



Weathering Increases Bedrock Water Storage Capacity but Not Necessarily Its Temporal Persistence in Humid Karst Hillslopes

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10 **Abstract.** Bedrock moisture plays a crucial role in buffering drought, sustaining vegetation, and recharging groundwater in mountainous landscapes with shallow soils. Stronger weathering is generally assumed to enhance these functions by increasing bedrock water availability. However, our understanding of how weathering regulates the storage, retention, and release of rock moisture is still limited. In this study, using two years of high-frequency moisture observations, we quantified relative water storage, temporal stability, and storage-deficit dynamics across the soil–bedrock continuums of two adjacent
15 hillslopes with contrasting weathering intensities in a representative humid karst catchment. Despite having 21% greater maximum dynamic storage capacity than the weakly weathered hillslope (WWH), the strongly weathered hillslope (SWH) realized only 91.65% of the water storage observed at WWH, indicating that greater storage space did not translate into greater realized storage. Strong weathering promoted rapid infiltration and drainage, thereby reducing bedrock water-storage persistence, as indicated by the faster groundwater recession and less persistent shallow groundwater at the SWH. More
20 importantly, these hydrological contrasts also shaped the responses of the two hillslopes to meteorological drought, with the SWH experiencing a greater decline in water storage than the WWH when annual precipitation notably reduced. This greater sensitivity to reduced rainfall may partly explain the shorter trees and smaller stem diameters observed at the SWH. Accordingly, these results challenge the assumption that stronger weathering could increase water availability, providing key insights into drought resilience, groundwater recharge, and runoff generation in karst critical zones.

25 1 Introduction

Global climate change is destabilizing terrestrial ecohydrological processes by increasing the frequency of extreme precipitation and prolonged drought (Chen et al., 2025a). These hydroclimatic extremes can suppress vegetation growth and ecosystem carbon uptake and even large-scale canopy deaths (Gebrechorkos et al., 2025; Luo et al., 2024). However, vegetation growing in shallow soils over weathered bedrock can remain comparatively resilient under similar climatic
30 extremes (Leite et al., 2025; McCormick et al., 2021; Rempe & Dietrich, 2018). This resilience is often attributed to rock moisture stored within bedrock pores and fractures, which can sustain plant water use after water in the shallow soil has become limited (McCormick et al., 2021; Salve et al., 2012). The ecohydrological importance of this subsurface reservoir has



been documented across diverse climatic and geological settings (Leite et al., 2025; Li et al., 2023; McCormick et al., 2021). Although previous studies have examined plant use of rock moisture and its seasonal variations, high-frequency observations that jointly quantify water-storage and its persistence across the soil–bedrock continuum remain limited. This uncertainty limits the representation of accessible bedrock water in catchment, land-surface, and Earth system models and consequently constrains accurate assessments of ecosystem resistance to drought at regional and global scales under climate changes (Chen et al., 2024; Fan et al., 2019).

A key challenge is that weathered bedrock is not readily represented by the hydraulic relationships developed for granular soils, whose water-retention and hydraulic-conductivity functions can often be inferred from measurable properties (Assouline, 2021; Nasta et al., 2013). Weathered bedrock instead comprises matrix pores and fractures with markedly contrasting hydraulic behaviors (Katsura et al., 2024; Liu et al., 2025a). Water stored in small matrix pores may persist for extended periods, whereas connected fractures can rapidly transmit seasonal recharge along preferential flow paths (Pedrazas et al., 2021). Bedrock water-storage dynamics therefore depend on both the storage provided by the rock matrix and the geometry and connectivity of fractures (Hahm et al., 2020; Salve et al., 2012). Because matrix porosity and fracture architecture evolve during weathering, contrasting weathering profiles may differ not only in the amount of water they can store but also in how long that water is retained (Wang et al., 2026). Resolving this capacity–persistence relationship is critical for predicting hillslope water availability, groundwater recharge, and runoff generation (Leite et al., 2025; Wang et al., 2023).

Evidence from mountainous landscapes indicates that subsurface water-storage capacity is strongly regulated by the degree and vertical extent of bedrock weathering (Hahm et al., 2019; Rempe & Dietrich, 2018). This control arises because differences in weathering reorganize the structure and pore architecture of deeper profiles (Wang et al., 2026). Weathering-induced rock fragmentation can create longer, less tortuous, and more highly connected flow paths, thereby increasing hydraulic conductivity and promoting preferential flow (Zhang et al., 2021). As these structural alterations propagate downward, the subsurface domain that actively participates in water storage and flow expands. As Pedrazas et al. (2021) found that seasonally dynamic rock moisture was concentrated primarily within the pervasively fractured and oxidized zone. However, more intense weathering does not necessarily produce a proportional increase in water capacity because the resulting hydraulic behavior depends on how weathering redistributes fine particles, rock fragments, pores, and flow pathways (Katsura et al., 2024; Wang et al., 2026). For example, Xiang et al. (2026) found that strongly weathered bedrock acted as a transmissive zone that promoted downward leakage and limited water accumulation at the soil–bedrock interface. By contrast, harder and less-weathered bedrock had sharply lower porosity and consequently formed a compact hydraulic barrier that impeded vertical water movement and maintained a more persistent water-retaining zone above it (Wang et al., 2026). These contrasting responses indicate that the degree of weathering determine not only the volume available for water storage but also the partitioning of incoming water between retention and drainage (Fu et al., 2025; Wang et al., 2026). However, continuous in filed observations remain too limited to determine whether stronger weathering increases water storage capacity or instead accelerates depletion.



The unresolved balance between water storage and release is particularly important in karst landscapes with shallow soils, which cover approximately 15% of the global land surface and provide drinking water to one quarter of the world's population (Goldscheider et al., 2020). Intense carbonate dissolution has produced a highly heterogeneous, coupled surface–subsurface structure characterized by a network of matrix pores, fractures, and conduits (Hartmann et al., 2014). This heterogeneous structure can retain and slowly release water through matrix pores and soil-filled fractures while simultaneously transmitting water rapidly through connected fractures and conduits (Perrin et al., 2003; Zhang et al., 2024a). Consequently, the precipitation in karst regions does not necessarily result in persistent subsurface water storage because a substantial fraction of recharge may bypass slowly draining domains and be rapidly discharged from the shallow critical zone (Liu et al., 2026). The capacity of karst bedrock to sustain groundwater discharge and ecosystem water supply during rain-free periods therefore depends on the proportion of recharge retained within matrix pores and poorly connected fractures rather than on total precipitation alone (Lapides et al., 2025; Liu et al., 2025a). When this retained storage is limited, meteorological drought can rapidly propagate into groundwater and streamflow deficits, helping to explain the greater frequency of hydrological drought reported in karst regions than in non-karst regions (Nerantzaki & Nikolaidis, 2020; Shi et al., 2024). Quantifying bedrock moisture storage and depletion is therefore essential for assessing groundwater recharge, dry-season water availability, and drought vulnerability in karst regions.

Southwest China contains the world's largest contiguous exposed karst region, characterized by shallow, highly permeable soils and rapid hydrological exchange between the land surface and groundwater (Chen et al., 2025b; Li et al., 2023). Unlike the arid and semiarid karst regions examined in many previous studies, this humid karst region receives abundant monsoonal rainfall. Nevertheless, rapid subsurface routing can raise groundwater levels during wet periods while depleting shallow storage between rainfall events, increasing groundwater-flooding risk and drought vulnerability (Long et al., 2014; Zhang et al., 2024a). The partitioning of rainfall among surface runoff, shallow storage, and subsurface drainage is governed largely by soil thickness and the pore-network architecture of the critical zone (Fu et al., 2016). However, these controls are not static. As bedrock weathering proceeds, soil thickness and pore-network architecture co-evolve, continually altering subsurface water exchange, retention, and release (Wang et al., 2024). Despite the apparent ecohydrological importance of this co-evolution, surprisingly few studies have explicitly linked the degree of bedrock weathering to water-storage dynamics and groundwater responses.

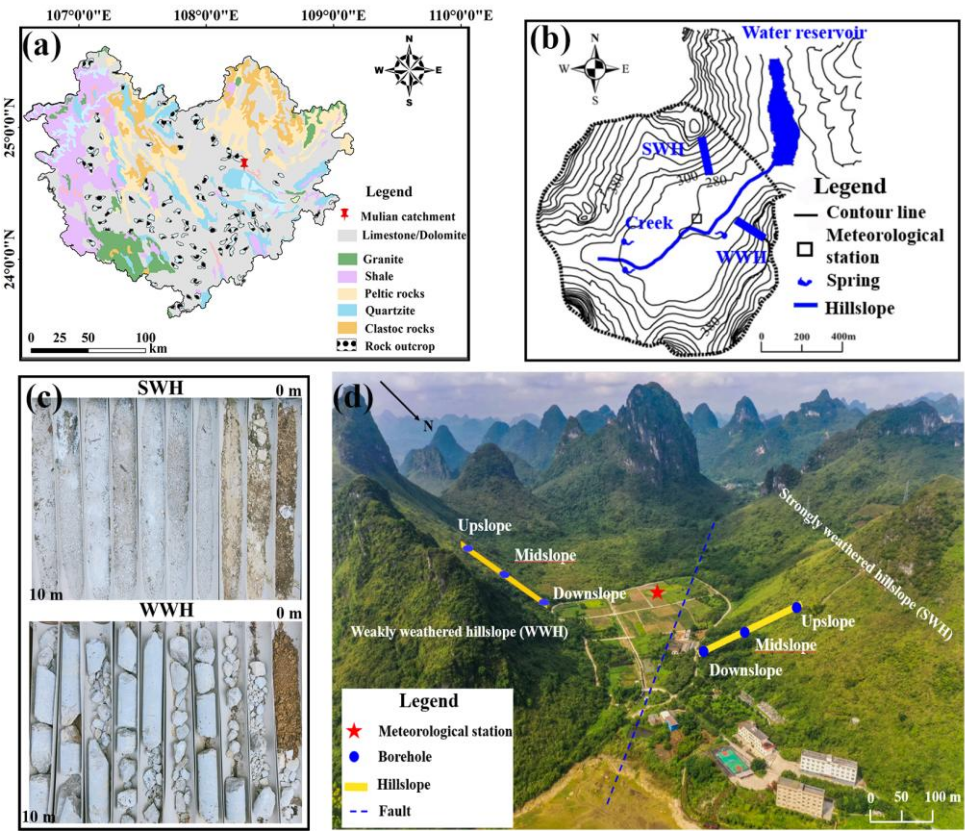
Based on the above evidence, a potential tradeoff may exist between storage expansion and drainage enhancement as bedrock weathering progresses. We hypothesized that a greater degree of bedrock weathering would expand the active storage domain but enhance subsurface drainage, thereby reducing water-storage persistence. To test this hypothesis, we compared the water storage dynamic of two adjacent karst hillslopes with contrasting weathering intensities using two years of high-frequency observations of soil and bedrock water content. We quantified the respective contributions of soil and bedrock to dynamic profile water storage and evaluated differences in relative water storage, temporal stability, storage-deficit dynamics, and groundwater responses between the two hillslopes.



