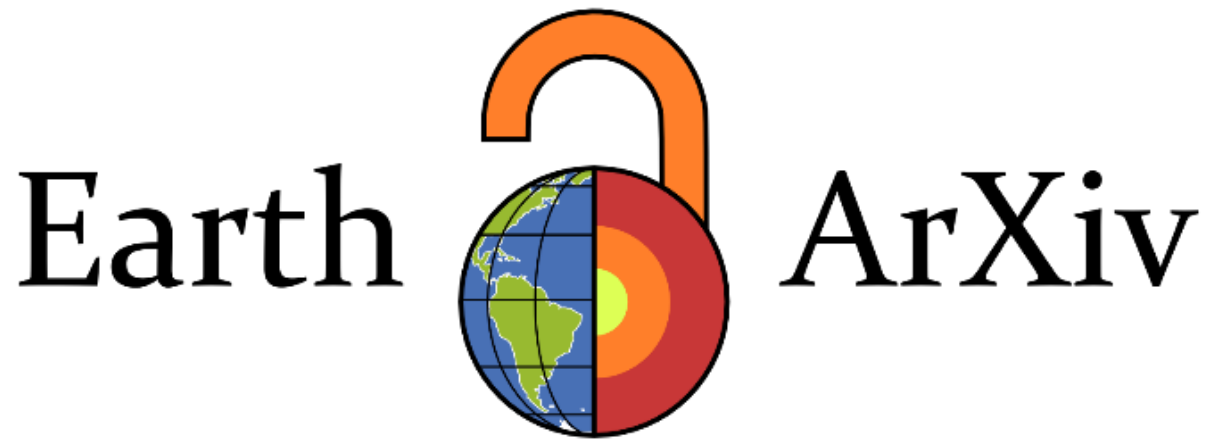




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Earthquakes prime landscapes for typhoon-induced permeability loss

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Abstract

Bedrock aquifers supply drinking water to half the global population, and earthquakes can disrupt this resource for years by changing their permeability. Previous studies proposed several mechanisms to explain coseismic permeability changes but lacked an observational tool to distinguish them; moreover, all assumed an immediate onset. Here, we analyse groundwater and seismic data

from eastern Taiwan capturing the responses of the aquifer and the surrounding rock substrate to the 2024 M7.2 Hualien earthquake. Tracking substrate and aquifer responses separately enables us to distinguish permeability changes driven by mechanisms confined to the aquifer from those affecting the entire subsurface. We observe two distinct ways in which earthquakes modify aquifer permeability: coseismic shaking produces immediate permeability increases that co-evolve with substrate changes. Months later, two consecutive typhoons triggered abrupt permeability decreases — an observation unique to the postseismic period. As the substrate remains unaffected by these typhoons, we attribute this response to the combined effect of the earthquake, which increased sediment availability and enhanced infiltration capacity, and typhoon-induced overland flow, delivering the sediment into the aquifer, clogging fractures. Our results demonstrate that large earthquakes can prime landscapes for delayed hydrological disruption, with potentially long-lasting consequences for groundwater supply and landslide susceptibility.

Keywords: Earthquake hydrology, postseismic recovery, permeability changes, groundwater recovery

1 Introduction

Bedrock groundwater is the largest terrestrial freshwater reservoir [1] that provides drinking water to ~ 50 % of the global population [2]. Perturbations to this system can have dramatic socioeconomic and ecological consequences [3] and may alter landscape dynamics, for example by impacting landslide distributions [4]. Among these, earthquake-driven perturbations have been observed for thousands of years [5]. They appear, for example, as sudden changes in river discharge [6], the appearance or disappearance of springs [7], or abrupt changes in water composition [8]. At depth, groundwater level disturbances occur step-like as recharge or discharge events that can cause excess discharge on the order of cubic kilometres [9] or in the form of changes in recharge and drainage dynamics that can last for years before they recover [10] and, thus, disturb this crucial resource for a long time. Such perturbations have been attributed to changes in aquifer permeability [11].

Previous studies have invoked a variety of mechanisms for coseismic permeability changes, such as static strain changes [12], the sudden consolidation of material leading to increased pore pressure [13], comparable to liquefaction, modifications in aquifer confinement [14, 15], or the clearing and clogging of pore spaces and fractures [16]. However, direct observations of the mechanisms remain limited. While tidal transmissivity monitoring [17] has been useful in identifying coseismic permeability changes, it is unable to distinguish between the mechanisms triggering the change. For example, the commonly invoked clogging and clearing cannot be distinguished from rock-mechanical changes, i.e., changes in the mechanical state of the rock matrix itself, including the geometry of the fracture network. Furthermore, while sediment concentrations in rivers increased following large earthquakes [18], it was unclear whether and when the sediments reach the aquifers and, ultimately, clog them. Various studies

have independently observed coseismic disturbances followed by multiannual recovery in the elastic properties of the rock substrate [19–21] and in the aquifer [11, 22, 23]. Yet we do not know whether substrate and permeability changes respond to earthquakes independently or in a coupled fashion, which could be addressed through a joint study of the two components.

Here, we investigate the convergence and divergence of time-dependent observations of aquifer permeability and substrate changes in response to large earthquakes. This enables us to distinguish mechanisms associated with substrate changes, which would manifest in both records simultaneously, from mechanisms confined to the aquifer, such as fracture clogging and clearing. Specifically, we compare the time-dependent transmissivity [24], which is proportional to the permeability, with changes in the seismic wavespeed [25].

1.1 Taroko National Park and the 2024 Hualien earthquake

We analyse data from Taroko National Park in eastern Taiwan (see Fig. 1), which provides a dense network of groundwater wells and seismic stations from the coast to the high mountains (> 3000 m.a.s.l.). Eastern Taiwan sits on a subduction zone that produces a steep topography, allowing for fast sediment mobilisation, and causes regular large earthquakes. Furthermore, Taiwan experiences regular typhoons during the summer monsoon, causing intense rainfall and flooding that can quickly mobilise large amounts of sediment. In April 2024, within our study period (2021–2025), an M7.2 megathrust earthquake occurred just offshore the city of Hualien at a depth of ~ 35 km. The Hualien earthquake is the largest seismic event recorded around Taiwan in the last quarter century, causing thousands of coseismic landslides [26] and perturbing Taroko National Park’s fractured aquifers and the hydrological system as a whole. Less than four months later, Taiwan experienced a typhoon season that included one of Taiwan’s most intense typhoons in recent history, Kong-Rey [27], delivering more than 1000 mm of precipitation to our field site. Together, this extraordinary combination of events and the independent monitoring of both systems allow us to test whether coseismic permeability changes and substrate modifications are coupled and identify the environmental conditions necessary to trigger fracture clogging.

2 Results

2.1 The impact of earthquakes on the raw water level

Groundwater data from four boreholes show coseismic changes (see Fig. 2), illustrating the range of drainage dynamics and coseismic responses. Prior to the M7.2 Hualien earthquake, the water level in all four wells was gradually decreasing. All wells record a sudden step increase coincident with the earthquake, followed by a decay on the order of 6 hours in all wells but EHW-03. In the following months, two wells — EHW-03 and EH-04 — return to their pre-earthquake declining trend, whereas EHW-01 and EHW-02 record a steeper decrease than before the earthquake. After the 2022 M6.9 Taitung earthquake, we find more subtle groundwater changes. Of these four wells, EHW-02 yields robust transmissivity estimates (see supplementary discussion).

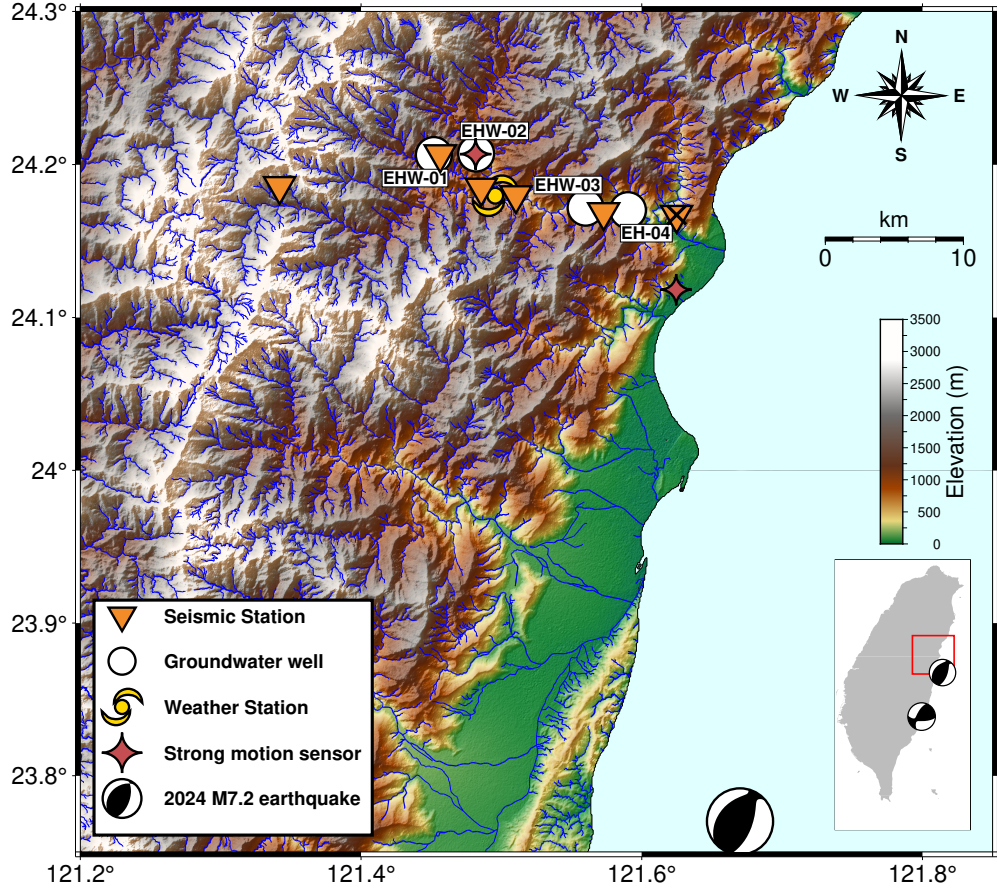


Fig. 1 Locations of sensors, from which data were used. Data from the seismic station marked with a cross (TQ01) have been used to produce Fig. 3. We used data from well EHW-02 co-located with the strong motion sensor F072 to perform the tidal analysis. The remaining three wells were used only for the groundwater compensation of $\delta v/v$. The map inset contains the moment tensor solutions for the M7.2 2024 Hualien and M6.9 2022 Taitung (outside the map limits) earthquakes.

