Permeability and elastic properties of rocks from the northern Hikurangi margin: Implications for slow-slip events

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Abstract

Fluid flow and pore-pressure cycling are believed to control slow slip events (SSEs), such as those that frequently occur at the northern Hikurangi margin (HM) of New Zealand. To better understand fluid flow in the forearc system, we examined the relationship between elastic properties, compaction, porosity, and permeability of Cretaceous-to-Pliocene sedimentary rocks from the Raukumara peninsula. We found that the permeability of the deep wedge is too low to drain fluids, but fracturing increases permeability by orders of magnitude, making fracturing key for fluid flow. In weeks to months, plastic deformation and clay swelling heal the fractures, restoring the initial permeability. We conclude that overpressures at the northern HM might partly dissipate during SSEs due to enhanced permeability near faults. However, in the weeks to months following an SSE, healing in the prism will lower permeability, forcing pore pressure to rise and a new SSE to occur.

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Key Points:

- Elastic properties, plastic deformation, and permeability of northern Hikurangi margin rocks
- Permeability-porosity relationship in accretionary prisms
- Clay swelling and plastic deformation controls permeability healing, providing a mechanism justifying slow-slip event cyclicity

Abstract

Fluid flow and pore-pressure cycling are believed to control slow slip events (SSEs), such as those that frequently occur at the northern Hikurangi margin (HM) of New Zealand. To better understand fluid flow in the forearc system, we examined the relationship between elastic properties, compaction, porosity, and permeability of Cretaceous-to-Pliocene sedimentary rocks from the Raukumara peninsula. We found that the permeability of the deep wedge is too low to drain fluids, but fracturing increases permeability by orders of magnitude, making fracturing key for fluid flow. In weeks to months, plastic deformation and clay swelling heal the fractures, restoring the initial permeability. We conclude that overpressures at the northern HM might partly dissipate during SSEs due to enhanced permeability near faults. However, in the weeks to months following an SSE, healing in the prism will lower permeability, forcing pore pressure to rise and a new SSE to occur.

Plain Language Summary

Earth's crust is composed of many tectonic plates fitting together like jigsaw puzzle pieces. Tectonic plates subduct in the mantle along active converging margins, where the forces driving such a convergence can trigger large earthquakes. However, these subduction zones often deform without producing earthquakes, but through slow-slip. The Hikurangi Margin (HM) of New Zealand is a well-studied subduction zone, producing both earthquakes and slow-slip events. The northern HM exhibits more frequent and shallower slow-slip events than the southern margin. Understanding what controls such differences can help improve the general understanding of subduction zone fault mechanics and earthquakes. One of the hypotheses is that the differences between the deformation of the northern and southern HM are controlled by the pore pressure at depth. We tested the elastic and fluid-transport properties of four samples from the northern HM and found that the overriding plate, if not fractured, would be impermeable to fluids. We also tested a fractured sample and observed efficient healing that resets the initial permeability. We conclude that fracturing the overriding plate is fundamental to draining the fluids carried at depth by the subducting plate, and slow-slip events may create new pathways for fluids to escape to the seafloor.

1 Introduction

At the shallow (<15 km depth) portion of the plate interface of subduction zones, scientists have found that convergence between the tectonic plates is often accommodated by modes of slip in between fast earthquakes and aseismic creep (Saffer & Wallace, 2015). Slow-slip events (SSEs) represent one class of such transient phenomena, which can lead to several centimeters of slip over several days to months (Schwartz & Rokosky, 2007). The relatively large seismic moment released by shallow SSEs, comparable to that of earthquakes (Passarelli et al., 2021), proves the importance to understand SSEs and how they influence the seismogenic character of a convergent margin. Frictional properties and stress heterogeneities along the plate interface might favor SSEs (Barnes et al., 2020; Bell et al., 2010; Im et al., 2020; Rabinowitz et al., 2018). Subducting oceanic crust and sediments release large volumes of fluids (i.e., seawater and CO_2) whose pressure can exceed hydrostatic conditions when confined within low permeability rocks, lowering the effective stress on the shallow megathrust or splay faults and creating conditions conducive to SSEs (Kitajima & Saffer, 2012; Tsuji et al., 2008; Warren-Smith et al., 2019).