100 **2 Materials and Methods**

2.1 Site Description

The study was conducted within the Mulian catchment (1.14 km²; 24°43' N, 108°18' E), situated in Guangxi Province, southwest China. The catchment exhibits a representative peak-cluster depression morphology, characterized by a central flat basin surrounded by high-relief ridges and overlapping hills—with elevations ranging from 272 to 647 m above sea level (Fig. 1a). Geologically, the region overlies extensive Carboniferous carbonate successions, primarily composed of dolomite and dolomitic limestone (Zhang et al., 2026a). Within this small watershed, a SE-trending fault bisects the terrain into two distinct hillslope systems with divergent subsurface architectures: a soil-rock mixture habitat on the NE side and a continuous bedrock hillslope on the SW side, where bedrock is extensively exposed (Zhang et al., 2026a) (Fig. 1b).



110 **Figure 1: Location, geological setting, and monitoring layout of the study area. (a) Regional lithological map showing the location of the Mulian catchment. (b) Topographic map showing the strongly weathered hillslope (SWH), weakly weathered hillslope (WWH), creek, spring, reservoir, and meteorological station. (c) Representative borehole cores from 0–10 m depth at the SWH and WWH. (d) Aerial view of the two adjacent hillslopes showing the monitoring transects, boreholes at the downslope, midslope, and upslope positions.**



115 The study area is characterized by a subtropical monsoon climate, with a mean annual precipitation (MAP) of 1,446 mm (Zhang et al., 2024a). Rainfall exhibits marked seasonality—the wettest month being June (268.7 mm) and the driest being February (24.2 mm). Generally, over 60% of the annual precipitation falls between April and September, reflecting distinct wet and dry seasons. The mean annual temperature is 19.4 °C, with monthly extremes recorded in July (27.4 °C) and January (9.0 °C).

120 Soil distribution is strongly tied to topography and exhibits significant spatial heterogeneity. In contrast to the relatively deep soils in depressions (20–200 cm), hillslope soils thin progressively upslope, ranging from 0–60 cm. These soils are predominantly clay loam or clay, characterized by high clay content (30–60%), rich organic matter (1–12%), and abundant rock fragments (>2 mm) which account for 10–40% of the total mass. Groundwater is shallow (1–6 m) and exhibits seasonal fluctuations in response to precipitation, with numerous epikarst springs emerging at the foot of slopes. Vegetation patterns

125 are closely coupled with these edaphic conditions. The overstory is dominated by *Radermachera sinica*, *Celtis biondii*, *Rhus chinensis*, and *Schefflera octophylla*. The shrub layer is primarily composed of *Vitex negundo*, *Ligustrum lucidum*, and *Pyracantha fortuneana*, while the herbaceous community is characterized by *Miscanthus floridulus*, *Bidens pilosa*, and *Allantodia virescens*.

2.2 Determination of Weathering Degree for the Two Hillslopes

130 To determine the degree of bedrock weathering, boreholes were drilled to a depth of approximately 5 m at the downslope, midslope, and upslope positions of each hillslope, and continuous rock cores were recovered. Weathering-related characteristics, including core integrity, fracture development, pore occurrence, and visible alteration, were documented through core logging. Representative rock and regolith samples were analyzed for major-oxide composition by X-ray fluorescence spectrometry (XRF) using a Philips PW2404 spectrometer following GB/T 14506.28–2010 (Li & Ji, 2015). The

135 chemical index of alteration (CIA), which has been used to identify progressive weathering in carbonate profiles (Chen et al., 2025b), was calculated from the major-oxide data. The weathering degree of each hillslope was determined by integrating the rock-core characteristics, major-oxide compositions, CIA values, and their vertical variations (Fig. 1c and 2).

The rock cores reveal a pronounced structural contrast. The strongly weathered hillslope (SWH) cores consist predominantly of friable, highly fractured, and visibly altered materials, whereas the weakly weathered hillslope (WWH) cores contain

140 largely intact and blocky dolomite with small pores, calcite vugs, and localized dissolution fractures (Fig. 1c). The geochemical results reinforce this structural distinction. Relative to the WWH, the SWH generally has higher SiO₂, Al₂O₃, Fe₂O₃, Na₂O, K₂O, and CIA values, whereas the WWH is consistently enriched in CaO and MgO (Fig. 2a). Furthermore, vertical geochemical differentiation is also more pronounced in the SWH, where CaO and MgO contents increase with depth while SiO₂, Al₂O₃, Fe₂O₃, Na₂O, K₂O, P₂O₅, and CIA values generally decrease. By comparison, substantial geochemical

145 variation in the WWH is confined mainly to the near-surface layer (0–60 cm), with only minor changes at greater depths. Although most samples from both hillslopes plot within the weakly weathered field in CIA–Na/K space, the broader distribution of the SWH samples indicates greater spatial heterogeneity in weathering intensity (Fig. 2b). Collectively, these



structural and geochemical contrasts demonstrate that the SWH has a more vertically differentiated and spatially heterogeneous weathering profile, providing the basis for evaluating how weathering influences subsurface water storage and dynamics.

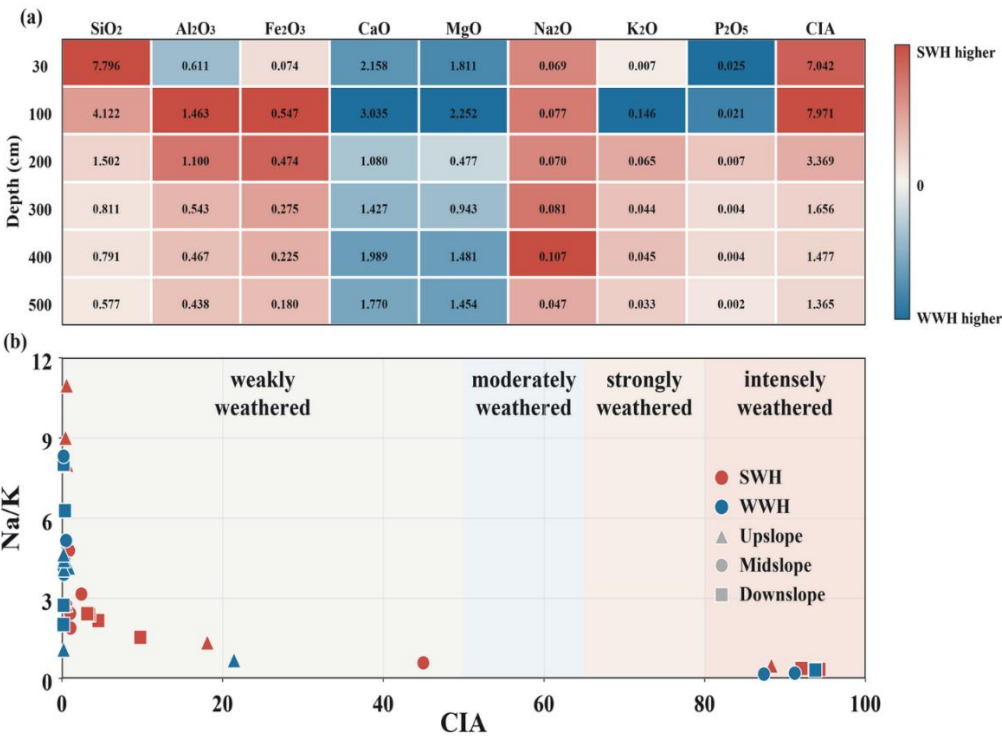


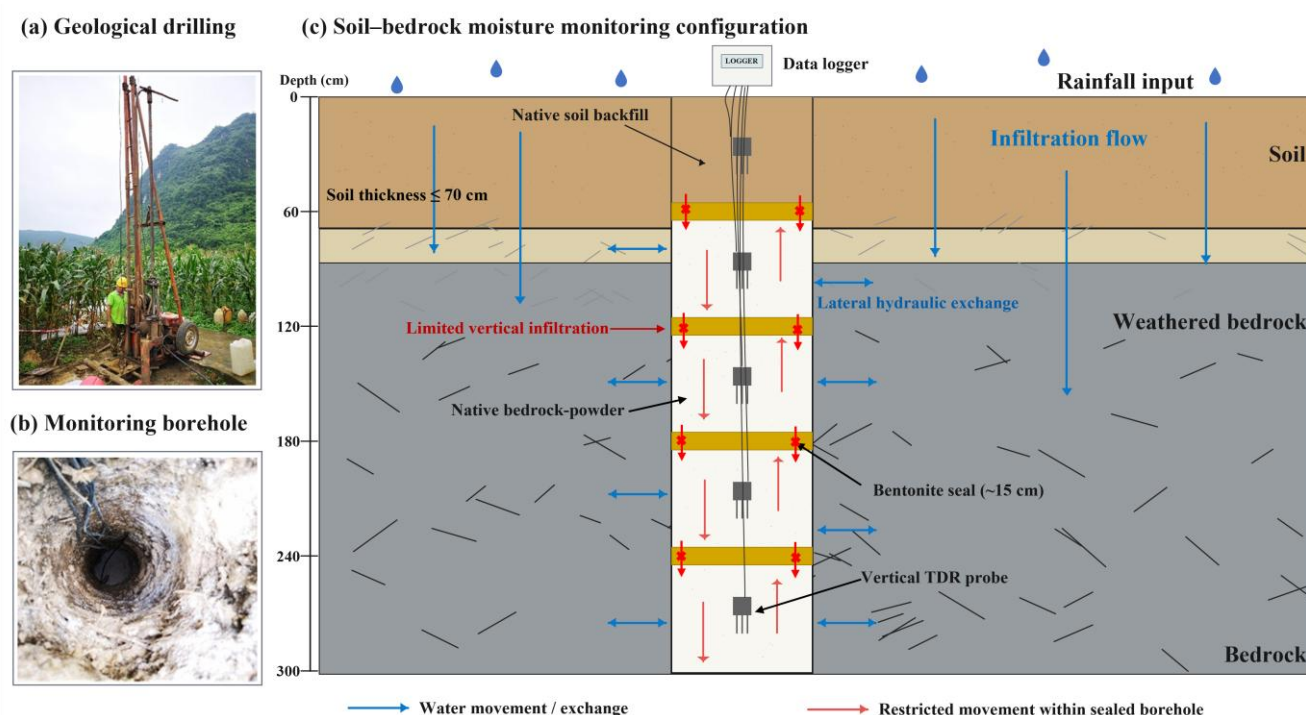
Figure 2: Major oxide differences and chemical weathering status of SWH and WWH. (a) Depth profiles of mean differences in major oxide concentrations and chemical index of alteration (CIA) between SWH and WWH. Red values indicate higher values in SWH, whereas blue values indicate higher values in WWH. (b) Relationship between CIA and Na/K ratio for samples from different hillslope positions. Colors denote hillslope type, and symbols denote hillslope position. Background shading indicates weathering intensity classes from weakly to intensely weathered.