Hence, the raw data already demonstrate that large seismic disturbances are followed by a period of pore pressure readjustment; we discuss the causes of this adjustment later in this article.

2.2 Time-dependent transmissivity

We obtain the time-dependent transmissivity from July 2021 to November 2025 by fitting the semidiurnal tidal response in the piezometric head to the solid Earth tides and barometric fluctuations. This analysis yields time-dependent transfer functions that we invert for the transmissivity, assuming a leaky-aquifer model (see 4.2 for details). Independent pump tests, performed in 2013 for depth ranges within the wells (see supplementary Table S1), provide a reference order of magnitude for the transmissivity.

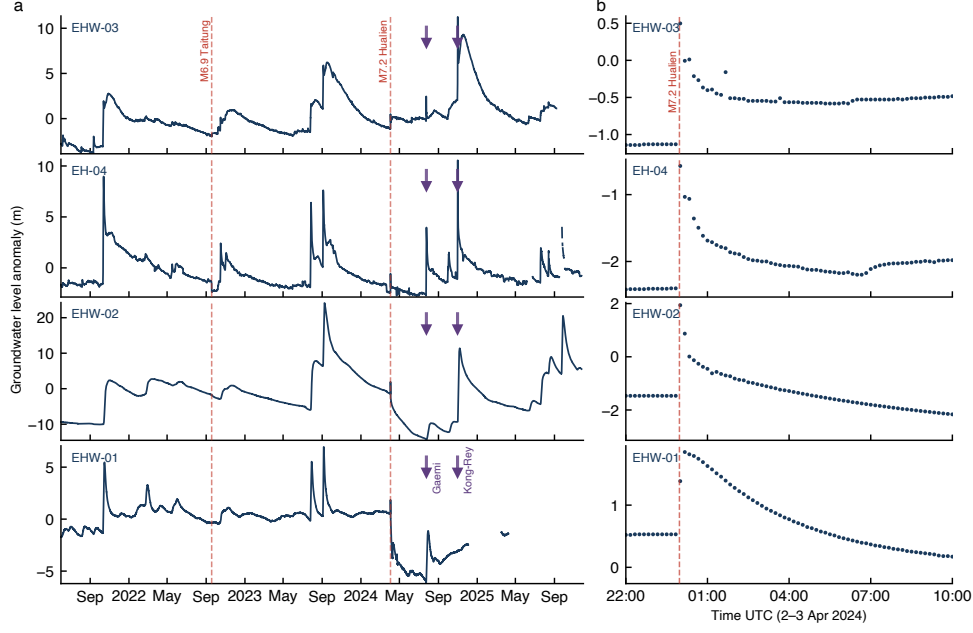


Fig. 2 Piezometric groundwater level records from Taroko National Park. Panel **a** shows water level records for the study period from four wells. The water level exhibits step-like changes and alterations in drainage dynamics at the marked M6.9 Taitung and M7.2 Hualien earthquake sequences. Panel **b** shows a zoom-in on panel **a** of the hours following the 2024 M7.2 Hualien earthquake. For the tidal analysis, we use data from EHW-02.

Until April 2024, the transmissivity deviates less than 20 % from its median value of $\sim 1.1 \times 10^{-6} \text{ m}^2/\text{s}$. Pump tests indicate a somewhat higher value of $\sim 1.7 \times 10^{-6} \text{ m}^2/\text{s}$ (see Fig. 3) that is sufficient to validate the order of magnitude of the values retrieved from our tidal analysis. Small variability in the transmissivity can often be explained by rising (falling) groundwater, which leads to lower (higher) effective stresses and thereby causes fracture apertures to increase (decrease) [28]. To emphasise co- and postseismic changes, we correct for this pore pressure effect (see 4.2.1), yielding a pore pressure-corrected transmissivity time series.

After pore pressure correction, the transmissivity exhibits several abrupt excursions. Two of these changes are transmissivity increases that coincide with large regional earthquakes — the 2024 M7.2 Hualien earthquake and the 2022 M6.9 Taitung earthquake sequence. Most prominently, following the Hualien earthquake, which caused ground-shaking of 0.5 g, the pore pressure-corrected transmissivity increased by about 80 %. For about two weeks after the earthquake, our transmissivity estimate exhibits large uncertainties, which may be due to high-frequency noise in the water level record. Therefore, we choose not to interpret the transmissivity’s immediate response to the M7.2 earthquake. Over the rest of the study period, the pore pressure-corrected transmissivity gradually decreases, superimposed with two step decreases.

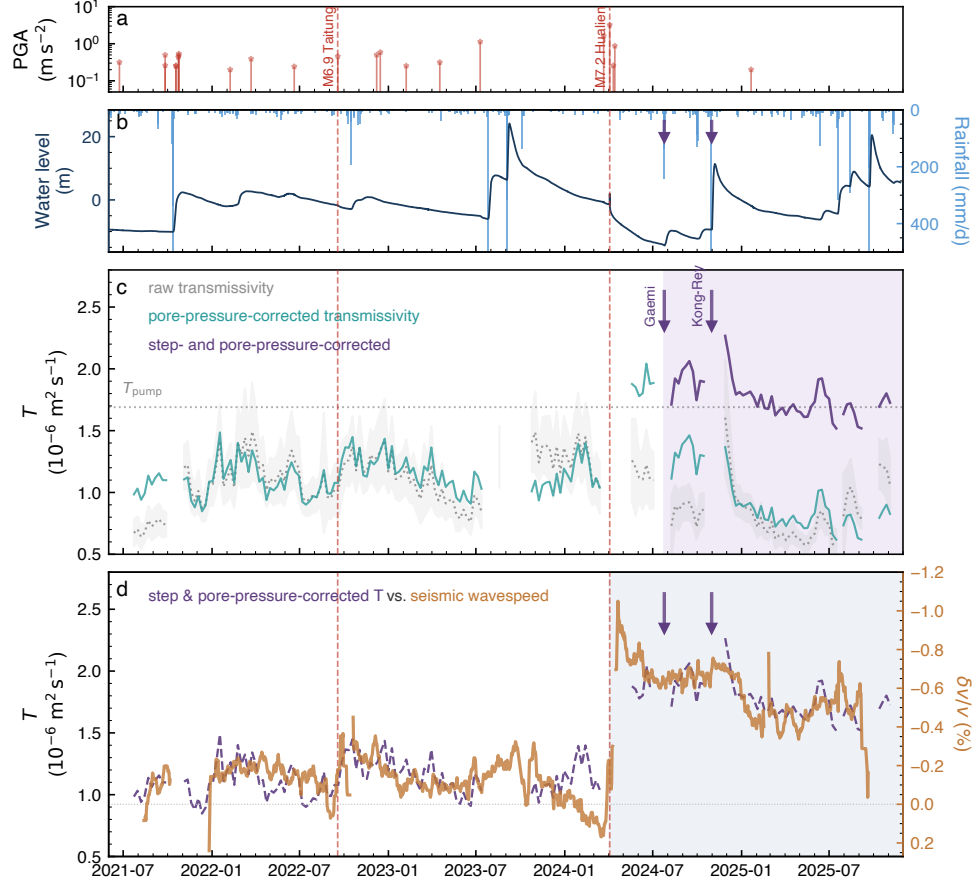


Fig. 3 Co-evolution of the transmissivity and the seismic wavespeed ($\delta v/v$). **a** Peak ground accelerations (PGAs) measured for regional earthquakes $M > 4.5$. **b** Piezometric water level records measured at EHW-02 and daily rainfall at the automatic weather station COT820. **c** From tidal inversion, we retrieve the raw transmissivity (dotted grey curve) and its $\pm\sigma$ error envelopes. For comparison, we show the transmissivity estimated from hydraulic pump tests T_{pump} . We apply a pore pressure correction (mint curve) and remove steps introduced by the typhoons Gaemi and Kong-Rey (purple curve). **d** The low-frequency seismic wavespeed co-evolves with the fully corrected transmissivity. The shading in panel c emphasises the typhoon-induced, lasting transmissivity decrease, while panel d is shaded for times following the Hualien earthquake. We computed $\delta v/v$ from data in the frequency band 0.5-1 Hz from station TQ01. We exclude $\delta v/v$ values with a correlation coefficient of $R_{av} < 0.25$ and transmissivity values with standard deviations $> 5 \times 10^{-7} \text{ m}^2/\text{s}$. Note that the $\delta v/v$ axis is inverted.