The northern Hikurangi margin (HM) of North Island, New Zealand, is a subduction zone with a shallow forearc and plate interface, where sediment accretion, compaction, and deformation have been modulated for millions of years by underthrusting seamounts (Gase et al., 2021; Sun et al., 2020). Subducting topography (e.g., seamounts) may cause stress heterogeneities (Bangs et al., 2023; Leah et al., 2022; Sun et al., 2020) and fluid pressure transients (Shaddox & Schwartz, 2019) that can lead to SSEs, several of which have been characterized in great detail by onshore geodetic and offshore absolute pressure gauge (APG) data (Yohler et al., 2019). Offshore Gisborne SSEs occur every 1-2 years and can last several weeks, during which 5 to 30 cm of slip may be accommodated (Wallace, 2020). Temporal variations in the character of earthquake focal mechanisms within the subducting oceanic crust provide compelling evidence for low effective stress before an SSE (Warren-Smith et al., 2019). This observation suggests that increases in fluid pressure enable SSEs and that the slip itself is accompanied by fluid release. Nevertheless, fluid transport through the accretionary wedge in this deformation cycle is not yet well understood (Antriasian et al., 2018).

The physical properties of accreted sediments of the northern HM and their relationship to slip phenomena have been studied recently with the use of cores and data from IODP expeditions (e.g., Wallace et al., 2019). The resulting studies have shed new light on the frictional properties, shallow dewatering, and faulting near the seafloor (Aretusini et al., 2021; Boulton et al., 2019, 2022; Dutilleul et al., 2021; Fagereng et al., 2019; French & Morgan, 2020; Shreedharan et al., 2022). However, to understand how fluid flow and deformation interplay in the deeper prism, we also must consider the physical properties of older, compacted, and diagenetically mature strata (Bland et al., 2015, Bassett et al., 2022). Here we present and discuss laboratory testing performed on rock samples from the subaerial northern HM as proxies of deep rocks in the prism to better understand fluid transport within the subduction zone.

2 Materials and Method

To test the compaction, elastic, and transport properties of rocks from the northern HM, we collected and performed experiments on outcrop samples from the Raukumara peninsula (Figs 1, S1) presenting different ages and degrees of diagenesis. In the central part of the peninsula, we collected a fine-grained sandstone from the Jurassic-to-Early Cretaceous Torlesse Supergroup forming the backstop for the accretionary wedge (sample MO02) (Adams & Graham, 1996; Mortimer et al., 2014). Just east of sample MO02 location, we sampled a calcareous fine-grained sandstone with a silty matrix from the Late Cretaceous-to-Paleocene Tinui Group (sample MT07) that likely represents an early passive margin deposit, now deeply buried in the accretionary wedge (Mortimer et al., 2014). Closer to the East coast, we collected a siltstone (sample GB13) from the middle Miocene Tolaga Group, which was deposited in slope basins after subduction initiated along the HM (van de Lagemaat et al., 2022), and a glauconitic fine-grained sandstone (sample FB12) from the Pliocene Mangaheia Group.