2.3 Soil and Bedrock Relative Moisture Monitoring and Evaluation

In November 2019, six boreholes were drilled across the SWH and WWH using a geological drilling rig (Table S1). Three boreholes were positioned at the upslope, midslope, and downslope locations of each hillslope (Fig. 1d). The boreholes were 7.5 cm in diameter and extended to a depth of 300 cm through the soil and weathered bedrock. Drilling revealed that soil thickness did not exceed 70 cm at any of the monitoring sites (Table S1). TDR sensors (TDR-305N, Acclima, USA) were installed at 60-cm intervals to a depth of 300 cm, providing five monitoring layers per borehole (0–60 cm in soil and 60–300 cm in weathered bedrock; Fig. 3). The probes were installed vertically within their respective layers. Probes in the soil were backfilled with native soil, whereas those below the soil–weathered bedrock interface were embedded in native bedrock



165 powder to ensure hydraulic contact. Low-permeability bentonite seals approximately 15 cm thick were placed between adjacent monitoring layers, except above the uppermost layer, to restrict vertical flow and prevent artificial preferential flow along the boreholes (Salve et al., 2012; Tokumoto et al., 2012). Moisture measurements were recorded continuously at 5-min intervals from January 2020 to December 2021, providing two years of high-frequency observations across the soil–bedrock profiles.



170 **Figure 3: Field installation and configuration of the soil–bedrock moisture monitoring system. (a) Geological drilling and (b) the completed monitoring borehole. (c) Layout of the TDR probes installed at 60-cm intervals to a depth of 300 cm. Blue arrows indicate rainfall infiltration through the surrounding formation and lateral hydraulic exchange between the formation and backfill, while red arrows indicate the limited vertical water movement within the sealed borehole. The schematic is not to scale.**

175 For the bedrock intervals, the TDR probes measured the VWC of the backfill rather than that of the intact surrounding rock. Although the backfill exchanged water with the adjacent formation and approached equilibrium in water potential, differences in porosity, bulk density, and pore structure prevented its VWC from being interpreted as the absolute VWC of the surrounding bedrock. We therefore used relative changes in VWC to characterize temporal moisture dynamics. Four aspects of the monitoring design support this application. (1) TDR probes installed in fine-grained backfill have been successfully used to detect wetting-front arrival and temporal moisture variations in weathered granite, limestone, and fractured bedrock (Katsura et al., 2008; Leite et al., 2021; Li et al., 2020). (2) Because the bentonite seals restricted vertical

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flow between monitoring intervals, event-driven moisture changes within each interval resulted predominantly from lateral hydraulic exchange with the adjacent formation and therefore reflected changes in its wetness (Salve et al., 2012). (3) The continuous 5-min measurements offered a clear advantage over low-frequency or discrete measurements in resolving rainfall responses, including response time, magnitude, and recession at individual depths. (4) Identical drilling, sensor installation, backfilling, and sealing procedures were applied to all boreholes at both hillslopes. This standardized protocol minimized methodological differences, allowing systematic contrasts in relative moisture dynamics to be attributed primarily to differences in soil–bedrock structure rather than installation artifacts.

To evaluate the reliability of the TDR probes for quantifying relative moisture changes across heterogeneous soil–bedrock materials, laboratory calibration tests were conducted using an acrylic vessel method (Leite et al., 2021; Vaz et al., 2013). Because dolomite weathering produces a progressive vertical transition from soil to weathered bedrock, a compositional gradient was established by mixing local soil with crushed and sieved weathered-bedrock material (<1 mm). Six treatments were prepared with soil mass fractions of 0%, 20%, 40%, 60%, 80%, and 100% to represent the range of materials and weathering conditions observed within the critical-zone profiles (Fig. 1c and S1). The materials were packed at dry bulk densities ranging from 1.2 g cm^{-3} for pure soil to 1.5 g cm^{-3} for pure weathered-bedrock powder, encompassing values reported for carbonate-derived soils and weathering products in Southwest China (Feng et al., 2024; Li et al., 2023). After packing, VWC was measured using five TDR sensors. The materials were then excavated to the probe depth (approximately 4 cm), and three 100-cm^3 ring samples were collected and oven-dried at 105°C for 24 h to determine reference VWC. This procedure was repeated under five moisture conditions, with four successive increments of approximately $0.05 \text{ cm}^3 \text{ cm}^{-3}$. The increments were selected with reference to the typical saturated VWC of approximately $0.30 \text{ cm}^3 \text{ cm}^{-3}$ in the weathered bedrock.

The calibration tests produced strong linear relationships between the TDR readings and reference VWC across all six treatments ($R^2 > 0.96$; Fig. S1; Table S2). However, systematic baseline offsets were observed. Uncalibrated TDR readings showed maximum underestimation in pure soil (mean bias of $-0.077 \text{ cm}^3 \text{ cm}^{-3}$). In contrast, weathered-bedrock treatments showed varying degrees of overestimation, with a maximum mean bias of $0.069 \text{ cm}^3 \text{ cm}^{-3}$. These differences likely resulted from variations in bulk density, mineral composition, electrical properties, and probe–matrix contact (Sakaki & Rajaram, 2006; Vaz et al., 2013). Despite these offsets, all calibration equations revealed slopes close to unity (0.79 to 1.22) and low residual errors ($\text{RMSE} \leq 0.019 \text{ cm}^3 \text{ cm}^{-3}$). Moreover, for a given monitoring depth, these baseline offsets are additive and time-invariant. As a result, they mathematically cancel out when calculating relative water content changes. We therefore used the raw TDR time series to quantify within-sensor relative water-content and storage dynamics, avoiding the direct application of calibration equations derived from crushed and repacked materials to intact, heterogeneous field bedrock (Salve et al., 2012; Tokumoto et al., 2012). Therefore, the calibration results demonstrate that the TDR measurements reliably captured temporal variations in relative moisture across the soil–bedrock continuum, while the observed slopes and residual errors define the uncertainty associated with the magnitude of these variations.



2.4 Determination of Relative Water Storage (RWS) and Temporal Persistence

While VWC varies considerably owing to variability in soil and bedrock properties, changes in relative water storage (RWS; mm) are more comparable across different layers, making it a useful indication for water balance calculations (Seyfried & Wilcox, 2006). For this reason, we calculated RWS at each depth increment as follow (Leite et al., 2025):

$$RWS = (VWC - VWC_{min}) \times D \quad (1)$$

where VWC is the volumetric water content measured every five minutes, VWC_{min} is the lowest volumetric water content recorded at that depth during the study, and D is the thickness represented by each TDR sensor. Because drilling revealed variable soil thicknesses of ≤ 70 cm among the boreholes (Table S1), the sensor at the 60-cm depth was used to represent the entire soil layer, with D equal to the soil thickness ($60 \leq h_s \leq 70$ cm). The sensor at 120 cm represented the weathered bedrock extending from the soil–bedrock interface to 120 cm, corresponding to a thickness of $120 \text{ cm} - h_s$. The sensors at 180, 240, and 300 cm represented the fixed 120–180, 180–240, and 240–300 cm intervals, respectively. For each monitoring day, total RWS over the 300-cm profile was calculated by summing the layer-specific RWS values derived from the daily mean VWC of each interval. The maximum water storage was then defined as the highest daily total RWS recorded during the entire study period. This estimate only represents a lower bound, as wetter years or additional large storms could have resulted in higher VWC—particularly in bedrock and deep soils (Leite et al., 2025).

Water-storage persistence in this study was defined as the extent to which stored water was maintained over time with limited relative fluctuation. We quantified this property using the temporal stability (T_s) of daily RWS, expressed as the complement of its coefficient of variation following Isbell et al. (2026). Before calculating T_s , daily RWS was normalized to the observed storage range for each hillslope, slope position, and depth interval:

$$RWS^* = \frac{RWS_t - RWS_{min}}{RWS_{max} - RWS_{min}} \quad (2)$$

$$T_s = 1 - CV(RWS^*) = 1 - \frac{\sigma_{RWS^*}}{\mu_{RWS^*}} \quad (3)$$

where RWS_t is the daily RWS at time t ; RWS_{min} and RWS_{max} are the minimum and maximum daily RWS for the corresponding hillslope, slope position, and depth interval; and μ_{RWS^*} and σ_{RWS^*} are the mean and standard deviation of the normalized daily RWS series, respectively. Normalization enabled comparisons among layers and hillslopes with different storage capacities. Higher T_s indicates that storage remained relatively high with limited variability, representing greater water-storage persistence, whereas lower or negative T_s indicates greater fluctuations and more frequent depletion (Isbell et al., 2026). Soil T_s was calculated using the 0–60 cm interval, whereas bedrock T_s was calculated from normalized daily RWS averaged across the 60–300 cm intervals.

It is worth noting that using discrete, point-based TDR measurements to generalize water content across an entire 300-cm interval introduces inherent uncertainties, particularly within the uppermost soil layer. Small rainfall events that only induce moisture fluctuations in the near-surface soil layers may evaporate or be transported before reaching the 60-cm sensor, leading to a localized underestimation of topsoil water storage. Nonetheless, this limitation does not compromise our primary



objective of tracking bedrock critical zone hydrologic dynamics, as minor precipitation events rarely penetrate deeply enough to recharge the complex drainage network within the underlying weathered bedrock (Sala et al., 1992).