A more subtle transmissivity increase occurs with the September 2022 Taitung earthquake sequence, which had magnitudes of M6.5 and M6.9 and occurred $\sim 100 \text{ km}$ south of our network at a shallow depth of around 10 km. After those events, the transmissivity recovers transiently for about 10 months to pre-earthquake levels.

Two additional transmissivity decreases occur well separated from any earthquakes within the time series. Specifically, four months after the Hualien earthquake, the transmissivity drops by $\sim 55\%$ coincident with the landfall of the typhoon, Gaemi. The raw transmissivity even reaches values below its pre-earthquake baseline. Another $\sim 25\%$ decrease coincides with typhoon Kong-Rey, three months later. However, we find no imprint on the transmissivity from typhoons before the 2024 Hualien earthquake, although some delivered significantly more rainfall than Gaemi (see Fig. 4). The two transmissivity decreases do not recover within the study period.

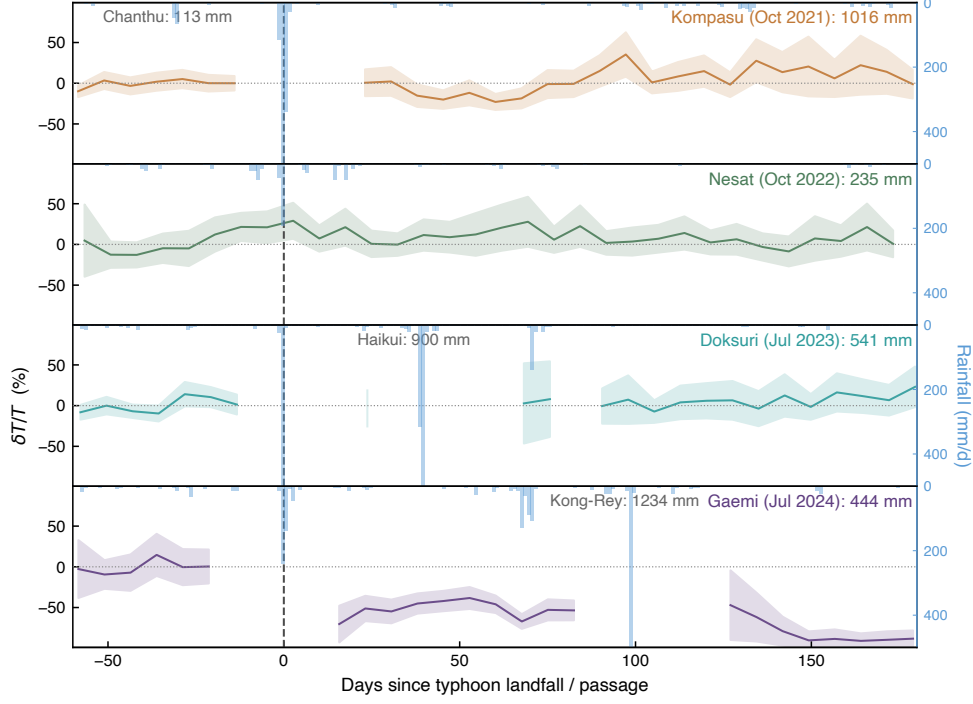


Fig. 4 Transmissivity response to major typhoons. The relative pore pressure-corrected transmissivity ($\delta T/T$) only shows a statistically significant change in response to the Gaemi and Kong-Rey typhoons, which occur three and seven months after the 2024 M7.2 Hualien earthquake. For all other typhoons, including Nesat that follows the significantly smaller Taitung earthquakes, no change is detectable. For $\delta T/T$, we show the $\pm\sigma$ error envelopes. We exclude $\delta T/T$ values with $\sigma > 50\%$. For each typhoon, we indicate the total rainfall ± 3 days from its occurrence. Data gaps result from increased noise level in the tidal inversion due to barometric fluctuations and strong water level changes.

2.3 The seismic wavespeed co-evolves with the transmissivity

As a measure of the solid substrate's state, we compare the transmissivity to seismic wavespeed changes ($\delta v/v$). To obtain $\delta v/v$ estimates for several depth ranges, we

compute $\delta v/v$ from several frequency bands, where higher frequencies result in $\delta v/v$ estimates for shallower depths [29]. High-frequency $\delta v/v$ is anticorrelated to water level, which is expected from the pressure-dependency of elastic properties [e.g., 30–32]. Here, we aim to analyse properties of the substrate independently from water level variations in the aquifer. Therefore, we focus the analysis on low-frequency (0.5–1 Hz) $\delta v/v$ (Fig. 3d) that shows only minimal impacts of water level fluctuations (see supporting Figs. S4–5).

Considering co- and postseismic behaviour, the seismic wavespeed and the transmissivity show remarkable similarities: The only coseismic transmissivity changes coincide with the only coseismic wavespeed changes. Furthermore, transmissivity increases are not only accompanied by decreases in $\delta v/v$, but the postseismic recovery to a base value also occurs on similar time scales. For the weaker Taitung sequence, $\delta v/v$ and the transmissivity recover after about 10 months. For the 2024 Hualien earthquake, the recovery is still ongoing at the end of our time series but follows the same trajectory for $\delta v/v$ and the pore pressure and typhoon-corrected transmissivity. This co-evolution holds across all stations and for depth ranges $>\sim 200$ m (see supplementary Figs. S6 & S7). Shallower depths show residuals of groundwater signals — even after correction.

In contrast to the transmissivity, the seismic wavespeed does not show any changes unrelated to pore pressure changes following the 2024 Gaemi and Kong-Rey typhoons. From the parallels and contrasts in the response of the seismic wavespeed and the transmissivity to earthquakes and typhoons, we can constrain the mechanism driving each kind of permeability change.

3 Discussion

3.1 Earthquakes and typhoons induced two kinds of transmissivity changes

After removing the pore pressure effect from the transmissivity, we find two kinds of sudden transmissivity changes: Coseismic transmissivity changes and typhoon-triggered transmissivity changes. Because the seismic wavespeed is only impacted by the earthquakes and direct pore pressure effects, and because the postseismic recovery transient of the transmissivity remains unaffected by the 2024 typhoons, we conclude that these two kinds of transmissivity changes are fundamentally different. From their symptomatic differences and, crucially, from the observation that the typhoons do not disrupt the transmissivity’s postseismic recovery trajectory, we can make some inferences about the underlying mechanisms.

The 2024 typhoon season

There is a sharp contrast in the response of the aquifer to typhoons before and after the Hualien earthquake: typhoons typically do not impact the transmissivity beyond their direct effect on the water level, but typhoons occurring in the months following the earthquake produce lasting transmissivity decreases (Fig. 4). We do not find a change in the aquifer’s response to rainfall following any other earthquake, for example the

2022 Taitung sequence, raising the question of how the Hualien event differed from preceding earthquakes. Compared to the larger Hualien earthquake, all other earthquakes produced substantially less coseismic landsliding [33, 34], which can drive transmissivity changes in two ways: increased effective stress from sediments redeposited by typhoon-induced flooding or fracture clogging by sediment particles [16]. Two observations are inconsistent with increased effective stress by redeposited sediment as the driving mechanism. First, we estimate that the added sediment thickness to produce only the first typhoon-induced transmissivity drop is at least ~ 5 m (see 4.2.2). By contrast, Taiwan’s largest typhoon in recent history caused a net erosion of less than 20 cm [35]. Second, effective stress modifications impact the seismic wavespeed. However, for the 2024 typhoons, we do not observe wavespeed changes unrelated to pore pressure changes. Furthermore, we would expect a notable transmissivity recovery as erosion gradually removes deposited sediments. We can thus rule out effective stress modifications as the mechanism for the 2024 typhoon-induced transmissivity changes.

Fracture clogging by sediment particles [16], capable of decreasing transmissivity, may have been enabled by the combination of increased sediment availability through coseismic landsliding and overland flow capable of delivering sediments into the aquifer. Because this clogging affects only a small part of the volume that the seismic waves are sensitive to and the effect of clogging on the elastic properties of the bulk substrate is likely negligible in general, it does not impact the seismic wavespeed. This is consistent with the absence of any $\delta v/v$ signal unrelated to pore pressure during either typhoon, across all stations and frequency bands. In Taiwan, [18] observed increased sediment-concentrations in rivers during typhoons that followed the 1999 Chi-Chi earthquake. That sediments delivered to rivers by typhoons are subsequently flushed into aquifers is supported by our transmissivity observations, likely facilitated by the large pressure gradients created during the flooding that enable infiltration within hours to days [36, 37]. Large earthquakes prime this mechanism in two complementary ways: They destabilise steep mountain terrain and thereby deliver sediment from the hillslopes [38, 39], and they generate macrofractures that increase the infiltration capacity of the landscape [40]. Smaller earthquakes occurring before the Hualien earthquake likely provided insufficient sediment in our study area to facilitate fracture clogging. The typhoon-triggered transmissivity decreases show no sign of recovery, indicating that the clogging of the millimetric-to-centimetric fractures that control permeability [41] may persist for years.

Co- and postseismic transmissivity changes

Coseismic transmissivity increases, as we observe for the 2022 Taitung and 2024 Hualien earthquakes, differ in several ways from typhoon-induced transmissivity drops. First, in contrast to typhoon-induced changes, earthquake-driven transmissivity changes show signs of recovery. This recovery trajectory remains unbroken by the 2024 typhoon season. Second, the coseismic transmissivity increases are accompanied by coseismic wavespeed decreases. Recovering coseismic transmissivity increases could be linked to coseismic static stress changes [42], the unclogging or clearing of fractures by seismic wave passage [11, 16], or to changes in fracture geometry [28].

