$F(1,14) = 1.44$, $p = 0.24$	$F(1,14) = 3.30$, $p = 0.091$ †
Soil depth (SD)	$F(5,155) = 4.71$, $p < 0.001$ ***	$F(5,155) = 21.34$, $p < 0.001$ ***	$F(5,155) = 6.09$, $p < 0.001$ ***	$F(5,155) = 6.63$, $p < 0.001$ ***
Om $\times$ T	$F(2,8) = 4.76$, $p = 0.04$ *	$F(2,8) = 0.67$, $p = 0.54$	$F(2,8) = 1.07$, $p = 0.38$	$F(2,8) = 0.80$, $p = 0.48$
Om $\times$ TP	$F(1,14) = 1.41$, $p = 0.25$	$F(1,14) = 0.06$, $p = 0.80$	$F(1,14) = 0.24$, $p = 0.63$	$F(1,14) = 0.28$, $p = 0.60$
Om $\times$ SD	$F(5,155) = 1.78$, $p = 0.11$	$F(5,155) = 1.22$, $p = 0.30$	$F(5,155) = 1.62$, $p = 0.15$	$F(5,155) = 0.74$, $p = 0.59$
T $\times$ TP	$F(2,14) = 1.64$, $p = 0.22$	$F(2,14) = 0.79$, $p = 0.47$	$F(2,14) = 0.79$, $p = 0.47$	$F(2,14) = 1.95$, $p = 0.17$
T $\times$ SD	$F(10,155) = 2.17$, $p = 0.02$ *	$F(10,155) = 0.54$, $p = 0.86$	$F(10,155) = 3.45$, $p < 0.001$ ***	$F(10,155) = 3.56$, $p < 0.001$ ***
TP $\times$ SD	$F(5,155) = 1.43$, $p = 0.21$	$F(5,155) = 0.40$, $p = 0.84$	$F(5,155) = 1.49$, $p = 0.19$	$F(5,155) = 0.66$, $p = 0.65$