250 2.5 Determination of Low-Storage Conditions and Water Storage Deficits

To characterize the relative depletion of subsurface water storage, we used normalized relative water storage (RWS^*) as an indicator of relative storage state in both soil and bedrock layers. Because RWS^* expresses daily storage relative to the observed minimum and maximum storage at each hillslope position and depth interval, it provides a dimensionless measure of local water availability that can be compared across layers with different storage capacities. Here, we refer to the study by
 255 Rahman et al. (2024) and Zhang et al. (2024b), defining days with an $RWS^* < 0.30$ as low water storage days. In addition, the absolute daily relative water storage deficit (RWSD) for each monitoring unit was used to quantify the difference between the maximum observed storage and the storage on day t :

$$RWSD_t = RWS_{max} - RWS_t \quad (4)$$

where RWS_{max} is the maximum daily RWS observed for a given hillslope, slope position, and depth interval during the study
 260 period, and RWS_t is the daily RWS on day t .

To further characterize the rates at which storage deficits developed and were relieved, we calculated the daily change in RWSD between two consecutive days:

$$\Delta RWSD_t = RWSD_{t+1} - RWSD_t \quad (5)$$

Positive values of $\Delta RWSD$ indicate an increase in storage deficit and therefore represent deficit accumulation, whereas
 265 negative values indicate a reduction in storage deficit and therefore represent storage recovery. Deficit accumulation rate (DAR) was calculated from all positive daily changes in RWSD. This metric quantifies how rapidly water storage moved away from the maximum observed storage state and therefore describes the tendency toward low-storage conditions.

Recovery rate was calculated only when the preceding day was under low-storage conditions ($RWS_t < 0.30$). This restriction ensured that the recovery metric represented the ability of soil or bedrock storage to recover from a low-storage state. When
 270 $\Delta RWSD_t < 0$ and $RWS_t < 0.30$, deficit recovery rate (DRR) was calculated. Higher DRR values indicate faster reduction of storage deficits during low-storage conditions. For each hillslope and layer, DAR and DRR values were pooled across slope positions and used to construct cumulative probability distributions. These distributions allowed us to compare not only the magnitude and frequency of low-storage conditions, but also the rates at which water storage deficits accumulated and recovered in soil and bedrock across the two hillslopes.

275 2.6 Meteorological and Groundwater Data Collection

Hourly air temperature and relative humidity were obtained from a meteorological station located within the Mulian catchment and aggregated to daily mean values. Rainfall was measured using a field-calibrated tipping-bucket rain gauge with a resolution of 0.2 mm. Because the SWH and WWH were adjacent to the meteorological station, the same daily meteorological record was used to represent atmospheric forcing at both hillslopes.



280 Groundwater levels were monitored from April 2020 to December 2021 in 11-m observation wells located at the upslope, midslope, and downslope positions of both hillslopes. Water levels were expressed as depth below ground surface (m). Measurements at or below the 11-m well limit were recorded as >11 m and treated as censored observations. In addition, daily groundwater-level observations from an adjacent depression were used as a hydrological background reference. This record was used to characterize catchment-scale groundwater conditions and to contextualize the contrast between the wet
 285 year and the normal year.

To compare groundwater recharge and drainage behavior between hillslopes, we identified independent rainfall events and quantified four complementary metrics: response time, peak lag, the percentage of monitored days with the water table occurring within the 11-m well depth, and groundwater recession rate. Response time represented the interval between rainfall onset and the first detectable water-table rise, whereas peak lag represented the time from rainfall onset to the
 290 shallowest post-event water level. Groundwater recession rate was quantified as the mean daily deepening of the water table during consecutive rain-free periods following recharge events; larger values indicate faster post-event drainage. These event-response and recession metrics were selected to characterize the timing of rainfall recharge, groundwater persistence, and post-recharge drainage in fractured bedrock systems (Cai & Ofterdinger, 2016).

3 Results

295 3.1 Contrasting Hydroclimatic Regimes in the Two Monitoring Years

During the two-year monitoring period, annual precipitation differed markedly between years. In 2020, cumulative precipitation reached 2,245.90 mm, approximately 55% above the long-term annual mean of 1,446 mm (Fig. 4a). Rainfall was concentrated in several large summer and early-autumn events, including an approximately 282-mm storm in September. Soil relative moisture remained higher at the WWH than at the SWH throughout the year, whereas bedrock relative moisture
 300 was generally higher at the WWH but briefly became higher at the SWH during the high-rainfall period (Fig. 4b). The September storm also caused groundwater to rise abruptly from approximately 3–4 m below the ground surface to near-surface levels, followed by a gradual recession through late autumn and winter (Fig. 4c). During the summer wet season, relative humidity frequently reached 85–90%, mean daily air temperature approached 30°C, and daily maximum temperature approached 40°C (Fig. 4d and 4e). In 2021, cumulative precipitation was 1,325.70 mm, slightly below the long-term annual
 305 mean and only 59.03% of the 2020 total. Rainfall accumulated more gradually and lacked the large storm pulses observed in the previous year. Soil relative moisture at the WWH remained higher than at the SWH, while bedrock relative moisture was also higher at the WWH throughout most of the year, with no reversal comparable to that during the wet period of 2020 (Fig. 4b). Groundwater rose gradually during the spring–summer rainy period but did not approach the ground surface as abruptly as it did follow the September 2020 storm. The water table subsequently declined during the late-season dry period as
 310 rainfall inputs diminished (Fig. 4c).

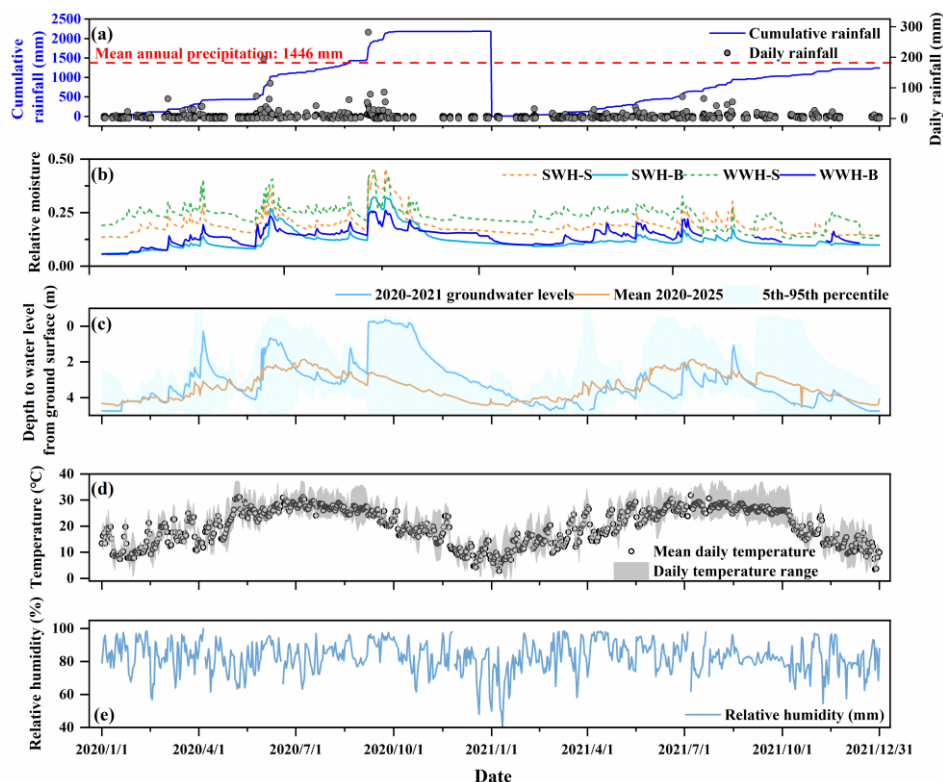
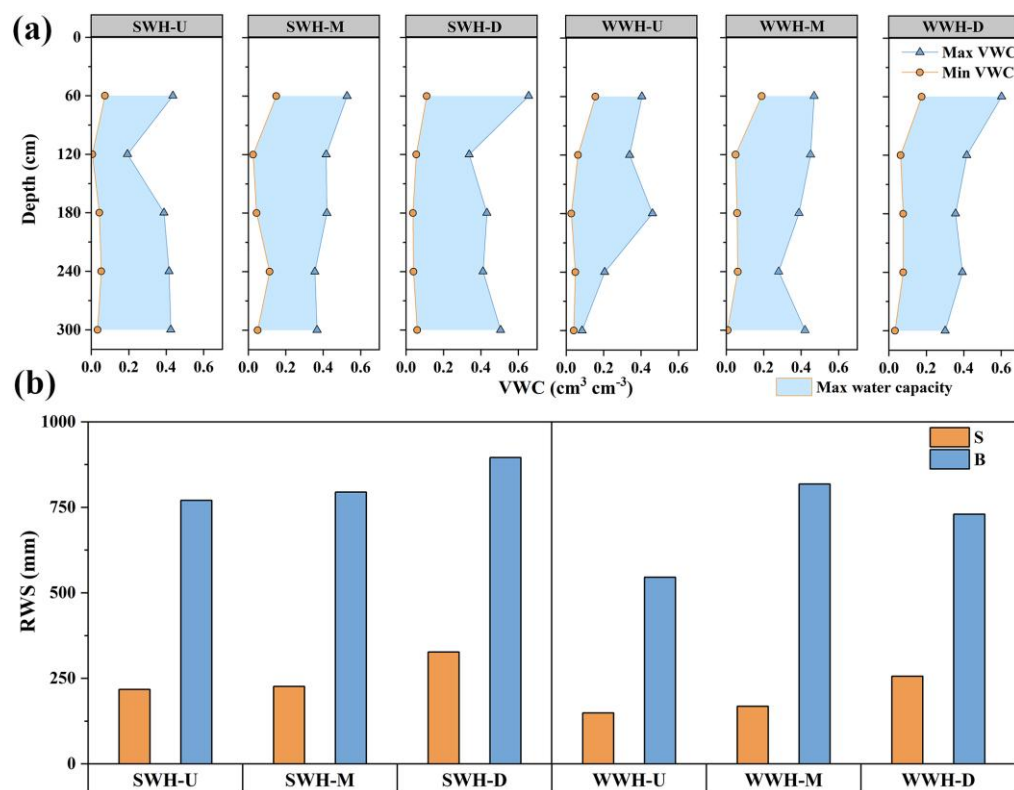


Figure 4: Hydroclimatic conditions and mean soil and bedrock relative-moisture dynamics during the 2020–2021 monitoring period. (a) Daily rainfall and annual cumulative rainfall; the red dashed line indicates the long-term mean annual precipitation of 1,446 mm. (b) Relative moisture in soil and bedrock averaged across the upslope, 315 midslope, and downslope positions of the strongly weathered hillslope (SWH) and weakly weathered hillslope (WWH). (c) Groundwater depth during 2020–2021, together with the 2020–2025 daily mean and 5th–95th percentile range. (d) Mean daily air temperature and daily temperature range. (e) Daily relative humidity. S and B in panel (b) denote soil and bedrock, respectively.