Coseismic static stress changes have generally been discarded as a major cause of seismic wavespeed changes, as they cannot explain their spatial pervasiveness, their sign consistency (almost invariably negative), or their logarithmic recovery [20, 21]. Moreover, the Taitung earthquake would not have caused significant static stress changes at our site due to its relatively large epicentral distance.

Clearing of fractures by seismic shaking, similar to the mechanism we invoked for the typhoon-induced transmissivity change, causes transmissivity to increase. If clearing of fractures was the dominant driver of the coseismic transmissivity increase, its postseismic recovery would correspond to gradual fracture clogging. Therefore, we would expect an interaction between the typhoon-induced fracture clogging and the postseismic recovery, disrupting or even resetting the postseismic recovery trajectory. Assuming that the aquifer recovers to a steady-state, we may even expect a transmissivity-increasing post-typhoon recovery. Instead, we observe an unbroken postseismic transmissivity recovery, which indicates that coseismic and typhoon-triggered transmissivity changes stem from two unrelated processes. Furthermore, as clearing or clogging of pore spaces does not impact the seismic wavespeed, the parallel post-seismic recovery of the seismic wavespeed and the transmissivity would have to be coincidental. For those reasons, we consider it unlikely that clearing of pore spaces was the main driver of the coseismic transmissivity increases.

The remaining candidate is a rock-mechanical change in pore space geometry that simultaneously affects the seismic wavespeed and the permeability. The seismic wavespeed is controlled by grain contact area, i.e., the area of contact within micro- and macrofractures [43]. The permeability, on the other hand, depends mostly on the aperture of these fractures [44]. Some processes, such as clogging of fractures, reduce the effective hydraulic aperture [44] but do not affect the contact area. However, in some cases, for example, when changes in effective stress open or close fractures, these properties may covary. Because permeability scales with the cube of fracture aperture [44] while seismic wavespeed changes are proportional to grain contact area [43], a geometric change that widens fractures while reducing contact area produces permeability changes that far exceed wavespeed changes — consistent with the observation. Following the coseismic change, contact area and fracture aperture recover on the same trajectory. Seismic wavespeed decreases and their recovery are an expression of the non-classical nonlinear response of materials to stress changes, in which grain contacts are disrupted and recover logarithmically [45–47]. The co-evolution of transmissivity and seismic wavespeed, and their parallel postseismic recovery, indicate that coseismic changes in fracture aperture and grain contact area are coupled, and that both recover on the timescale predicted by nonlinear nonclassical elasticity. This interpretation remains a hypothesis, but it is the only single-mechanism explanation consistent with all our observations.

3.2 The seismic wavespeed as a diagnostic tool for aquifer changes

Earthquakes with similar intensity do not necessarily affect the aquifer equally. In contrast to findings by [11], the peak ground acceleration (PGA) is not a reliable measure for the amplitude or even the presence of a transmissivity change. In particular, we

find earthquakes producing higher PGAs than the 2022 Taitung earthquake that are not associated with discernible transmissivity changes, consistent with observations by [23, 24]. For our data, we find that coseismic changes in seismic wavespeed are a good indicator for coseismic transmissivity modifications: every coseismic transmissivity change is accompanied by a coseismic variation in wavespeed and vice versa. This likely reflects differences in spatial sensitivities: the PGA is a point measure that is heavily influenced by local topography and source directionality [48–50], whereas the seismic wavespeed integrates changes over larger volumes [29]. When the underlying transmissivity change is rock-mechanical, the seismic wavespeed may serve as a proxy for postseismic aquifer recovery. However, when clogging or clearing dominates, as for the 2024 typhoons, the seismic wavespeed does not respond. If continuous groundwater observations are available in such cases, the seismic wavespeed can become a discriminator of the underlying transmissivity-changing mechanism rather than a proxy for its amplitude.

3.3 Implications for postseismic water availability

Here, we have identified two ways in which large earthquakes may impact the water availability for years to come: through rock-mechanical changes to the fracture network that show signs of recovery, and through priming the landscape for long-lasting fracture clogging. Because the transmissivity’s postseismic transient follows a similar trajectory to that of the seismic wavespeed, we can use observations of seismic wavespeed recovery to infer expected recovery times of rock-mechanical transmissivity changes. For the seismic wavespeed, recovery has been observed to take several months to over a decade [e.g., 19, 21, 51–53]. This implies that groundwater levels and flow rates will take similarly long to return to pre-earthquake baselines [as independently estimated by 11, 54].

While previous studies have invoked fracture clogging as a coseismic mechanism [10, 16, 22], we demonstrate that it manifests as an abrupt transmissivity drop, months after the earthquake, when sediments made available by coseismic landsliding are flushed into the aquifers. For the following year, there is no sign of recovery from these clogging-induced permeability decreases in our data. When, whether, and how aquifers will be unclogged remain important open questions, although earthquake-induced shaking [11, 16] or flushing by steep pressure gradients due to intense rainfall could potentially clear fractures. Long-lasting groundwater perturbations are consistent with observed impacts on vegetation health in the years following large earthquakes [55–57].

In steep landscapes like Taroko, transmissivity changes are unlikely to be spatially uniform. Topographic amplification focuses earthquake damage on mountain ridges [48–50], inducing stronger transmissivity changes and enhancing the known effect of postseismic redistribution of water from higher to lower elevations [58]. Storm-driven clogging, on the other hand, will preferentially take place in valleys, where topography concentrates overland flow and sediment-laden flooding [18], further amplifying the difference between valley and ridge transmissivities. Such effects are most consequential at convergent plate boundaries, where steep landscapes and megathrust earthquakes combine, altering permeability distributions and, thereby, pore pressures within the landscape for years, with implications for landslide susceptibility [39, 59].

4 Data and Methods

4.1 Groundwater records

We obtain water level records from four wells: EH-04, EHW-03, EHW-01, and EHW-02. Based on the lithological information, transfer function characteristics, pump test results and the drainage and recharge behaviour, we classify EHW-03 and EH-04 as unconfined wells. Their barometric transfer functions exhibit near-zero barometric efficiency, and tidal transfer functions derived from their piezometric records have broad optima leading to large uncertainties (i.e., $> 20^\circ$ for the tidal phase), precluding reliable tidal inversion. EHW-01 samples a confined but very low-permeability formation ($T \approx 1 \times 10^{-8}$), for which semidiurnal tidal forcing is insufficient to drive measurable flow from the aquifer into the borehole within a half-tidal period. For these three wells, we only use their piezometric records to correct $\delta v/v$ for water-level-induced fluctuations (see 4.4.2). EHW-02, which samples a semi-confined fractured aquifer, delivers robust tidal transfer functions.

Geologically, eastern Taiwan is dominated by intense tectonic deformation resulting in an extraordinarily steep topography and a lithology dominated by metamorphic rocks. In particular, schists, marbles, and quartzites are prevalent. EHW-02 was drilled through a sequence of fractured schists overlain by ≈ 17 m of colluvium. The resulting borehole was cased and closed to isolate it from atmospheric influences. The groundwater level is screened between 97 and 100 m depth at a 10 min sampling interval and archived as groundwater level above sea level in metres. To obtain the groundwater level anomaly time series shown in the manuscript, we demean the data.

4.2 Computing transmissivity time series from well records, solid Earth tides, and barometric records

To obtain the transmissivity time series, we closely follow the procedure proposed by [24], who extended earlier work by [17] and [60] to compute such time series from tidal areal strain time series, barometric records, and groundwater level records for semi-confined aquifers. The algorithm consists of a two-step procedure: First, the tidal and barometric transfer functions of the groundwater level are computed for narrow frequency bands. Thereafter, one can invert for the aquifer’s transmissivities $T(t)$ and shear moduli $G(t)$ that minimise the misfit to the observed transfer functions. This inversion requires knowledge of some well specifications and hydraulic parameters that can be derived from pump tests. Aside from assuming a constant vertical diffusivity D' for the leaky aquitard, we follow the procedure described in [24].

As previous authors, we focus our analysis on tidal modes around the semi-diurnal tidal bands, including the M2, N2, and S2 modes, which separate more cleanly from harmonic noise than diurnal tides [11, 17, 24, 60]. To avoid introducing high-frequency noise into our analysis, we detrend and taper the data before applying a simple spike and step removal that removes water level changes above 2.5 cm per 10 min. Then, we bandpass filter the piezometric time series between 1.6 and 2.4 cycles per day using a 4th-order zero-phase Butterworth filter, effectively removing harmonic noise outside

our band of analysis. To decrease the computational cost of the remaining steps, we decimate the water level and tidal data to a 60 min sampling interval.