We determined mineral abundances and assemblages of each sample through X-ray diffraction (XRD) analyses and transmitted light microscopy by preparing 30 μm in thickness thin-sections. To estimate density, porosity, compressional and shear ultrasonic wave velocities (i.e., Vp and Vs), and helium gas permeability, we prepared cylindrical core plugs with parallel end faces for each sample. Samples were tested at the UT Austin Rock-Deformation-Laboratory for confining pressures (P_c) up to 200 MPa (~12.5 km depth for hydrostatic pore pressure and overburden density of 2.6 g/cm^3) and deviatoric vertical force (F_v) ~2.6 kN. Each core plug was mounted inside a PVC jacket and between two core holders equipped with ultrasonic transducers and fluid ports, which are used to saturate and measure the core permeability. This sample assembly is mounted inside the triaxial cell (NER Autolab 1500) between the load cell and the vertical force piston. We define the mean stress as $\sigma_M = \frac{\sigma_1+\sigma_2+\sigma_3}{3}$, where $\sigma_2 = \sigma_3 = P_c$ and σ_1 is the maximum vertical stress: $\sigma_1 = \sigma_d + P_c$, where $\sigma_d = \frac{F_V}{A}$ is the deviatoric stress, and A is the sectional area of the core plug. We also define effective stress (σ') as the difference between the mean stress and the pore pressure: $\sigma' = \sigma_M - P_p$.

We measured ultrasonic velocities using the transmission method at room temperature and a frequency of ~800 kHz (Birch, 1960). To understand the effect of saturation on *V*p and *V*s, we measured the ultrasonic velocities of sample GB13 saturated with water previously chemically equilibrated with the sample. During 30 hours, we recorded the injection of 4.7 ml of this fluid, equivalent to 136% of GB13 pore-space volume.

Sample permeabilities were calculated through the transient method measuring the pressure equilibration of the helium gas contained in two volumes connected to the sample end-faces and flowing through the sample (Sutherland & Cave, 1980). To understand the effect of porosity reduction on the permeability of young, loosely consolidated rocks, we measured FB12 permeability before and after mechanical compaction, which was assumed to be isotropic. First, we measured ultrasonic velocities and permeabilities at P_c up to 70 MPa and σ_d = 5 MPa, then, we increased P_c stepwise to 100, 150 and 200 MPa and waited for 19, 24 and 5 hours to measure creep until the observed shortening rate was less than 1 μm/hour. Finally, we measured the sample permeability for varying P_c up to 200 MPa.

To study how fractures influence the permeability of HM rocks, we split sample MT07 through a Brazilian test producing a sub-vertical fracture connecting the opposite end-faces of the core plug. Then, to study the effect of stress on permeability healing, we kept the sample dry and measured permeability as a function of σ' and we collected three micro-computed tomographies (μCT) to seek evidence of variations in fracture aperture. A detailed chronology of the operations follows: On day 1, after the Brazilian test, we collected μCT dataset S1. Between day 2 and 9 we performed the first permeability test (kT1) for σ' between 24 and 65 MPa. During kT1 (days 3 to 5) we promoted healing by keeping σ' to 65 MPa. After kT1 and for the next 39 days, the sample remained inside the pressure vessel at σ'~0 MPa. Between day 48 and day 77, we performed the second permeability test (kT2) at σ' ranging 5.6 to 64 MPa. At the end of kT2 we removed the sample from the pressure vessel and acquired μCT dataset S2. Then, the jacketed sample was placed inside a humidity-controlled chamber equipped with a water container and a thermo-hygrometer. For 72 hours, a medium to low vacuum (<0.5 bar) was maintained to promote water evaporation, causing the chamber relative humidity to remain above 97% and activating clays such as smectites with pronounced swelling properties (Villar et al., 2005). Finally, we acquired μCT dataset S3, and produced a thin section perpendicular to the sample axis. On the thin section, we examined the morphology of the fracture for evidence of clay infilling, possibly caused by plastic deformation and triggered by clay swelling.

Each μCT dataset comprises 1600, 33.3 μm resolution, 16-bits TIFF images perpendicular to the sample axis, recording the entire sample except 4.37 mm at the top and bottom. After normalization and segmentation, we calculated fracture apertures (B) for each CT dataset by producing fracture aperture distribution projections (FADP) whose mean and standard deviation provided average apertures (B_m) and associated uncertainties. We report more details on the methods in the supporting information.