3.2 Water Storage Capacity: SWH Versus WWH

320 Relative volumetric water content (VWC) varied obviously between soil and weathered bedrock, with contrasting depth profiles between the two hillslopes (Fig. 5a). The SWH exhibited a broader VWC range throughout the weathered-bedrock zone, particularly at depths of 120–300 cm, whereas the WWH generally showed a narrower range, most notably at the upslope position. These differences indicate a larger dynamically active storage domain in the strongly weathered bedrock.



325 **Figure 5: (a) Minimum and maximum volumetric water content (VWC) within the upper 300 cm of each borehole. (b)**
Relative water storage (RWS) capacities of soil (S) and bedrock (B). D, M, and U denote the downslope, midslope,
and upslope positions, respectively.

Weathered bedrock dominated dynamic profile storage on both hillslopes (Fig. 5b), storing on average 3.19 and 3.65 times as much water as the soil in the SWH and WWH, respectively. Averaged across the three monitored slope positions, mean
 330 dynamic profile storage was 1,077.6 mm in the SWH, 1.21 times that in the WWH (889.0 mm). Storage also varied with slope position. Total storage decreased from downslope to upslope in the SWH, whereas bedrock storage in the WWH was greatest at the midslope and lowest at the upslope position. Overall, stronger weathering enlarged the dynamically active bedrock storage domain but was accompanied by greater spatial and vertical variability.

3.3 Temporal Trends in Relative Water Storage

335 Overall, the heatmap revealed pronounced temporal and spatial differences in RWS between the SWH and WWH (Fig. 6). Although the estimated maximum storage capacity was greater in the SWH (1077.60 mm) than in the WWH (889.0 mm), the two-year mean realized profile storage was slightly higher in the WWH (245.85 mm) than in the SWH (225.32 mm). This contrast indicates that greater storage capacity did not necessarily translate into more persistent water storage. Spatially, the WWH exhibited strong topographic differentiation, with the highest mean profile storage at the downslope position (376.04



340 mm), followed by the midslope (187.21 mm) and upslope (174.31 mm). In contrast, profile storage in the SWH varied less among slope positions, ranging from 214.41 to 235.56 mm. Notably, the downslope bedrock of the WWH maintained consistently high storage, particularly at 180-300 cm, where mean storage reached 95.10, 112.00, and 64.90 mm at 180, 240, and 300 cm, respectively. These values were substantially higher than those at comparable depths in the SWH, indicating a persistent deep-storage zone within the weakly weathered hillslope.



345 **Figure 6: Daily rainfall and relative water storage at different depths and slope positions during the monitoring period. The grey sections indicate missing data.**

The temporal stability analysis (Fig. 7) further confirmed the differences in the water storage dynamics between SWH and WWH. Averaged across slope positions and layers, temporal stability was higher in the WWH ($T_s = 0.38$) than in the SWH ($T_s = 0.26$), indicating greater overall persistence of water storage in the weakly weathered hillslope. In the soil layer, temporal stability showed no significant difference between the two hillslopes at any slope position, although WWH tended to exhibit slightly higher stability at the upslope and downslope positions. By contrast, the bedrock layer showed a clear hillslope-dependent pattern. At the upslope and midslope positions, bedrock temporal stability did not differ significantly between SWH and WWH. However, at the downslope position, WWH bedrock exhibited significantly higher temporal stability than SWH bedrock ($P < 0.001$), with temporal stability increasing from 0.09 in SWH to 0.55 in WWH. In contrast, the low temporal stability of SWH downslope bedrock indicates stronger temporal variability, consistent with a recharge-rapid recession storage behavior following rainfall events.



Inter-annually, annual rainfall decreased from 2245.90 mm in wet year to 1325.70 mm in normal year, a decline of 40.97%. Mean profile storage in the SWH decreased from 258.61 mm in wet year to 191.93 mm in normal year, corresponding to a 25.78% decline. In comparison, WWH profile storage decreased from 266.08 to 223.21 mm, a smaller decline of 16.11%. Thus, the combined RWS and temporal stability results indicate that WWH functioned as a more persistent deep-water reservoir, whereas SWH had greater maximum storage capacity but lower storage persistence.

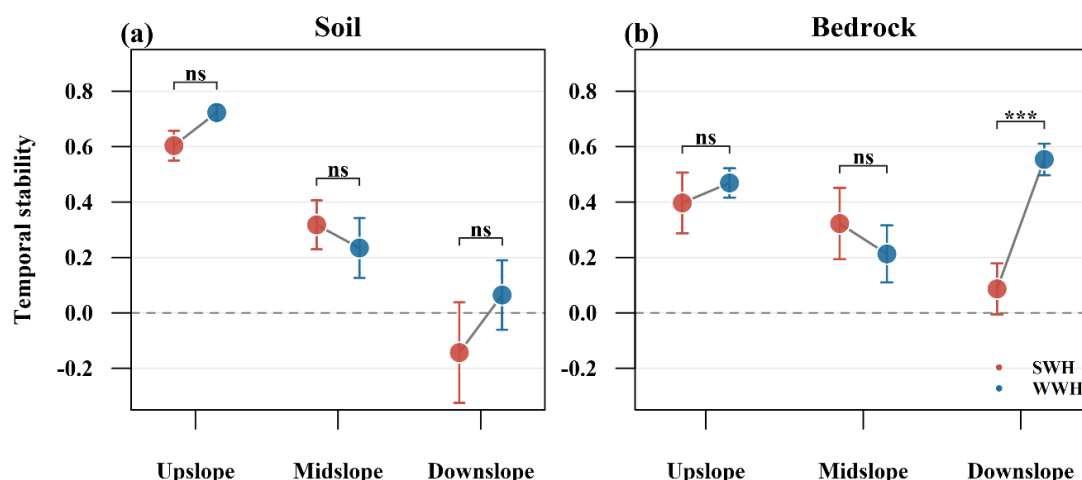


Figure 7: Temporal stability (Ts) of soil and weathered-bedrock water storage across hillslope positions. Higher Ts values indicate greater temporal stability. * indicate significant difference ($p < 0.001$) between the Ts of the SWH and WWH at the same slope position; ns indicates no significant difference.**

3.4 Low-Storage Dynamics and Storage-Deficit Responses

The contrast between the two years revealed pronounced differences in soil and bedrock water storage deficits and recovery between the SWH and WWH (Fig. 8 and 9). Across both soil and bedrock layers, the SWH consistently exhibited lower RWS* than the WWH, with the separation most evident during the normal year (Fig. 8b and 8c). In the soil layer, mean RWS* in SWH declined from 0.25 ± 0.16 in the wet year to 0.18 ± 0.07 in the normal year, whereas the WWH maintained higher values of 0.37 ± 0.16 and 0.32 ± 0.11 , respectively. In the bedrock layer, SWH RWS* decreased from 0.27 ± 0.21 to 0.19 ± 0.05 , whereas WWH remained higher, decreasing only from 0.35 ± 0.18 to 0.33 ± 0.14 . These patterns indicate that WWH retained a larger fraction of its available storage range under both wet and dry conditions.

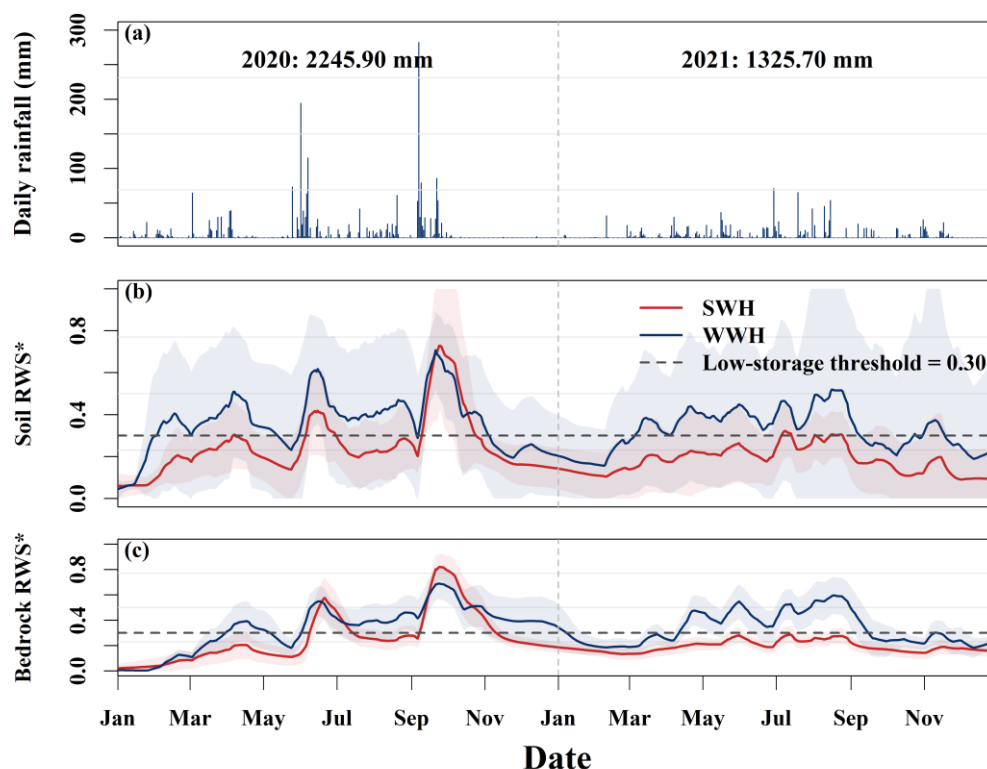


Figure 8: Temporal dynamics of normalized relative water storage (RWS*) in soil and bedrock, illustrating the criterion used to identify low-storage conditions. (a) Daily rainfall, with annual totals shown for 2020 and 2021. (b, c) Mean RWS* dynamics in the soil and bedrock layers for SWH and WWH. The horizontal dashed line indicates the low-storage threshold (RWS* = 0.30).