To obtain the tidal and barometric transfer functions (i.e., time-dependent functions for barometric and tidal gain and phase lag), we apply a harmonic least-squares fit to the areal solid Earth tide strain and the barometric pressures and the bandpass-filtered water level; thereby, we obtain complex functions containing amplitude and phase for each of the three frequencies. As in [23], we assume each frequency component k of the piezometric record to be the linear sum of the barometric and the tidal forcing:

$$\forall k \in [1, n], \tilde{h}_k = \frac{1}{\rho g} (Z_{a,k} \tilde{P}_k + Z_{\epsilon,k} \tilde{\epsilon}_k), \quad (1)$$

where $\tilde{h}$, $\tilde{P}$, and $\tilde{\epsilon}$ are the complex amplitudes of the piezometric water level, the atmospheric pressure, and the tidal strain, respectively. We set $\rho = 1000 \frac{\text{kg}}{\text{m}^3}$ and $g = 9.81 \frac{\text{m}}{\text{s}^2}$. To solve for the barometric and tidal transfer functions Z_a and Z_ϵ , we need to extract the values for at least two frequencies. It further requires us to assume that the frequency dependency between those frequencies is negligible [61]. Here, we solve for three quasi-semidiurnal frequencies (M2, N2, and S2), allowing us to create an overconstrained system of linear equations, to which we apply a complex least-squares solve, thereby obtaining additional uncertainty estimates:

$$\begin{aligned} \rho g \tilde{h}_1 &= Z_a \tilde{P}_1 + Z_\epsilon \tilde{\epsilon}_1 \\ \rho g \tilde{h}_2 &= Z_a \tilde{P}_2 + Z_\epsilon \tilde{\epsilon}_2 \\ \rho g \tilde{h}_3 &= Z_a \tilde{P}_3 + Z_\epsilon \tilde{\epsilon}_3 \end{aligned} \quad (2)$$

We solve this equation for time steps of 29.5 days (i.e., one lunar month) and in increments of 7.5 days. To refine our uncertainty estimates, we bootstrap the result 2000 times. We show the resulting transfer functions $Z_a(t)$ and $Z_\epsilon(t)$ in supplementary Fig. S2.

Finally, to obtain $T(t)$ from the estimated transfer functions, we use the solution for a two-layered model by [24] that requires us to solve two partial differential equations for the atmospheric pressure and the tidal response:

$$T \nabla^2 h + q = S \left(\frac{\partial h}{\partial t} - \frac{\gamma}{\rho g} \frac{\partial P}{\partial t} \right), \quad (3)$$

$$T \nabla^2 h + q = S \left(\frac{\partial h}{\partial t} - \frac{2\gamma G}{\rho g} \frac{\partial \epsilon}{\partial t} \right), \quad (4)$$

where q is the vertical leakage flow, G the aquifer's shear modulus, γ its uniaxial skempton coefficient or loading efficiency, T the transmissivity, and S the storativity. The full solution is provided in [24]. In contrast to them, we solve the equation for a time-invariant vertical diffusivity $\frac{dD'}{dt} = 0$, which reduces the number of unknowns and thereby decreases the noise level in the inversion procedure. We characterise the upper layer as a leaky aquitard with a thickness of 20 m and a storativity of $S' = 8 \times 10^{-4}$, and the lower layer as an aquifer with a storativity of $S = 7.2 \times 10^{-5}$ (see 4.3).

Furthermore, the algorithm requires us to define the well’s screen and casing radii, which are $r_w = r_c = 0.0254$ m.

As we do not have any direct measurements of the uniaxial Skempton coefficients γ and γ' and the hydraulic diffusivity D' , we calibrate them directly from the transfer functions. To this end, we exploit the dependency of the loading efficiency ratio $R = \frac{1-\gamma'}{1-\gamma}$, required to calibrate our model, on the median phase difference between tidal and barometric transfer functions $\tilde{\Delta}\phi \approx 19^\circ$ and the vertical diffusivity D' (as described in appendix B2 of [24]). For a given D' , we find the R that produces this phase difference analytically. We independently assess numerical stability by checking the fraction of time steps at which the inversion pins at its search bounds across a range of D' - R pairs, preferring pairs that minimise this fraction. This procedure yields $R \approx 0.25$ and $D' = 2.12 \times 10^{-3} \frac{\text{m}^2}{\text{s}}$, agreeing to first order with independent pump test results (see 4.3). Finally, we invert for $T(t)$ from the tidal phase response. The tidal amplitude gain constrains the shear modulus $G(t)$, which we do not interpret for reasons laid out in 4.2.1.

4.2.1 Postprocessing the transmissivity

Removing the water level dependency in $T(t)$

We find a subtle water level dependency for $T(t)$. Higher pore pressures lead to an extension of existing pores and cracks, increasing their permeability. This dependency can be corrected as demonstrated by [24]. We do so by fitting our observations of transmissivity and effective stress changes to a model relating the two [28]:

$$\left(\frac{T}{T_0}\right)^{1/3} = 1 - \alpha \ln \frac{\sigma^*}{\sigma_0^*}, \quad (5)$$

where σ^* is the effective stress, T_0 and σ_0^* are the mean transmissivity and mean effective stress, respectively. α is the proportionality coefficient that we find through a least-squares misfit optimisation in a simple gridsearch process. To avoid noise introduced by co- and postseismic transmissivity changes, we only consider observations before the Hualien earthquake. The resulting correction and fit are shown in supplementary Fig. S3.

Step removal in $T(t)$

Coinciding with the Gaemi typhoon, we find a step decrease in T (see 2.2) of $-6 \times 10^{-7} \text{m}^2/\text{s}$. For Kong-Rey, the corresponding step is $-3 \times 10^{-7} \text{m}^2/\text{s}$. To facilitate the comparison of the postseismic transient behaviours in $\delta v/v$ and T , we remove these steps simply by subtracting them from the time series following each typhoon’s landfall.

Water level dependency in $G(t)$

We find a strong positive correlation between the tidal transfer function’s gain and the piezometric water level, which carries through into the inferred $G(t)$ (see supplementary Figs. S2 and S3). As structural shear modulus changes and pore pressure-induced

gain changes are in practice indistinguishable, we chose not to interpret $G(t)$. $T(t)$ is unaffected as it is derived solely from the tidal phase response.

4.2.2 Estimating sediment load changes

After finding α for Eq. 5, we can simply infer the amount of effective stress required to produce the observed typhoon-triggered transmissivity changes from the fitted model. As a lower-bound estimate, we assume undrained conditions, i.e. we set the skempton coefficient to 0. The thickness of newly deposited sediment required to produce the step change introduced only by typhoon Gaemi ($-6 \times 10^{-7} \text{ m}^2/\text{s}$) is 4.8 m, assuming a mass density of 1800 kg/m^3 .

4.3 Aquifer and aquitard properties inferred from pump tests

The Taiwan Geological Survey performed hydraulic pump tests at our borehole site (see supplementary Table S1). These tests yielded values for the specific storativity S_s and the hydraulic conductivity K at various depth ranges. We infer storativity values for aquifer and aquitard and transmissivity values for the aquifer from these values. Specifically, we obtain $T_{PT}^{87-100 \text{ m}} = K\Delta h = 1.3 \times 10^{-7} \text{ m/s} \times 13 \text{ m} = 1.69 \times 10^{-6} \text{ m}^2/\text{s}$ assuming an isotropic hydraulic conductivity. We plot this value as reference in Fig. 3. We directly take $S = 7.2 \times 10^{-5}$ from the depth range of 28 to 100 m. For the aquitard, we choose the geometric mean, $S' = 8 \times 10^{-4}$, computed from the S values derived from the intervals 30-31.5 m with $S_s = 1.1 \times 10^{-5} \text{ m}^{-1}$ and 39-40.5 m depth with $S_s = 1.4 \times 10^{-4} \text{ m}^{-1}$, yielding $S' = S'_s \times h' = 2.2 \times 10^{-4}$ and $S' = 2.8 \times 10^{-3}$, respectively, assuming an aquitard thickness of $h' = 20 \text{ m}$. We can also use these values to validate our choice for D' as $D' = K'/S'_s$ [62], again assuming an isotropic hydraulic conductivity. Using this relation, we obtain values spanning between $7.86 \times 10^{-5} \text{ m}^2/\text{s}$ and $2.45 \times 10^{-3} \text{ m}^2/\text{s}$, depending on the chosen depth range - placing our chosen $D' = 2.12 \times 10^{-3} \text{ m}^2/\text{s}$ at the upper end of the range. Generally, pump test values vary by around one order of magnitude, depending on the depths from which they were derived, leaving substantial room for data uncertainty.

4.4 Obtaining seismic wavespeed changes from ambient seismic noise

To obtain the $\delta v/v$ time series, we follow the stretching method as proposed by [25] and implemented in the SeisMIC software package [63]. During preprocessing, we taper, detrend, anti-alias filter, and decimate the data to 4, 8, or 20 Hz, depending on the octave frequency band in which we subsequently bandpass filter. Our chosen frequency bands, 0.5-1 Hz, 1-2 Hz, 2-4 Hz, 4-8 Hz, and 8-16 Hz, provide us with depth sensitivities of tens of metres (16 Hz) down to 1-2 km (0.5 Hz) [29]. Finally, we impose one-bit-normalisation (i.e., we discard the data's amplitude information) and whiten their amplitude spectra [64].

The actual algorithm to retrieve relative seismic wavespeed changes consists of a two-step procedure: First, one correlates continuous seismic data from time t_n to remove the source-side effect, extracting empirical Green's functions (EGF). Thereafter, the EGF from t_n is stretched against a reference to obtain a homogeneous

velocity change at t_n (see [63] for a detailed description of the algorithm). In practice, we cross-correlate hourly seismic data recorded on the three directional components (vertical, north, east), thereby obtaining three cross-component correlations or EGFs per station (E-N, E-Z, and N-Z). To estimate $\delta v/v(t)$, we compare 2-week-long averages of those EGFs with stretched versions of the temporal average of all EGFs, which acts as our reference, corresponding to 0 % velocity change. Specifically, we compare the early coda of the EGFs between lag times of $\tau_s = 5/T_0$ and $\tau_e = 10/T_0$, where T_0 is the low-corner period of the applied bandpass filter. This early coda is dominated by surface waves, for which our depth estimates are valid [29]. We find the wavespeed change for each day by stretching the reference by up to 5 % and identifying the stretching factor producing the highest correlation for each daily smoothed EGF. We then apply a weighted average to the three $\delta v/v$ estimates from a station [65], obtaining one $\delta v/v$ curve per station with a one-day sampling interval. For representation purposes, we roll the results so that the 1st February 2024 has 0 % wavespeed change.