Soil pH, gypsum, SIC and $\delta^{13}\text{C}$ -SIC along topographic positions

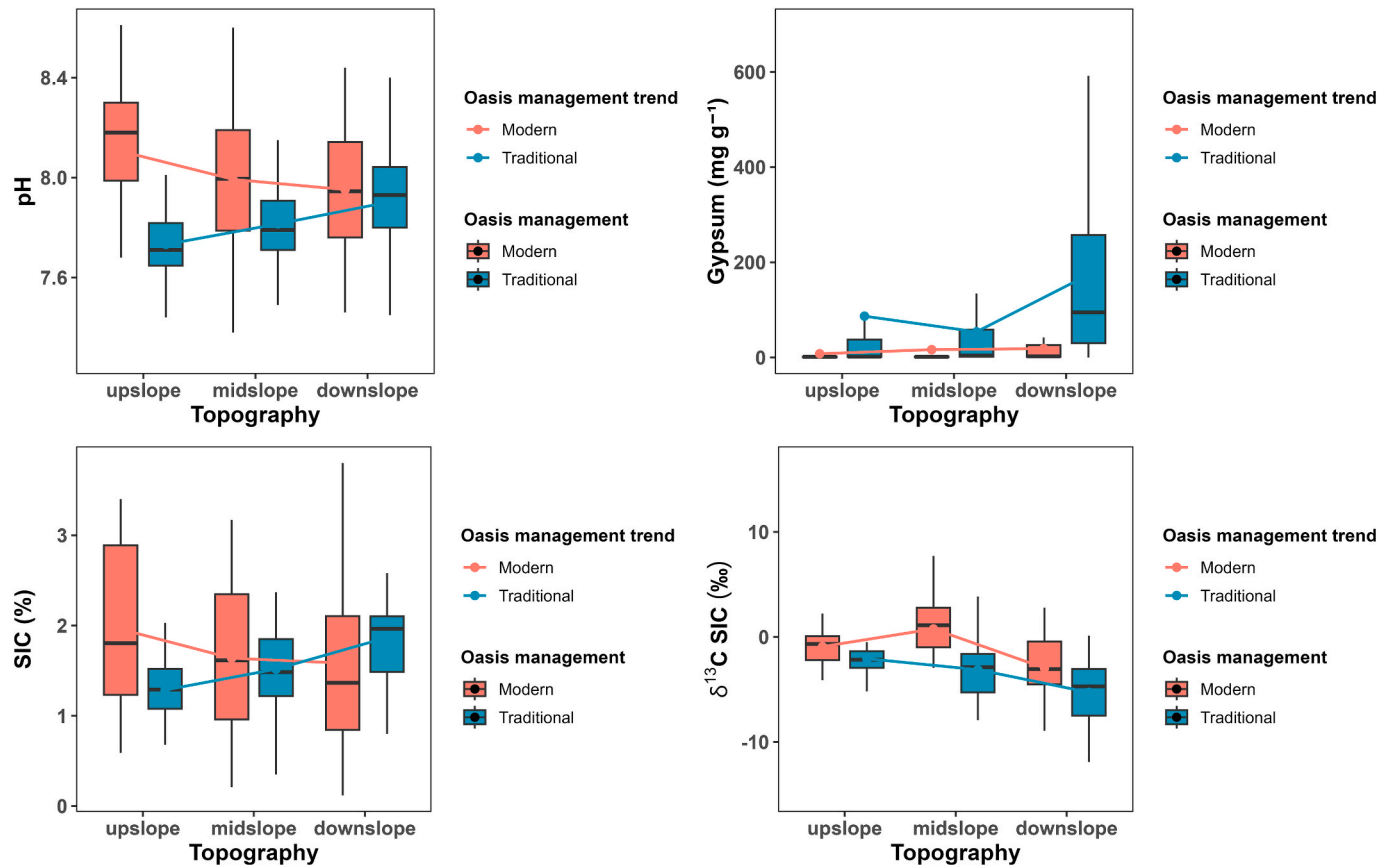


Fig. 3. Changes in soil pH, gypsum content, soil inorganic carbon (SIC) content, and carbon isotope composition of SIC ($\delta^{13}\text{C}$ SIC), with topographic position in traditional and modern oasis systems. Statistics of main effects of topography and oasis management and their interaction can be found in Table 2 and in Supplementary Table S2.