The contrast between the two hillslopes became more evident when the low-storage days and absolute storage deficits were considered together. In wet year, the SWH already experienced substantially more low-storage days than the WWH, with low-storage days frequencies of 79.78% in soil and 74.86% in bedrock, compared with 35.52% and 27.87% in the WWH (Table S3). Under the normal rainfall conditions of 2021, low-storage days increased to 93.15% in SWH soil and 95.89% in SWH bedrock, while the corresponding values in the WWH were much lower, 44.66% and 54.80%. Mean daily water deficits showed the same pattern. In normal year, the daily water deficits of SWH reached 179.68 ± 16.28 mm in soil and 585.85 ± 33.84 mm in bedrock, both exceeding the WWH values of 116.71 ± 24.53 mm and 385.86 ± 120.89 mm. Thus, the SWH was characterized by more persistent low-storage conditions and larger absolute water deficits, whereas the WWH showed stronger drought buffering in both soil and bedrock.

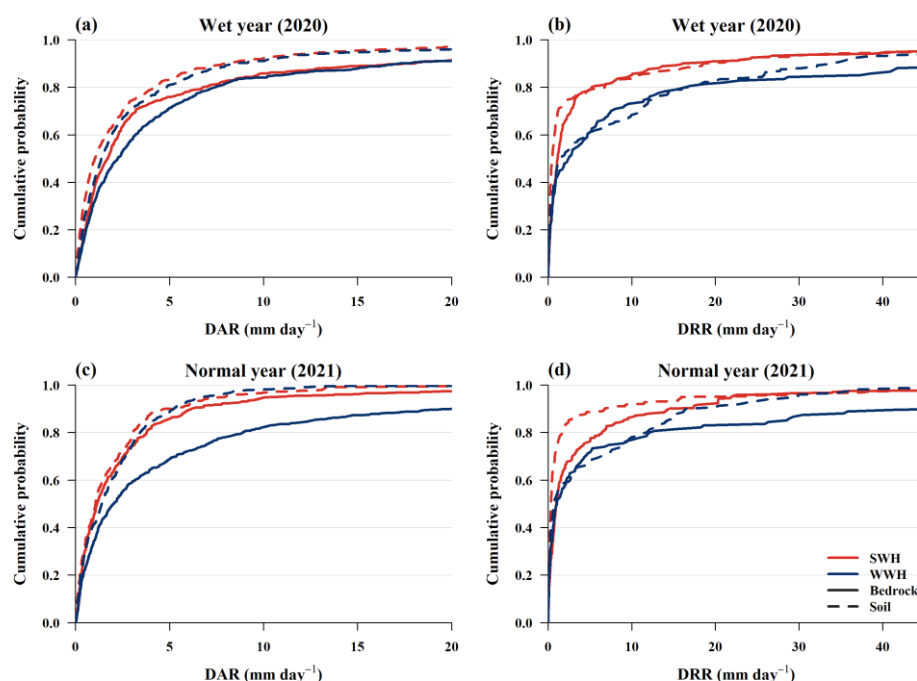


Figure 9: Cumulative distributions of water storage deficit rates and deficit recovery rates in soil and bedrock layers.

Analysis of cumulative distributions of deficit accumulation rate (DAR) and deficit recovery rate (DRR) further highlighted differences in storage dynamics (Fig. 9). Across both years, WWH bedrock DAR curves were consistently shifted toward higher values relative to SWH, reflecting larger episodic changes in deficit accumulation. This should not be interpreted as more severe storage depletion in WWH, because the WWH simultaneously maintained higher RWS*, lower mean daily deficits, and fewer low-storage days. Instead, these distributions indicate that WWH bedrock storage fluctuated within a wetter and more buffered domain, whereas SWH remained closer to the low-storage threshold for most of the record (Fig. 8b and c). Recovery dynamics reinforced the functional advantage of WWH. During periods of low storage, WWH exhibited higher DRR than SWH, particularly in the bedrock layer, as shown by right-shifted curves in the DRR panels (Fig. 9b and d). This indicates that, once low-storage conditions occurred, deficits in WWH were relieved more rapidly than in SWH.

4 Discussion

4.1 Weathering Expands Shallow Bedrock Water Storage Domain

Karst landscapes are typically characterized by shallow and discontinuous soils, which inherently limit soil water storage (Wang et al., 2019). However, plant water acquisition is not confined to the soil layer. Roots can penetrate carbonate bedrock along fractures and dissolution features, allowing vegetation to access water stored within weathered bedrock (Nie et al., 2014). In northern Yucatán, roots were observed throughout a 9-m limestone profile, with some reaching the underlying



aquifer (Estrada-Medina et al., 2013). Such deep roots can supply plants with up to five times as much water as shallow roots during drought, and a similar reliance on bedrock water has been observed in Southwest China (Bleby et al., 2010; Liu et al., 2014). These observations indicate that the use of deep rock water is widespread in shallow-soil landscapes and that water stored in weathered bedrock can constitute an ecologically important reservoir. Consistent with this understanding, our results showed that weathered bedrock, rather than soil, constituted the dominant dynamic water-storage domain on both hillslopes, where soil thickness did not exceed 70 cm. Dynamic bedrock storage was 3.19 and 3.65 times the observed dynamic storage of the overlying soil in the SWH and WWH, respectively (Fig. 5). This reflects the limited storage capacity of shallow soil and the much greater thickness of bedrock participating in wetting and drying than is commonly assumed. Similar contributions of bedrock water have been documented in other shallow-soil landscapes (Hahm et al., 2022; Rempe & Dietrich, 2018; Salve et al., 2012). Rock moisture storage exceeded soil water storage on a seasonally dry hillslope in northern California (Rempe & Dietrich, 2018), while limestone stored approximately twice as much dynamic water as soil in shallow upslope terrain on the Edwards Plateau (Leite et al., 2025). Furthermore, Leite et al. (2025) also found that the dominant storage domain shifted from bedrock to soil as soil thickness increased downslope. These findings demonstrate that the relative contributions of soil and bedrock depend strongly on profile structure and confirm that bedrock water must be explicitly considered when assessing subsurface water-storage capacity in shallow-soil karst landscapes. Although weathered bedrock dominated dynamic storage at both hillslopes, the capacity of this reservoir varied with the degree of weathering. Weathering progressively modifies matrix porosity and fracture architecture, thereby controlling the thickness of the active storage zone and the depth to which infiltrating water can penetrate (Lebedeva & Brantley, 2020; Pedrazas et al., 2021). In this study, dynamic water storage was 1,077.6 mm in SWH and 889.0 mm in WWH, indicating 21% greater storage in the strongly weathered hillslope (Fig. 5). The significant ($p < 0.05$) positive relationship between the CIA and RWS provides further evidence that the volume of water involved in wetting and drying increased with weathering degree (Fig. 10). This relationship is consistent with the more friable and strongly weathered rock cores and the greater vertical geochemical differentiation observed in the SWH (Fig. 1c). Chemical dissolution, mineral weathering, and physical fracturing can increase matrix and fracture porosity, deepen the weathering profile, and expose a larger rock volume to infiltrating water (Goodfellow et al., 2016; Holbrook et al., 2014). Geophysical observations in granitic terrain have similarly identified thick weathered profiles as major potential water reservoirs (Holbrook et al., 2014), while borehole measurements on sedimentary hillslopes have shown that seasonally dynamic rock moisture extends approximately to the depth of pervasive fracturing and oxidation (Pedrazas et al., 2021). At a broader spatial scale, Hahm et al. (2019) found that a deeply weathered critical zone supported substantially greater dynamic water storage than an adjacent, thinly weathered profile under similar climatic conditions. Therefore, stronger weathering expands the volume of bedrock contributing to dynamic profile water storage. This finding indicates that estimates based solely on soil water can substantially underestimate the active subsurface storage capacity of karst hillslopes. Incorporating weathering degree and weathering-profile depth into hydrological models may therefore improve predictions of water retention, drought development, and groundwater recharge under climate change.

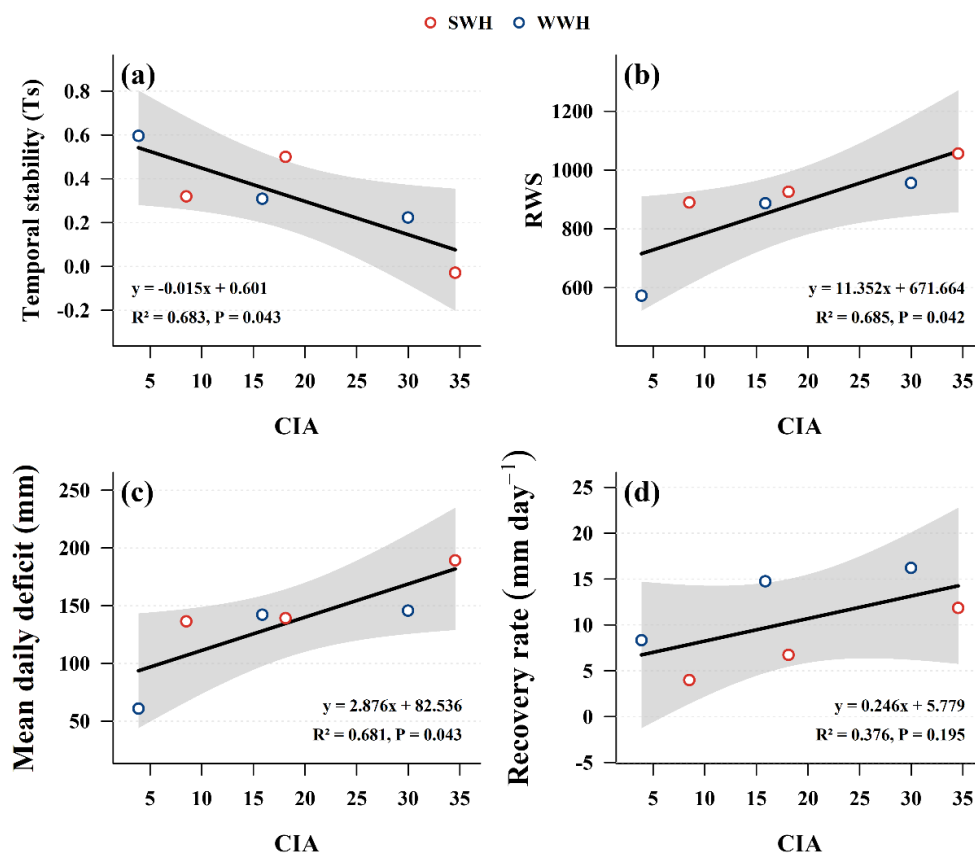


Figure 10: Relationships between bedrock weathering degree and water-storage dynamics. Correlations between the chemical index of alteration (CIA) and (a) temporal stability, (b) relative water storage, (c) mean daily storage deficit, and (d) recovery rate. Red and blue symbols represent the strongly weathered hillslope (SWH) and weakly weathered hillslope (WWH), respectively. Solid lines indicate linear regressions, and shaded bands indicate 95% confidence intervals.