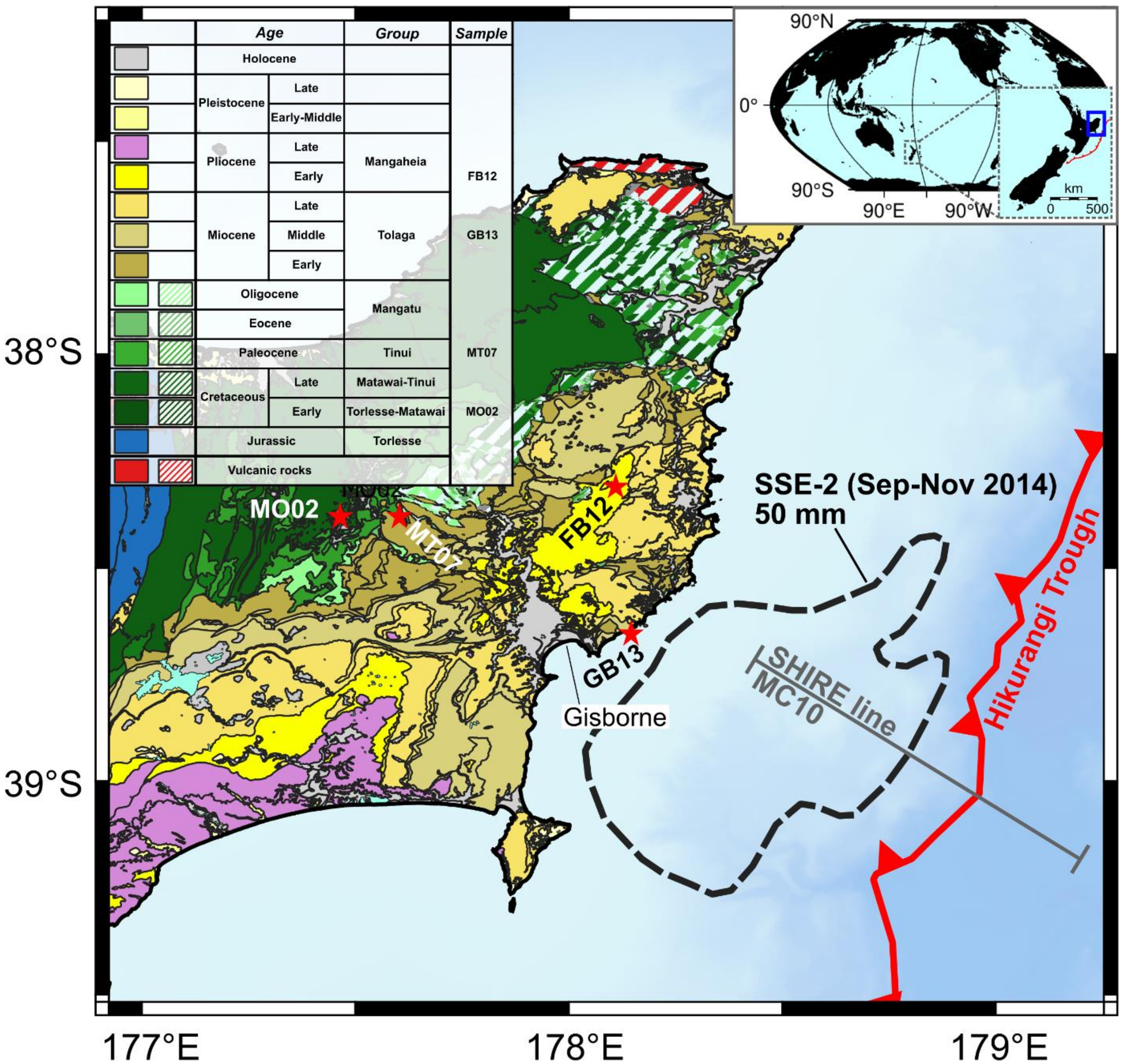

Figure 1. Geologic map of the Raukumara peninsula with the position of the rock samples used in this study (Mazengarb & Speden, 2000). The offshore dashed line contour marks the 50 mm geodetic slip model for the September-November 2014 SSE (Warren-Smith et al., 2019). The offshore line indicates the seismic line MC10 from the SHIRE project (Gase et al., 2021).