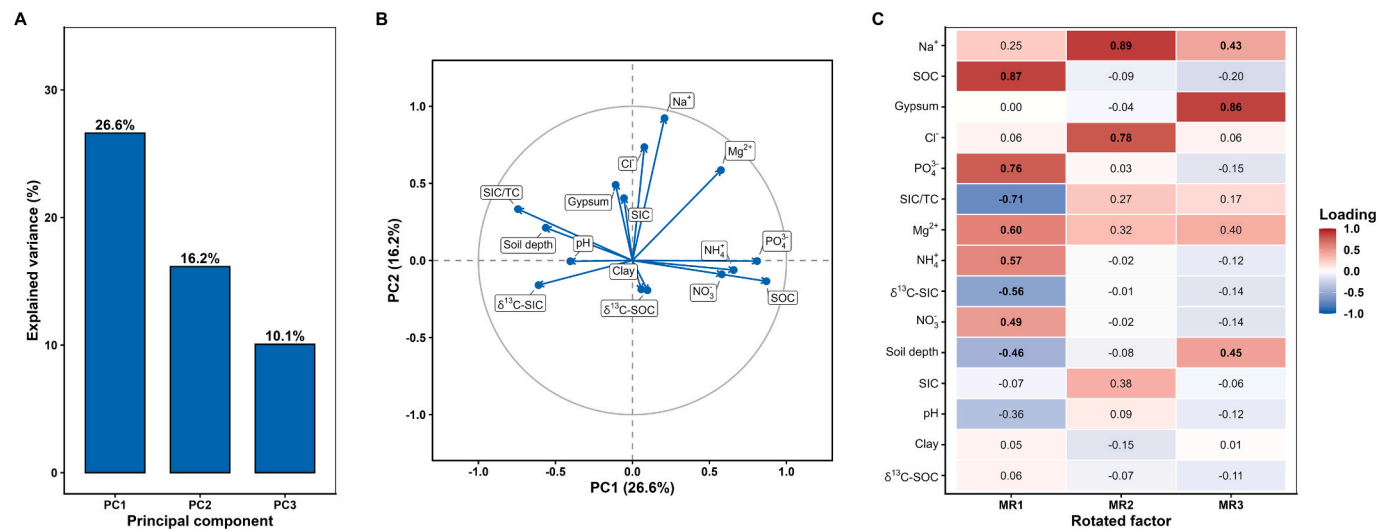


Fig. 4. Principal component analysis (PCA) of the multivariate relationship between soil inorganic carbon (SIC), its contribution to soil total carbon (SIC%STC), and its carbon isotope composition ($\delta^{13}\text{C}$ SIC) and potential proximal soil physicochemical drivers including soil pH, soil depth, soil organic carbon (SOC), gypsum, soil cations and anions, and soil texture. (A) Scree plot illustrating the percentage of explained variance across the principal components. (B) Correlation circle for PC1 and PC2. (C) Varimax-rotated factor-loading heatmap for the first three rotated factors. Data derived from modern and traditional oasis types in the Djerid region, Tunisia, sampled across topography, tree position and soil depth.

3.3. Contribution of biological CO₂ sources to soil inorganic carbon (SIC) formation

The relative contribution of pedogenic carbonates derived from soil-respired (biogenic) CO₂ varied significantly with soil depth, topography, and oasis management (Fig. 5). Topsoils (0–5 cm) showed the highest contributions (53% ± 2%), which decreased markedly below 30 cm (<30%).

Tree proximity (under-tree vs between-tree positions) had no significant effect on the relative contribution of biogenic CO₂. In contrast, oasis management influenced these patterns, with higher contributions in traditional systems (40%) compared to modern systems (27%). Topography also affected the distribution, with higher contributions in downslope positions than in upslope positions.

4. Discussion

4.1. Soil inorganic carbon (SIC) dominates soil carbon storage in oasis systems

Soil inorganic carbon (SIC) accounted for 76% of total soil carbon (STC) in the investigated oasis agroecosystems, confirming its dominant role in soil C storage under hyper-arid conditions. This is consistent with global patterns, where SIC contributions increase with aridity and can

exceed SOC in dryland soils (Zeng et al., 2023), partly because carbonates are relatively soluble minerals that can dissolve, be redistributed as dissolved Ca²⁺ and inorganic carbon, and reprecipitate under evaporative conditions. In hyper-arid environments such as the Sahara, the Alxa region in China, or the Atacama Desert, SIC commonly represents the majority of soil C, with reported contributions ranging from ~90% to >97% depending on climatic severity and vegetation cover (Valdivia-Silva et al., 2012; Zhu et al., 2024).

The dominance of SIC in these systems is primarily controlled by climate, parent material, and soil chemical conditions. In drylands, low precipitation relative to evapotranspiration limits leaching and promotes evaporative concentration of Ca²⁺ and alkalinity, favoring carbonate precipitation and retention. These processes are further regulated by soil pH, with alkaline conditions stabilizing carbonates and constraining dissolution; conversely, carbonate equilibria also buffer soil pH by neutralizing acidity through carbonate dissolution-precipitation reactions (Slessarev et al., 2016). As a result, SIC accumulation is strongly coupled to aridity gradients and becomes increasingly important as water availability declines (Plaza et al., 2018; Huang and Wei, 2025; Liao et al., 2026).