4.2 Greater Storage Capacity Does Not Ensure Greater Storage Persistence

Despite having 21% greater maximum dynamic storage capacity than the weakly weathered hillslope (WWH), the strongly weathered hillslope (SWH) realized only 91.65% of the mean water storage observed on the WWH, indicating that greater storage space did not translate into greater realized storage. This mismatch was further supported by the significant positive correlation between CIA and storage deficit and the significant negative correlation between CIA and temporal stability ($p < 0.05$; Fig. 10). One possible explanation for this capacity–persistence decoupling is that weathering modifies storage space and hydraulic connectivity simultaneously. Chemical alteration and physical fracturing can deepen the weathered zone, increase porosity, and transform fresh bedrock into a thicker assemblage of weathered and fractured materials with greater storage potential (Holbrook et al., 2014; Pedrazas et al., 2021; St. Clair et al., 2015). At longer timescales, however,



455 interactions between subsurface flow and the advancing weathering front may further enhance hydraulic connectivity and
 promote downward drainage, allowing infiltrating water to pass through the profile rather than remain in matrix storage
 (Lachassagne et al., 2021; Lebedeva & Brantley, 2020). This dual effect may be particularly pronounced in karst terrain,
 where increases in carbonate-rock matrix porosity can be accompanied by an order-of-magnitude increase in saturated
 hydraulic conductivity (Leite et al., 2023). The effect of this expanded storage space on water-storage persistence therefore
 460 depends on the hydraulic connection between matrix storage and transmissive pathways. For example, Xiang et al. (2026)
 found that strongly weathered bedrock functioned as a transmissive layer that promoted downward leakage and limited water
 accumulation above the soil–bedrock interface. By contrast, when the connection between storage and transmissive
 pathways is limited, additional water may be retained for extended periods. Evidence for this behavior comes from the
 Edwards Plateau, where rock moisture replenished by major storms persisted for more than two years under limited vertical
 465 fracture connectivity (Leite et al., 2025). These contrasting observations indicate that weathering does not inherently reduce
 storage persistence. Rather, its effect depends on whether the resulting weathered profile is dominated by matrix storage,
 poorly connected fractures, or an efficiently connected drainage network. Because porosity, fracture connectivity, and
 hydraulic conductivity were not measured directly in this study, this mechanism remains a process-based interpretation
 requiring further verification.
 470 In contrast to storage magnitude and temporal stability, storage-deficit recovery rate was not significantly related to CIA (Fig.
 10), indicating that weathering degree alone did not determine how rapidly depleted storage was replenished. Recovery is an
 event-scale process jointly controlled by rainfall characteristics, antecedent wetness, and the hydraulic connectivity among
 storage domains. Hillslope studies have shown that subsurface hydrological responses are often initiated only after rainfall
 exceeds a critical threshold (Lehmann et al., 2007; Tromp-Van Meerveld & McDonnell, 2006). Rainfall depth determines the
 475 amount of water available to overcome this threshold, whereas rainfall intensity influences the rate of wetting-front
 propagation and the activation of preferential pathways (Singh et al., 2021; Xiao et al., 2022). The higher profile recovery
 rates observed during the wet year were therefore likely associated with larger and more intense rainfall events. However,
 the rainfall threshold required for recovery also depends on antecedent soil and bedrock moisture, which partly reflects the
 amount of unfilled storage in the upper profile (Zhang et al., 2026b). Under dry antecedent conditions, incoming rainfall
 480 must first replenish shallow storage deficits before reaching deeper storage domains, whereas wetter antecedent conditions
 lower this threshold and facilitate the development of hydrologically connected flow paths (Detty & Mcguire, 2010; Jencso
 et al., 2009; Penna et al., 2011). Once connected, fractures may rapidly replenish hydraulically accessible bedrock storage
 while simultaneously transmitting water past poorly connected matrix domains or beyond the monitored profile (Hopp &
 McDonnell, 2009; Salve et al., 2012). Therefore, weathering degree exerted a clearer control on the magnitude and
 485 persistence of dynamic storage than on the event-scale rate at which storage deficits recovered. More broadly, these findings
 indicate that the hydrological buffering function of weathered karst bedrock cannot be inferred from storage capacity alone.
 Whether bedrock storage can sustain groundwater discharge during periods of reduced precipitation also depends on how
 persistently water is retained and how strongly storage domains are connected to drainage pathways.



4.3 Implications of Bedrock Water Dynamics for Catchment Hydrology

490 Bedrock water dynamics influence catchment hydrology by controlling how infiltrated precipitation is partitioned among shallow critical-zone storage, recharge to deeper groundwater, lateral subsurface flow, and spring discharge (Ren et al., 2026; Salve et al., 2012; Wang et al., 2020). This partitioning is initially constrained by antecedent storage deficits of the profile, because substantial groundwater recharge and streamflow generally occur only after water deficits in the weathered bedrock have been replenished (Hahm et al., 2022). In our study, the SWH had a 21% greater maximum dynamic storage capacity

495 than the WWH but maintained less water within that capacity, resulting in a larger storage deficit. Consequently, a greater proportion of incoming rainfall was required to replenish shallow bedrock storage before the recharge signal reached the monitored water table. Meanwhile, stronger weathering likely produced a larger and more heterogeneous network of pores and fractures, dispersing infiltrating water among multiple storage domains and flow paths (Delbart et al., 2016). Such distributed infiltration provides a plausible explanation for the longer groundwater response times and peak lags observed at

500 the SWH (Fig. 11b and c). Once these pathways became hydraulically connected, however, their high transmissivity allowed water to drain or be redistributed rapidly beyond the shallow monitoring zone (Lachassagne et al., 2021), as indicated by the faster groundwater recession at the SWH (Fig. 11e). Similar combinations of delayed recharge and efficient drainage have been reported where heterogeneous weathered layers and connected fractures regulate vertical percolation and preferential flow (Ren et al., 2026; Xiang et al., 2026). The groundwater records in this study support the interpretation: the water table

505 remained shallower than the 11-m observation limit for only 29.5%–49.2% of the monitoring period at the SWH, compared with 74.2%–99.8% at the WWH. Thus, the SWH combined greater transient storage potential with shorter shallow-groundwater residence, whereas the WWH retained shallow groundwater more persistently within the monitored profile. Importantly, the rapid recession and frequent decline of the SWH water table below 11 m do not necessarily indicate poor water retention at the catchment scale. Water leaving the monitored profile may recharge deeper groundwater, migrate

510 laterally toward lower topographic positions, or eventually contribute to spring discharge (Wang et al., 2020). Bedrock flow can bypass local monitoring domains and re-emerge farther downslope, thereby sustaining streamflow beyond the hillslope where infiltration initially occurred (Asano et al., 2020; Kosugi et al., 2006). In karst systems, such redistribution between diffuse storage and preferential pathways also governs the timing and magnitude of spring responses (Chang et al., 2019). The perennial spring observed at the SWH is therefore consistent with the transfer of shallow bedrock water into deeper or

515 lateral flow paths, although it does not by itself identify the pathways involved (Wang et al., 2020). Distinguishing among these potential destinations will require deeper groundwater observations, tracer experiments, and continuous measurements of spring discharge. Nevertheless, the combined storage and groundwater evidence suggests that bedrock weathering alters not only how much water a hillslope can store, but also where that water is retained, how long it remains locally available, and how it is subsequently released. Catchment models should therefore represent both weathering-dependent storage

520 capacity and fracture connectivity when predicting groundwater recharge, drainage, and the persistence of spring flow.

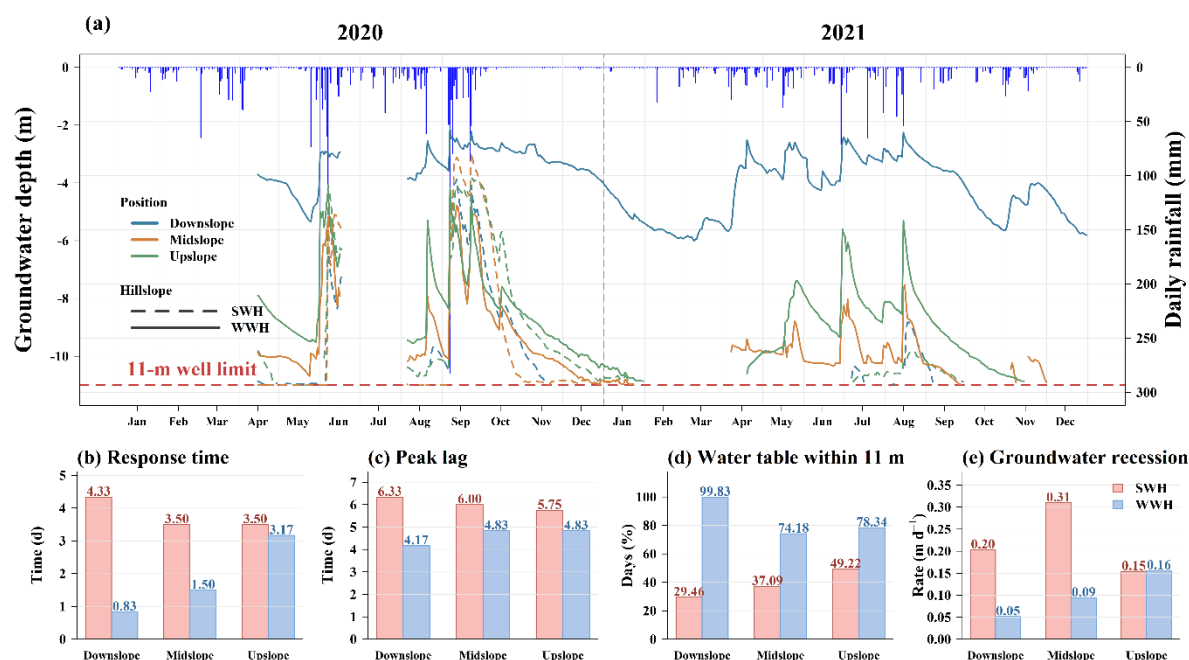


Figure 11: Rainfall-driven groundwater dynamics across the strongly weathered (SWH) and weakly weathered (WWH) hillslopes during 2020–2021. (a) Daily rainfall and groundwater depth at the upslope, midslope, and downslope positions; the red dashed line denotes the 11-m well limit. (b–e) Event response time, peak lag, percentage of monitored days with groundwater shallower than 11 m, and groundwater recession rate, respectively. Groundwater records were missed from 18 June to 5 August 2020.