3 Results

The four samples (Fig S1) contain more than 35 wt% quartz and feldspars. The remaining minerals are calcite, and clays: chlorite, kaolinite, micas, illite, and smectite group minerals (Fig S2). Clays and swelling clays (i.e., illites and smectites) represent at least 24 wt% and 13 wt%, respectively (Fig S3). Porosities vary between 7 and 18%, where the tighter samples (MT07 and MO02) have a longer diagenetic or metamorphic history. Microphotography reveals that the grain size varies significantly among the four samples: Sample GB13 has the smallest grain size (<20 μm).

Ultrasonic velocity measurements (Fig 2A) show that Vp and Vs increase with σ_m, and the younger samples (FB12 and GB13) generally have lower wave speeds. Vp to Vs ratios vary between 1.6 and 1.95, with the least consolidated and youngest sample (FB12) exhibiting the highest values. After saturation, sample GB13 Vp increased by ~250 m/s on average while Vs decreased by ~100 m/s on average, increasing the Vp to Vs ratio from ~1.65 to ~1.95.

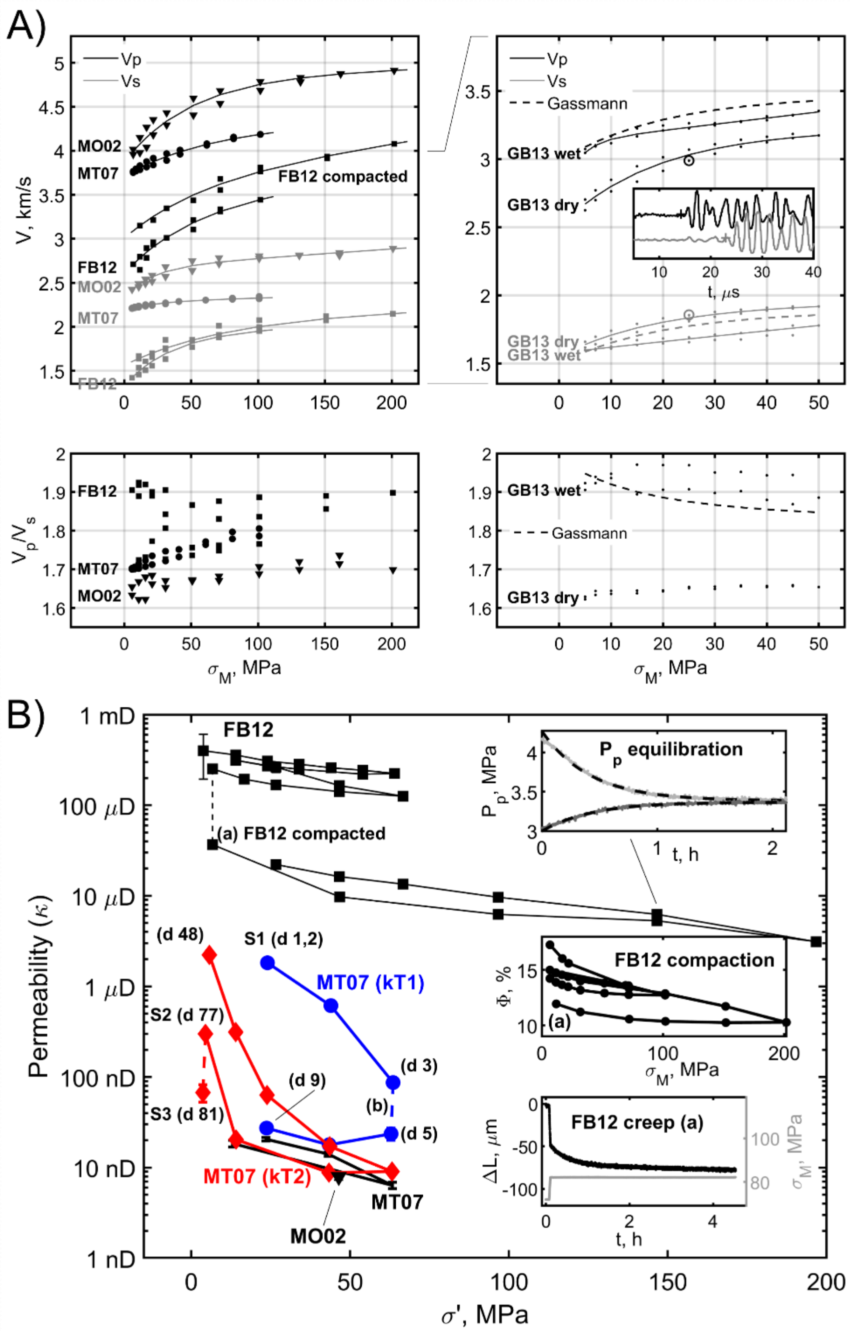

Figure 2. A) Left-top panel: ultrasonic velocities measured on the samples MO02, MT07 and FB12 as a function of σ_M. Right-top panel: ultrasonic Vp and Vs for sample GB13 when dry and saturated with water. Dashed lines indicate the theoretical saturated velocities from Gassmann fluid substitution (Gassmann, 1951). The inset shows examples of P and S waveforms recorded for the dry sample at the conditions indicated by the circled dots. Bottom panels: Vp to Vs ratios for the laboratory data and the fitting curves reported in the panels above. B) Permeabilities for samples FB12, MT07, and MO02 as a function of σ'. (a) "FB12 compaction" reports the loss of permeability due to the step-by-step increase of σ'; partial compaction and loss of porosity in sample FB12 are shown in "FB12 creep" and