4.4.1 Quality controlling $\delta v/v$

In eastern Taiwan, extreme weather events lead to flooding, sliding, and slumping that affect the sensors. As a result, stations may fail, or their ground coupling may deteriorate. Generally, we disregard stations without available data around the M7.2 Hualien earthquake or whose data quality does not satisfy the threshold described below. As a result, we discard six stations from our analysis: TQ00, TQ03, TQ04, TQ05, TQ07, and TQ11 (none of which is plotted in Fig. 1).

The quality of a $\delta v/v$ estimate can be assessed using the correlation coefficient between the stretched reference and the daily correlation function. After averaging the three components, we obtain an average correlation coefficient that is inherently lower than the individual values but still represents a reliable measure for the stability of each $\delta v/v(t_n)$ [65]. After inspection, we opt to discard $\delta v/v$ time series that exhibit significant portions of data points with $R_{av} < 0.4$. As a result, we consider only some frequency bands for some stations.

4.4.2 Water level compensation

To remove the influence of groundwater fluctuations on $\delta v/v$, we compute linear fitting coefficients for a simple least-squares fit between each $\delta v/v$ and the water level curve with a soft L1 loss criterion, where the loss function is defined as $\rho(z) = 2(\sqrt{1+z} - 1)$ and z corresponds to the least-squares loss [66]. We estimate the linear fitting coefficients for all $\delta v/v$ points with $R_{av} > 0.25$ and the period before the Hualien earthquake to avoid the influence of the coseismic wavespeed drop and recovery. Finally, the water-level-corrected version of $\delta v/v$ is then:

$$(\delta v/v)_{corr}(t) = (\delta v/v)(t) - (\delta v/v)(t)_{lqm}, \quad (6)$$

where $(\delta v/v)_{lqm}$ is the modelled pore pressure-induced seismic wavespeed change.

4.5 Peak ground acceleration

To compute the PGA, we download data from stations F072 and F026 from the Taiwan Strong Motion Instrumentation Program Network (TSMIP), a network of force-balance accelerometers. Data are downloaded for the five minutes following the origin time of each seismic event with $M \geq 4.5$, hypocentral depth $z \leq 40$ km, longitude $120^\circ \leq \lambda \leq 121.5^\circ$, and a latitude of $22.5^\circ \leq \phi \leq 25.5^\circ$. From the resulting traces, we remove the instrumental response and determine the envelope’s maximum for each of the instrument’s three components. We define the maximum value for each event as the event’s PGA. For events after 07 April 2022, we consider the PGA from F072, which is co-located with well EHW-02 but has no available data before this date. We do not plot PGAs below a threshold of 0.2 m/s^2 .

Supplementary information. This article is accompanied by a supplementary file in PDF format, containing additional figures and a discussion about Taroko’s aquifers.

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Declarations

4.6 Data availability

The seismic data from the TQ network [67] used to compute the seismic velocity change time series can be downloaded from the Geofon FDSN service [68]. Piezometric water levels for the wells are openly available with a one-day-sample interval from the Taiwan hydrogeological information system at <https://hydro.geologycloud.tw/>. The barometric and precipitation data for weather station C0T820 were obtained from Taiwan’s Central Weather Administration (CWA) at <https://codis.cwa.gov.tw>. To compute the PGA, we obtained data from stations F072 and F026 from the Taiwan Strong Motion Instrumentation Program (TSMIP) network, provided by the CWA at <https://gdms.cwa.gov.tw>. The digital elevation models used in the map plots are obtained from the Shuttle Radar Topography Mission [70]. Locations of rivers and streams shown in Fig. 1 are copyrighted by OpenStreetMap contributors and available from <https://www.openstreetmap.org>, while we obtained the moment tensors from the global CMT project [71] at www.globalcmt.org.

4.7 Code availability

The research codes required to reproduce this work are available on GitHub. We computed the solid Earth tide predictions using the PyGTide software [72]. Seismic velocity changes were estimated using the SeisMIC software suite [63]. We plotted maps with PyGMT [73]. Moreover, we heavily relied on the Python community packages NumPy [74], SciPy [66], and Matplotlib [75].

4.8 Author contribution

Conceptualization: PM, NB, LI, NH
Data Curation: PM, WAC, JMC
Formal Analysis: PM
Investigation: JMT, WSC, JMC, WAC, WYC
Methodology: PM, NB, LI
Software: PM
Supervision: NH, JMT
Visualization: PM
Writing — Original Draft: PM
Writing — Review & Editing: PM, JMT, NB, NH, LI

4.9 Competing interests

The authors declare no competing interests.

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Supplementary material for: Earthquakes prime landscapes for typhoon-induced permeability loss

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1 Supplementary discussion: Taroko's aquifers

All four boreholes, from which we obtain data, are drilled into fractured metamorphic bedrock typical of eastern Taiwan. However, their hydraulic properties vary substantially, impacting their water level dynamics and suitability for tidal analysis. EHW-03

and EH-04 are drilled through unconfined aquifers hosted in marble and gneiss, respectively, with high transmissivities that lead to rapid recharge, apparent in their raw water level records (see Fig. 1). Their tidal transfer functions yield poorly constrained phase estimates, precluding a tidal inversion. EH-04 is located adjacent to the Liwu river and, due to its high hydraulic conductivity, its piezometric water level closely tracks river stage, explaining the absence of persistent postseismic water level anomalies at this well. EHW-01 is confined within a low-permeability schist ($T \approx 1 \cdot 10^{-8} \frac{\text{m}^2}{\text{s}}$). Its transmissivity is so low that diffusion within the aquifer is slower than a half-period of the semi-diurnal tide, yielding no exploitable tidal signal. EHW-02 samples a semi-confined fractured aquifer isolated from the surface by ~ 17 m of colluvium acting as a leaky aquitard. Its intermediate transmissivity, $\sim 1 \cdot 10^{-6} \frac{\text{m}^2}{\text{s}}$, allows pore pressure diffusion within the tidal period while confinement suppresses atmospheric noise, making it the only well suited for our tidal inversion.

2 Supplementary figures

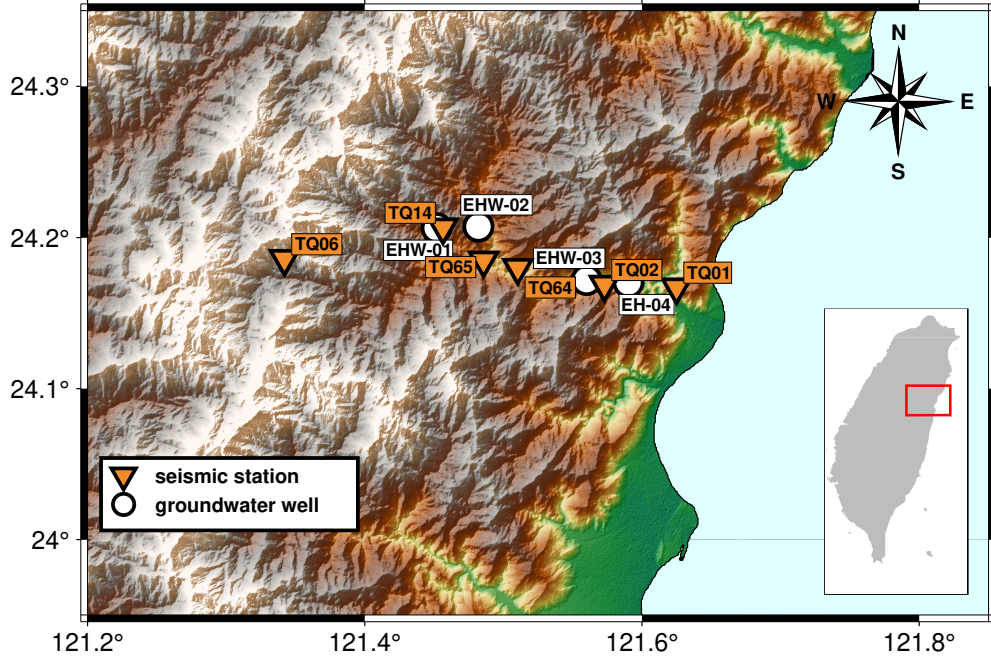


Fig. S1 Location and code of seismic stations and groundwater wells that we used for this supplement.