4.4 Ecological Implications of Shallow Bedrock Water Persistence

The redistribution of bedrock water described above has direct ecological consequences because water can support vegetation only while it remains within, or hydraulically connected to, the rooting zone (McCormick et al., 2021). Although the SWH had greater maximum storage capacity, its rapid drainage shortened the period during which stored water remained available. By contrast, the more persistent shallow storage at the WWH provided a more reliable water supply between rainfall events. Consistent with this hydrological contrast, trees at the WWH were generally taller and had greater DBH than those at the SWH (Table S1). Larger trees typically have greater transpiration demands and may draw on deeper water sources through more extensive root systems, particularly during the dry season (Bleby et al., 2010). Taken together, these hydrological and vegetation observations suggest that vegetation development depends not only on the amount of water that the weathered profile can store, but also on how long that water remains accessible to plant roots. Nevertheless, because tree size also reflects species composition, stand age, nutrient availability, and disturbance history, the observed vegetation pattern provides supporting evidence rather than direct proof of hydrological control.



Although tree size alone cannot establish hydrological control, rooting architecture and water-use traits provide a plausible
 540 mechanism linking water persistence to vegetation development. Karst woody plants exhibit contrasting belowground
 adaptations. Deep-rooted species can access relatively stable water stored in bedrock fractures during drought, whereas
 shallow-rooted species may develop extensive lateral roots to exploit spatially fragmented water sources near the surface
 (Nie et al., 2014). For example, two common woody species in the karst region of Southwest China had maximum rooting
 depths of less than 33 cm but developed widely spreading lateral root systems (Ma et al., 2020). Against this background, the
 545 rapid depletion of shallow water at the SWH would be expected to favor a more conservative water-use strategy, whereas the
 persistent and accessible water at the WWH could permit a more “profligate” strategy. This inference is further supported by
 Liu et al. (2025b). These findings indicate that vegetation responses to bedrock weathering depend on the compatibility
 between subsurface water persistence and plant water-use traits. As climate change intensifies drought and hydroclimatic
 variability, accounting for this coupling will be essential for predicting which species and communities can persist, decline,
 550 or recover in karst landscapes and for selecting restoration species suited to future water regimes.

4.5 Implications, Limitations, and Future Work

The conceptual model developed here challenges the common assumption that stronger weathering necessarily increases
 subsurface water availability by creating greater storage capacity (Fig. 12). Although the SWH had a larger dynamically
 active storage domain, its lower temporal stability, greater wet-to-dry storage decline, and faster groundwater recession
 555 indicate that much of this water was only temporarily retained. Stronger weathering may therefore increase storage capacity
 without providing greater drought resilience. In contrast, the WWH had a smaller dynamic storage domain but maintained
 more persistent profile water and groundwater. Under wet antecedent conditions, persistent storage may leave limited
 capacity to accommodate additional rainfall. The WWH may consequently have greater potential for rapid surface or
 subsurface runoff once local storage approaches saturation. These contrasting responses show that weathering modifies not
 560 only how much water can be stored, but also how long it remains available and how rapidly it is redistributed. This
 reinterpretation carries practical implications for ecosystem management in karst regions. In many areas of southwest China,
 restoration efforts have prioritized enhancing soil thickness and promoting weathering to increase water storage; our results
 suggest that this strategy may not always be effective. Where weathering has progressed to the point of creating highly
 connected fracture networks, the associated increase in storage space may be offset by rapid drainage that limits water
 565 availability during dry periods, whereas areas with moderate weathering and persistent storage may be more suitable for
 sustaining woody vegetation. This does not imply that weathering is inherently detrimental to ecosystem function, but rather
 that the relationship between weathering and water availability is non-monotonic and depends on the balance between
 storage creation and drainage enhancement. However, matrix porosity, fracture architecture, and hydraulic connectivity were
 not measured directly in this study. The interpretation that stronger weathering reduced storage persistence by enhancing
 570 drainage therefore remains a process-based inference from the observed moisture and groundwater dynamics rather than a
 direct verification of its structural controls. Storage capacity, persistence, and hydraulic connectivity should therefore be

considered separately—and in combination—when evaluating drought buffering, runoff generation, and restoration potential in karst critical zones (Hahm et al., 2022; Leite et al., 2025).

The TDR measurements should be interpreted within several methodological constraints. Laboratory calibration showed strong linear responses across the soil–bedrock mixtures ($R^2 \geq 0.96$), supporting the use of TDR for detecting temporal changes in moisture. However, the probes below the soil–bedrock interface were installed in weathered-bedrock powder to ensure hydraulic contact, rather than directly in intact and fractured bedrock. Differences in material composition, pore structure, and probe contact can therefore introduce matrix-specific offsets in the measured volumetric water content (Salve et al., 2012). In addition, the discrete sensors installed at 60-cm intervals capture the dominant profile-scale dynamics but cannot fully represent small-scale variations among matrix pores, fractures, and conduits. The TDR-derived values should consequently be regarded as site-specific estimates of relative water-content change and storage dynamics. The consistent installation and calibration procedures support comparisons between hillslopes, depths, and monitoring periods within this study, but the absolute values should not be directly compared with measurements from other regions without site-specific calibration.

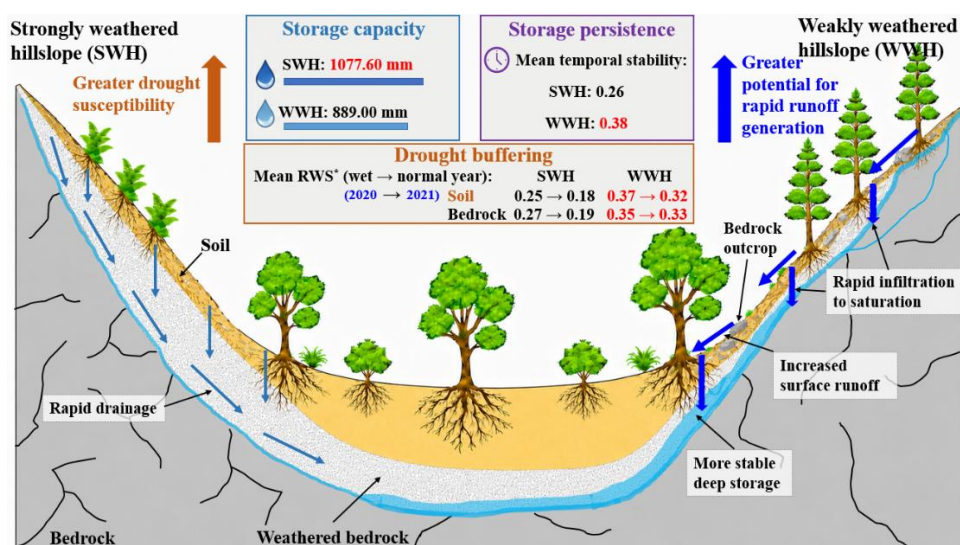


Figure 12: Conceptual model of the weathering-related tradeoff between dynamic storage capacity and drought resilience in karst hillslopes.

Future work should first combine direct characterization of bedrock structure and hydraulic properties with high-frequency moisture and groundwater monitoring. Core-based porosity measurements and borehole imaging could quantify matrix porosity and fracture aperture, density, and orientation, while hydraulic and tracer tests could evaluate fracture connectivity and the exchange between matrix storage and transmissive pathways. These measurements would directly test whether the lower storage persistence of the SWH resulted from stronger hydraulic connections between dynamically active storage domains and drainage pathways. Event-scale analyses could then determine how these structural controls interact with rainfall characteristics and antecedent wetness to regulate response thresholds, wetting-front propagation, storage recovery,



595 and post-event drainage. Concurrent monitoring of hillslope runoff, spring discharge, groundwater levels, and deep
percolation would help determine whether rapid bedrock wetting contributes to local storage, lateral subsurface flow,
groundwater recharge, or flood peaks. Second, stable isotopes of precipitation, soil water, bedrock water, groundwater, and
xylem water should be combined with sap-flow and plant-water-potential measurements to identify the water sources used
by vegetation and evaluate the ecological accessibility of bedrock storage. Finally, replicated observations across broader
600 weathering gradients, lithologies, and climatic regions would be valuable for testing whether the relationships among storage
capacity, persistence, and hydrological risk identified here are transferable beyond the two studied hillslopes.

5 Conclusions

Two years of continuous monitoring across two adjacent humid karst hillslopes showed that weathered bedrock, rather than
shallow soil, was the dominant dynamic water-storage domain. Although the strongly weathered hillslope (SWH) had 21%
605 greater maximum dynamic storage than the weakly weathered hillslope (WWH), it maintained only 91.65% of the mean
profile storage observed at the WWH and exhibited lower temporal stability, more frequent low-storage conditions, and
greater depletion as precipitation declined. Groundwater observations reinforced this contrast: the SWH experienced more
transient recharge and faster recession, whereas the WWH maintained profile water and shallow groundwater more
persistently. These observations indicate that stronger weathering can expand the volume of bedrock involved in water
610 storage while simultaneously promoting drainage through hydraulically connected pathways, thereby limiting the duration
for which stored water remains locally available. The more developed woody vegetation on the WWH was consistent with
this more reliable water supply, although direct plant uptake of bedrock water was not measured. Collectively, our findings
challenge the assumption that greater weathering and storage capacity necessarily enhance water availability or drought
resilience. Instead, the hydrological and ecological functions of weathered bedrock depend on the balance among storage
615 capacity, storage persistence, and hydraulic connectivity, all of which should be considered when predicting drought
propagation, groundwater recharge, runoff generation, and vegetation responses in karst critical zones.

Data Availability Statement

The data generated in this study are available from the authors upon request, and can also be accessed via the following link:
<https://data.mendeley.com/datasets/k7tgwwtwcj/1>.



620 **Author contributions**

Conceptualization: Hongsong Chen, Jun Zhang, Fa Wang; Data curation: Yutong Zhang, Jun Zhang, Fa Wang; Funding acquisition: Hongsong Chen, Jun Zhang, Jinjiao Lian; Investigation: Jun Zhang, Fa Wang, Huimin Ren; Methodology: Yutong Zhang, Jun Zhang, Fa Wang; Supervision: Hongsong Chen, Jun Zhang, Fa Wang.

Competing interests

625 The authors declare there are no conflicts of interest for this manuscript.

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