SIC stocks (0–1 m) in this study ranged from 9.0 to 52.0 kg C m⁻², with a mean of 25.4 kg C m⁻², which is high compared to unmanaged desert soils in China (13 ± 1.6 kg C m⁻²; Dong et al., 2024) and comparable to other Tunisian oasis systems (18–23 kg C m⁻²; Brahim et al.,

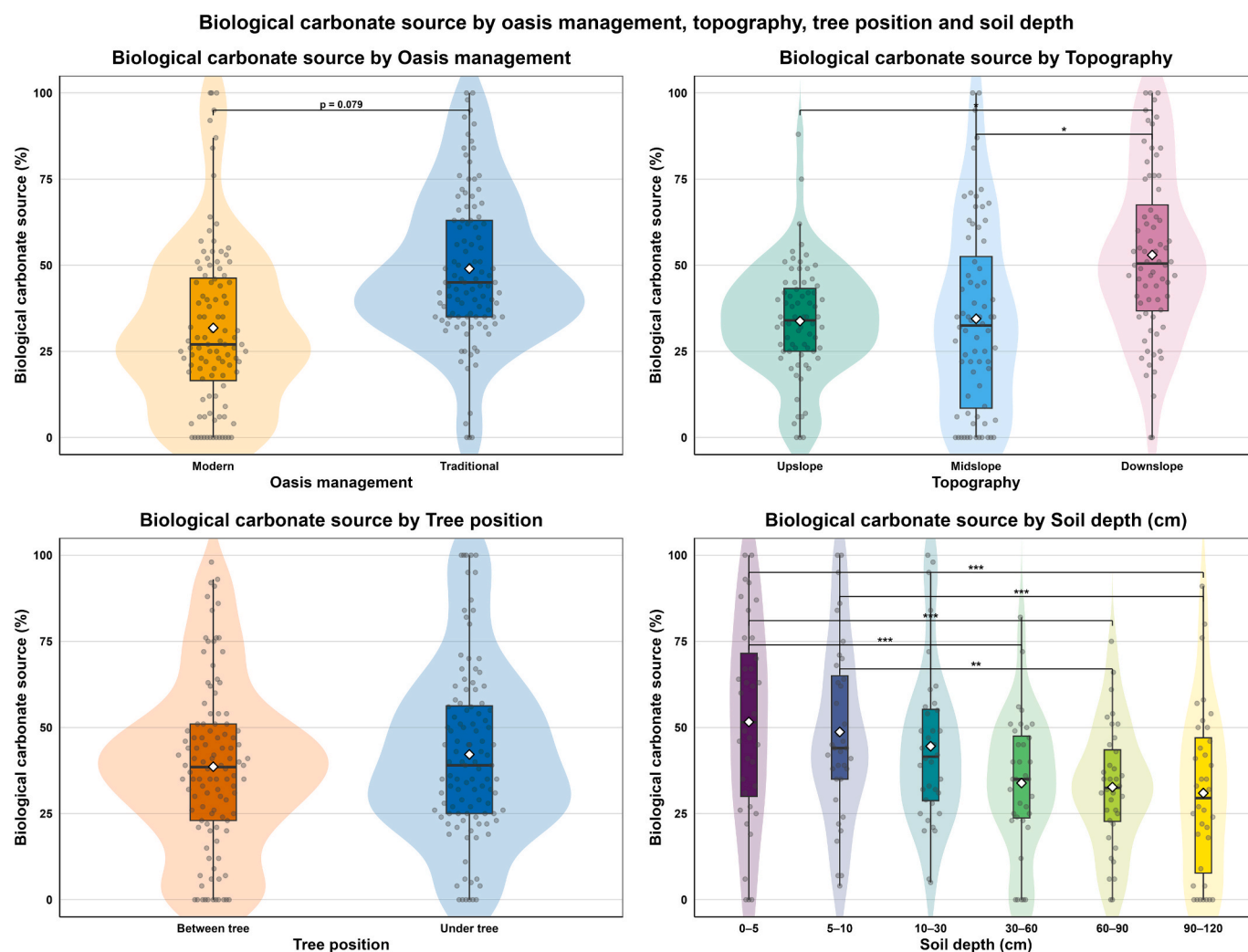


Fig. 5. Boxplots of the effects of oasis management, topography, tree position and soil depth on biologically derived SIC (%) in soils in the Djerid region, Tunisia. The partitioning of SIC was performed using differences in carbon isotope signatures ($\delta^{13}\text{C}$) between (i) ^{13}C depleted biologically sourced pedogenic carbonates deriving from decomposition of soil organic carbon and root respiration, and (ii) ^{13}C enriched lithogenic carbonates and pedogenic carbonates deriving from atmospheric CO₂.