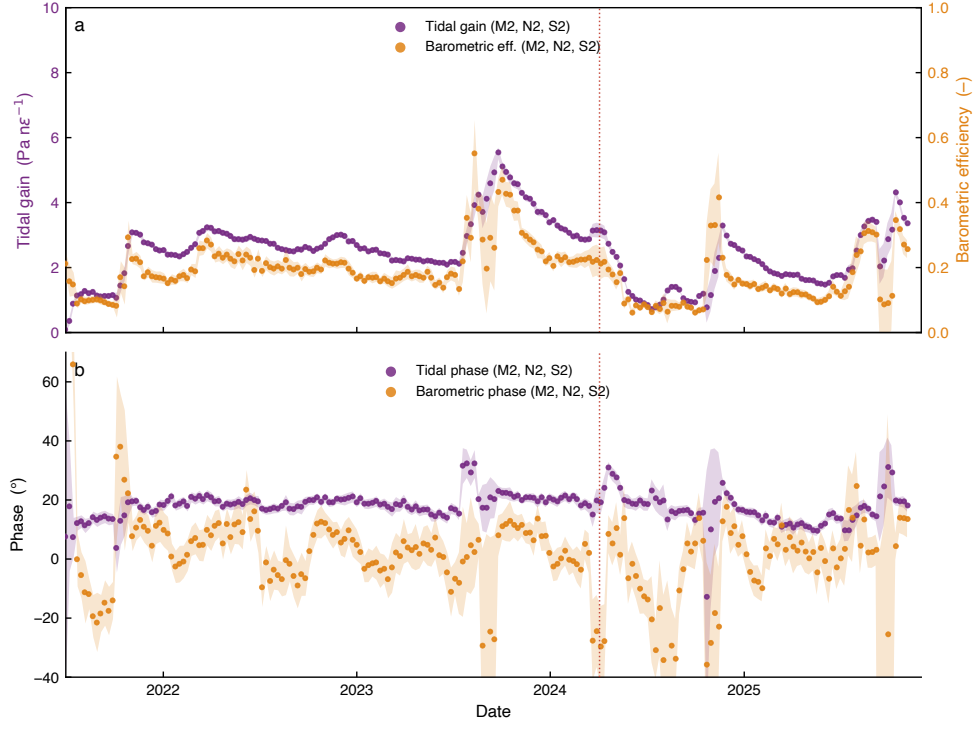


Fig. S2 Tidal and barometric transfer functions. The first part of the transmissivity inversion yields tidal and barometric gain (a) and phase (b) as a function of time. Both show a dependency on the groundwater level; this dependency is particularly apparent for the gain. The dashed red vertical line demarcates the M7.2 2024 Hualien earthquake.

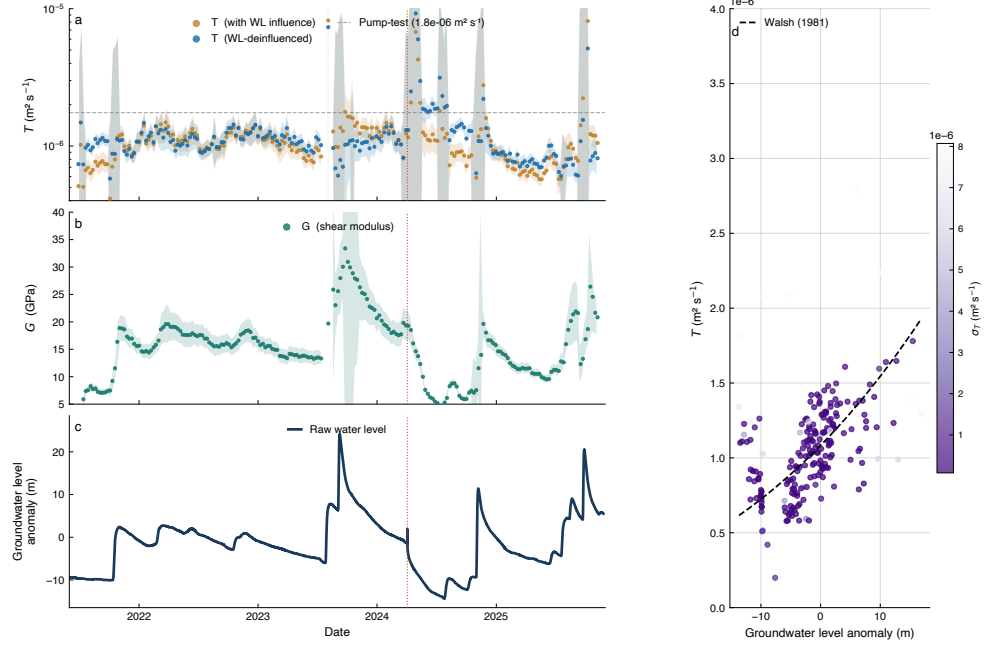


Fig. S3 Time-dependent transmissivity and shear modulus. The second part of the transmissivity inversion yields the aquifer's transmissivity $T(t)$ (a) and shear modulus $G(t)$ (b). Due to its strong water level dependency (shown in c), we do not interpret $G(t)$. In panel a, we compare T retrieved from hydraulic pump tests with the pore-pressure-compensated transmissivity. The dashed red vertical line demarcates the M7.2 2024 Hualien earthquake. Panel c shows the transmissivity as a function of the groundwater level anomaly up to the Hualien earthquake, with the best-fit curve for Eq. 5 derived from [1] superimposed as a yellow dashed line. The points are shaded according to their standard deviation.

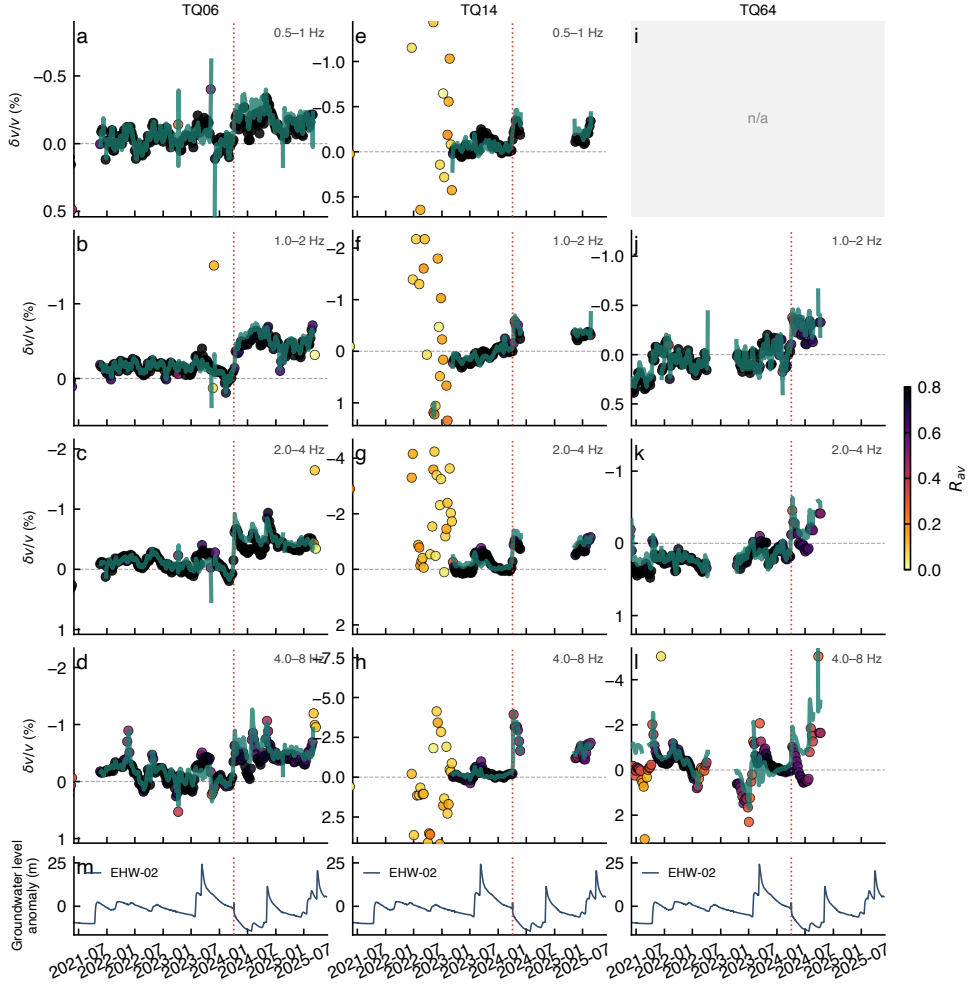


Fig. S4 Pore pressure correction of $\delta v/v$. We show $\delta v/v$ computed from the frequency bands and stations indicated in each panel. The pore-pressure-corrected $(\delta v/v)_{corr}$ is plotted in green, whereas the raw $\delta v/v$ is shown as a scatter plot with its colour depending on the average cross-correlation R_{av} as a quality metric. Note that lower frequency bands show lower pore pressure dependency. The performance of the pore-pressure correction varies from station to station. Panels marked with "n/a" are $\delta v/v$ time series that did not pass the quality control criteria.