2014). Importantly, SIC stocks were not significantly affected by oasis management or topography in our study, indicating that the overall magnitude of SIC storage is primarily constrained by geological and climatic factors rather than by contemporary management. The high SIC stocks are therefore likely dominated by lithogenic carbonates derived from calcareous parent material and aeolian inputs, with pedogenic contributions superimposed on this large background pool. Similar controls have been reported in other dryland systems, where parent material exerts a strong influence on SIC storage (Zeng et al., 2023; Dong et al., 2024).

In contrast to SIC stocks, the relative contribution of SIC to STC increased with soil depth, consistent with global observations (Huang et al., 2024). This vertical pattern reflects the contrasting origins of the two carbon pools: SOC is primarily biogenic and concentrated near surface organic inputs, whereas SIC is largely geogenic, associated with carbonate-rich parent material at depth. Our results show that soil depth is significantly structured SIC-related variables, but do not support a uniform increase of SIC concentrations with depth. Instead, depth-dependent patterns likely reflect the vertical distribution of lithogenic carbonates and the limited vertical redistribution of dissolved carbonate species under hyper-arid conditions. While processes such as carbonate dissolution in topsoil and re-precipitation at depth have been proposed for other systems (Mi et al., 2008), our data do not directly resolve these mechanisms and suggest that large, relatively stable carbonate pools dominate the soil profile.

4.2. Minor negative effects of agricultural management on soil inorganic carbon (SIC)

Although traditional and modern oases differed strongly in management history, with traditional systems representing centuries-old agroecosystems and modern systems reflecting more recent and intensive agricultural management, SIC content, SIC%STC, and SIC stocks were not significantly affected by oasis management in our study ($p = 0.78$ and $p = 0.57$ for SIC and SIC%STC, respectively), nor by tree proximity ($p = 0.30$). This indicates that contemporary management practices have not induced detectable changes in the overall magnitude of SIC pools. Consequently, our first hypothesis (H1), which proposed enhanced SIC loss under intensive management due to fertilisation-induced acidification, is not supported by our data.

This finding contrasts with previous studies and meta-analyses reporting SIC depletion under N fertilisation due to soil acidification and enhanced carbonate dissolution (Liao et al., 2026; Raza et al., 2020; Zamanian et al., 2018). However, the absence of a detectable management effect in our system suggests that these processes may be buffered under hyper-arid conditions. One likely explanation is the role of irrigation water derived from the Complex Terminal aquifer, which supplies Ca^{2+} and alkalinity (Table S4). Under conditions where evapotranspiration exceeds infiltration, such inputs can promote carbonate precipitation and partially offset acidification-driven dissolution (Dang et al., 2022; Ortiz et al., 2022; Tang et al., 2024).

Consistent with this interpretation, soil pH remained within a narrow circumneutral to slightly alkaline range (7.3–8.6) and was not significantly correlated with SIC content ($r < 0.2$, $p > 0.05$), indicating that acidification effects were limited at the system scale. Instead, SIC dynamics appear to be dominated by large, geogenic carbonate pools, with pedogenic processes superimposed but not sufficient to alter bulk SIC stocks.

While bulk SIC remained unchanged, isotopic evidence ($\delta^{13}\text{C}$ -SIC) indicates that management influences carbonate sources and formation pathways. The more negative $\delta^{13}\text{C}$ -SIC values observed in traditional systems suggest a greater contribution of soil-respired CO_2 to carbonate formation, consistent with higher biological activity and organic inputs. These results therefore point to shifts in carbonate cycling processes rather than to net gains or losses of SIC.

Overall, our findings suggest that in hyper-arid oasis systems,

agricultural management modifies SIC dynamics primarily at the level of carbonate transformation and source partitioning, whereas the total SIC pool remains buffered by strong geological and hydrological controls.

4.3. Topography-management interactions impact soil inorganic carbon (SIC) distribution

In contrast to our second hypothesis (H2), topographic position did not exert a consistent main effect on SIC content across the study area ($p = 0.84$), nor did it lead to a uniform downslope accumulation of carbonates. Instead, SIC patterns were governed by a significant interaction between topography and management, as well as by strong depth-dependent structuring (topography $\times$ depth, $p < 0.001$). This indicates that topography influences SIC distribution primarily through context-dependent patterns rather than through a uniform directional trend.

Specifically, opposing SIC patterns along the catena were observed between oasis types: in modern oases, SIC decreased from upslope to downslope positions, whereas in traditional oases the inverse pattern occurred. These contrasting trends likely explain the absence of a significant overall topographic effect. Rather than reflecting net gains or losses of SIC, these patterns point to spatial reorganization of carbonates within the landscape.

Topographic controls on SIC are generally mediated through water redistribution, evaporation, and associated solute transport in drylands (Obear and Soldat, 2016; Brahim et al., 2021; Wang et al., 2023). In more heterogeneous landscapes, slope position has been shown to strongly influence carbonate accumulation and redistribution (Díaz-Hernández et al., 2013; Gozukara et al., 2025). In our system, the differing responses between oasis types suggest that management history modulates how these processes operate along the catena. In modern systems, higher SIC at upslope positions may reflect limited leaching and localized carbonate precipitation (Xie et al., 2023; Dubeux et al., 2024; Su et al., 2025), whereas in traditional systems, long-term irrigation may enhance downslope accumulation of dissolved inorganic carbon followed by reprecipitation under evaporative conditions (Obear and Soldat, 2016; De Soto et al., 2019; Wang et al., 2023). However, these mechanisms cannot be directly resolved from our data and should therefore be interpreted cautiously.

Overall, our results highlight that SIC distribution in oasis systems emerges from the interaction between topographic gradients and land-use history, rather than from topography or management alone. This supports previous work emphasizing that carbonate dynamics in drylands are shaped by coupled hydrological and anthropogenic controls, which are often overlooked when these factors are considered in isolation (Díaz-Hernández, 2010; Shi et al., 2012).

4.4. Interrelationships between gypsum and soil inorganic carbon (SIC)

Gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) is a major mineral component in the study area and can influence SIC dynamics through shared geochemical controls and coupled dissolution–precipitation processes. However, the apparent associations between SIC and gypsum observed across PCA axes are interpreted here cautiously, because lower-order components explained progressively less variance and may partly reflect statistical rather than mechanistic contrasts. Varimax-rotated factor analysis provided a clearer interpretation: gypsum loaded strongly on MR3, whereas SIC showed only weak loading on MR2 and negligible loading on MR3, indicating that SIC and gypsum were not strongly coupled in the multivariate structure.

Gypsum accumulation is consistent with evaporitic conditions characterized by high salinity, strong evaporative concentration, and particularly pronounced near hypersaline systems such as the Chott el Jerid (Hardie, 1991; Warren, 2010; Brahim et al., 2021). Under such evaporitic conditions, carbonate precipitation generally precedes gypsum formation, reflecting progressive concentration of solutes.

Although gypsum dissolution during irrigation or episodic wetting may release Ca^{2+} that could subsequently contribute to pedogenic carbonate precipitation during drying phases, our data do not resolve this mechanism. Due to its higher solubility, gypsum can dissolve following irrigation or episodic wetting, releasing Ca^{2+} that may subsequently contribute to pedogenic carbonate precipitation under drying conditions (Herrero and Porta, 2000; Wong et al., 2009; Laudicina et al., 2013; Peng et al., 2024). With long-term agricultural management, fertilizer-induced acidification and hydrologically mediated Ca^{2+} losses, gypsum dissolution may replenish Ca^{2+} for carbonate precipitation and buffering SIC. Nevertheless, this remains a plausible mechanism rather than a directly demonstrated pathway in our dataset.