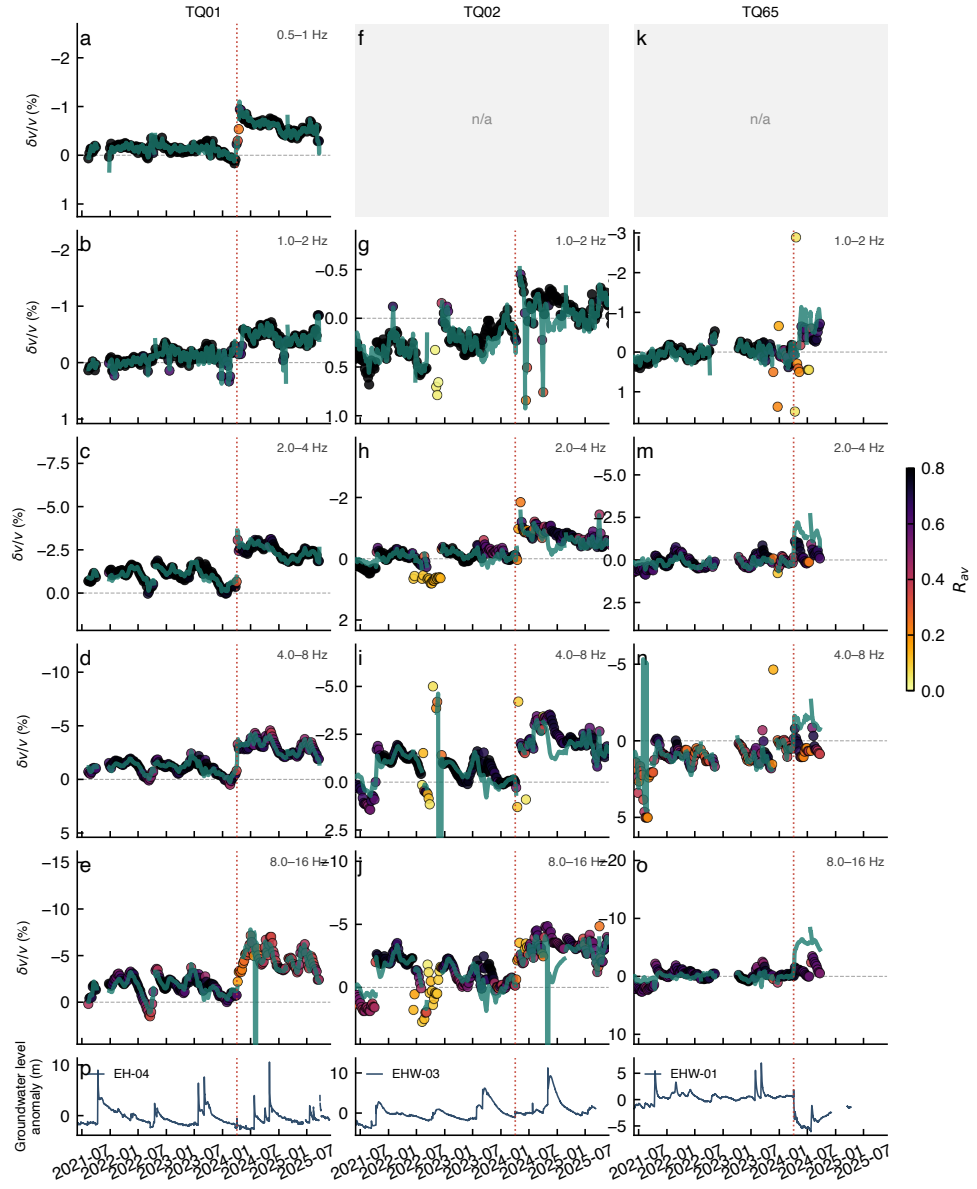


Fig. S5 Pore pressure correction of $\delta v/v$. Refer to Fig. S4's caption.

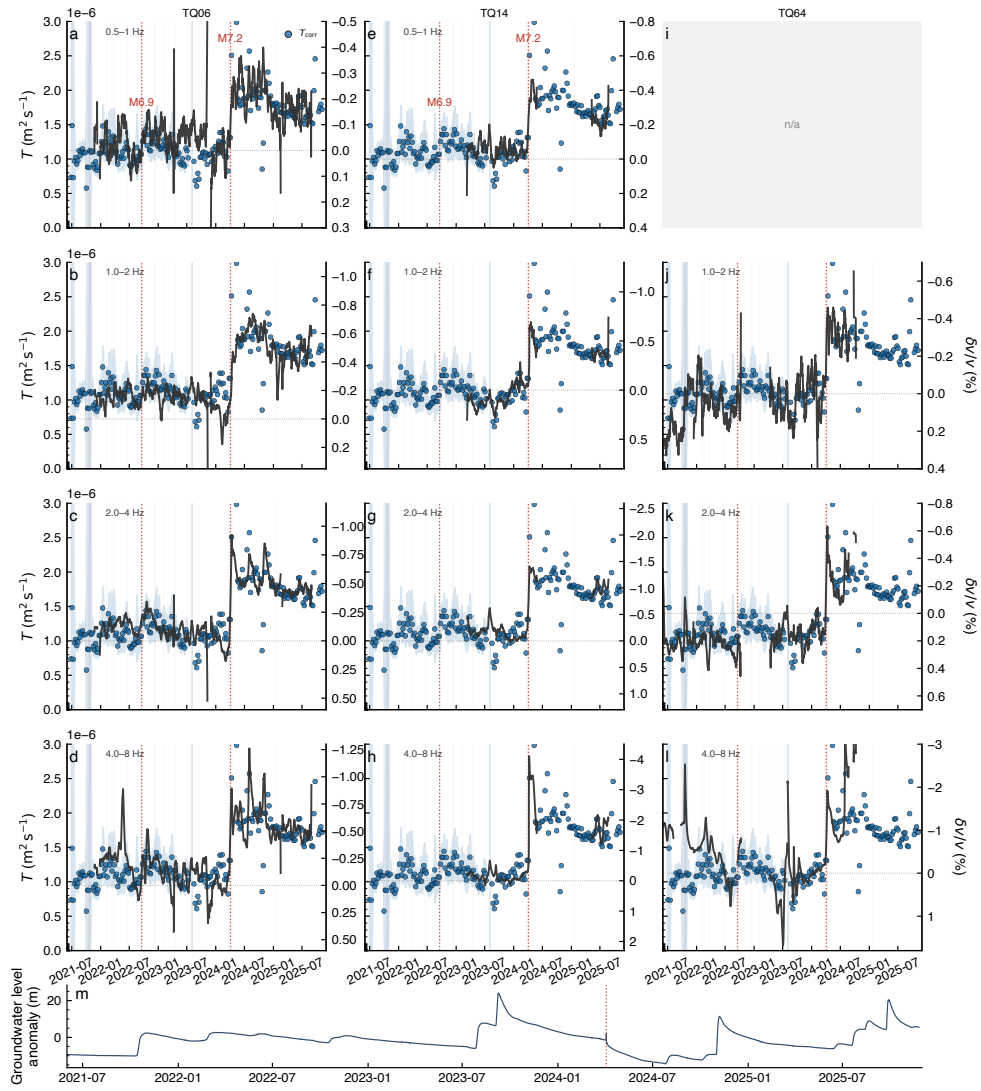
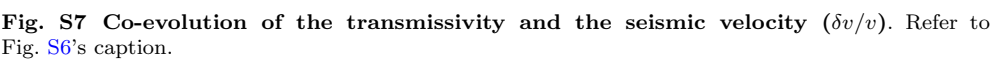


Fig. S6 Co-evolution of the transmissivity and the seismic velocity ($\delta v/v$). We show the pore pressure and typhoon corrected transmissivity (blue circles) against $\delta v/v$ (black solid curve) computed from the frequency bands and stations indicated in each panel. Panels marked with "n/a" are $\delta v/v$ time series that did not pass the quality control criteria. Note that lower frequencies show a clear covariance between $-\delta v/v$ and T . Origin times of the 2022 Taitung and 2024 Hualien earthquakes are indicated by red dotted lines. Panel p shows the piezometric record of EHW-02.



3 Supplementary tables

Table S1 Hydraulic pump test results for borehole EHW-02. We use the storativity values to parametrise the transmissivity inversion, whereas the hydraulic conductivity K serves purely as a validation metric.

$z_{\min}$ (m)	$z_{\max}$ (m)	K (m s ⁻¹)	S	S_s (m ⁻¹)
5.00	10.00	$4.50 \cdot 10^{-6}$	—	—
11.00	16.00	$2.80 \cdot 10^{-5}$	—	—
30.00	31.50	$2.70 \cdot 10^{-8}$	$1.70 \cdot 10^{-5}$	$1.10 \cdot 10^{-5}$
39.00	40.50	$2.90 \cdot 10^{-8}$	$2.20 \cdot 10^{-4}$	$1.40 \cdot 10^{-4}$
52.50	54.00	$2.20 \cdot 10^{-8}$	$4.20 \cdot 10^{-4}$	$2.80 \cdot 10^{-4}$
54.00	100.00	$4.00 \cdot 10^{-7}$	$4.60 \cdot 10^{-5}$	$1.00 \cdot 10^{-6}$
60.00	100.00	$6.10 \cdot 10^{-8}$	$4.00 \cdot 10^{-5}$	$1.00 \cdot 10^{-6}$
64.40	65.90	$2.80 \cdot 10^{-8}$	$1.10 \cdot 10^{-3}$	$7.40 \cdot 10^{-4}$
71.80	73.30	$6.70 \cdot 10^{-8}$	$2.00 \cdot 10^{-4}$	$1.40 \cdot 10^{-4}$
87.00	100.00	$1.30 \cdot 10^{-7}$	$1.30 \cdot 10^{-5}$	$1.00 \cdot 10^{-6}$
87.00	88.50	$4.40 \cdot 10^{-7}$	$1.90 \cdot 10^{-6}$	$1.30 \cdot 10^{-6}$
28.00	100.00	$3.10 \cdot 10^{-6}$	$7.20 \cdot 10^{-5}$	$1.00 \cdot 10^{-6}$

References

- [1] Walsh, J. B. Effect of pore pressure and confining pressure on fracture permeability. *International Journal of Rock Mechanics and Mining Sciences & Geomechanics Abstracts* **18**, 429–435 (1981).