Overall, the revised multivariate interpretation indicates that gypsum and SIC are not directly or strongly coupled, but are structured by partly different environmental controls within a broader evaporitic geochemical context. While these patterns are consistent with coupled dissolution–reprecipitation processes, the underlying mechanisms cannot be directly resolved from our data.

4.5. Sources of soil inorganic carbon (SIC)

The negative relationship between $\delta^{13}\text{C}$ -SIC and SOC across all samples indicates coupling between organic and inorganic carbon pools. More negative $\delta^{13}\text{C}$ -SIC values in soils with higher SOC suggest a greater incorporation of isotopically depleted, soil-respired CO_2 into pedogenic carbonates, consistent with previous studies linking carbonate formation to biological CO_2 inputs (Cerling, 1984; Kuzyakov and Bol, 2005; Tang et al., 2024; Liao et al., 2026).

Using a two-source mixing approach, we estimate that biologically derived CO_2 contributes approximately 27–40% of SIC, with higher contributions in traditional oasis systems, downslope positions, and topsoils. These patterns align with higher SOC contents and biological activity in traditional systems, where organic inputs and long-term management enhance soil respiration and CO_2 production (Raddadi

and Podwojewski, 2024). The remaining SIC fraction reflects lithogenic carbonates and pedogenic carbonates formed from atmospheric CO_2 , both of which exhibit relatively enriched $\delta^{13}\text{C}$ signatures (Cerling, 1984; Bayat et al., 2023).

Depth-dependent trends further support this interpretation. Higher relative contributions of biogenic CO_2 in topsoils correspond to elevated SOC, nutrient availability, and biological activity; in contrast, deeper soil layers are increasingly dominated by lithogenic carbonates, consistent with observations from other dryland systems (Dong et al., 2024).

While these isotopic patterns indicate shifts in carbonate sources and formation pathways, they do not provide direct evidence for net SIC loss or a complete transformation of lithogenic into pedogenic carbonates. Thus, our third hypothesis (H3) is only partially supported: management and environmental gradients influence the relative contribution of biogenic CO_2 to carbonate formation, but the overall SIC pool remains dominated by lithogenic carbonates.

Overall, our results highlight that SIC dynamics in oasis systems are characterized by active exchange between organic and inorganic carbon pools, with biologically mediated carbonate formation superimposed on a large and relatively stable geogenic carbonate reservoir. These coupled dynamics, modulated by topographic position and oasis management history, are synthesized in a conceptual model presented in Fig. 6.

5. Conclusions

Soil inorganic carbon (SIC) dominates soil carbon storage in Tunisian oasis agroecosystems, accounting for the majority of total soil carbon under hyper-arid conditions. Despite contrasting management practices, SIC contents and stocks were not significantly affected by agricultural management, indicating that large lithogenic carbonate pools buffer the system against short- to medium-term anthropogenic impacts. Instead, SIC dynamics were primarily expressed through changes in spatial distribution and carbonate sources. Topography–management interactions

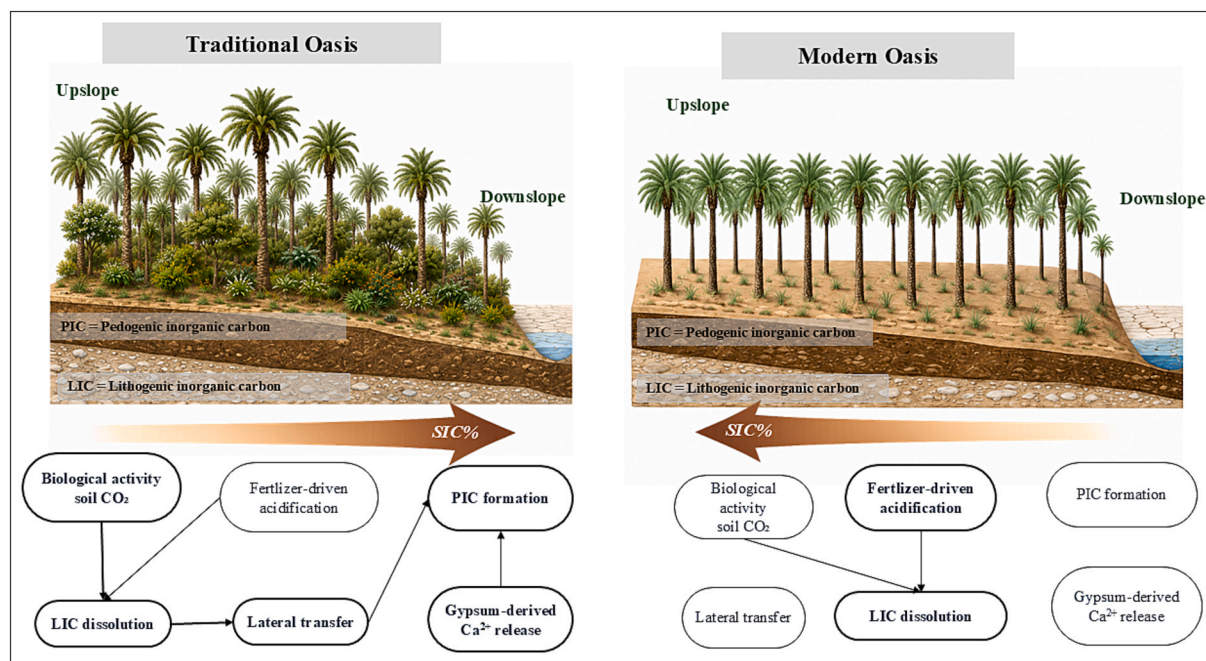


Fig. 6. Conceptual model of soil inorganic carbon (SIC) dynamics in traditional and modern oasis systems. The figure summarizes the inferred roles of lithogenic inorganic carbon (LIC) dissolution through biological activities increasing soil CO_2 and by fertilizer-derived acidification, topographic redistribution causing lateral transfer of dissolved inorganic carbon and Ca^{2+} , pedogenic inorganic carbon (PIC) formation, and gypsum dissolution stimulating PIC formation by providing precipitable Ca^{2+} . The brown gradient arrow represents the relative contribution of SIC to total soil carbon, increasing in traditional oases and decreasing in modern oases systems along the topographic gradient, from upslope to downslope positions. Downslope positions are influenced by the hypersaline Chott lake and connected groundwater systems, upslope positions are close to unmanaged Saharan desert.

structured SIC patterns along the catena, while isotopic evidence ($\delta^{13}\text{C}$ -SIC) revealed shifts in the relative contribution of biogenic versus geogenic carbon sources without indicating net SIC loss. Overall, our findings demonstrate that SIC in oasis systems is governed by strong geological and hydrological controls, with management influencing carbonate cycling processes rather than total carbon storage. These results highlight the importance of considering both landscape position and land-use history when assessing inorganic carbon dynamics in dryland agroecosystems.

CRediT authorship contribution statement

Wajdi Allagui: Writing – review & editing, Writing – original draft, Visualization, Methodology, Investigation, Formal analysis, Data curation. **Nadhem Brahim:** Writing – review & editing, Methodology. **Manel Allani:** Methodology. **Bouthaina Zougari:** Methodology. **Hatem Ibrahim:** Writing – review & editing. **Roland Bol:** Writing – review & editing, Validation. **Hamouda Aichi:** Writing – review & editing. **Wolfgang Wanek:** Writing – review & editing, Validation, Supervision.

Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Appendix A. Supplementary material

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.geoderma.2026.117865>.

Data availability

Data will be made available on request.

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