“FB12 compaction” insets, respectively. "Pp equilibration" inset: example of pore pressure (Pp) equilibration and fitting curves (dashed lines). The blue and red curves show the permeability of the fractured sample MT07 during the two permability cycles (kT1 and kT2). (d X) near data points indicates X days since stage S1 (Fig 3A). Day 77 was the end of kT2 and stage S2 (Fig 3B): the sample was CT-scanned and exposed to a humid atmosphere for 72 hours. Day 81 was stage S3: we CT-scanned and remeasured the sample permeability (Fig 3C,D,E).

Before compaction, sample FB12 permeability ranged between 200 and 400 μD. Then, we raised P_c twice to 70 MPa, causing the permeability to decrease by a factor of two and porosity by 3% (i.e., at σ_m~7 MPa, porosity varied from 17.3 to 14.2%). In the following two cycles, where P_c reached 200 MPa, porosity decreased to 13.9%, and the permeability declined by almost an order of magnitude. Concurrently, the ultrasonic Vp increased from 2.6 km/s to 4 km/s.

Samples MO02 and MT07, when intact, have permeabilities below 100 nD, regardless of σ'. The permeability of the fractured MT07 evolved between stages S1, S2, and S3. After S1 and during the permeability cycle kT1, the permeability dropped from 2 μD to 87 nD. After exposing the sample to σ' ~65 MPa for more than 48 hours (Fig 4B b), we continued kT1 and found that the permeability further decreased to 24 nD. The permeability remained ~2 orders of magnitude lower than the initial permeability, i.e., around 30 nD, when σ' was reduced. After 39 days, the new increase of σ' during the second permeability cycle kT2, caused the permeability to drop to 9 nD. During the following decrease of σ', the permeability resembled pre-fracturing values. The last measurement of kT2 was performed at σ'=4.5 MPa and permeability was 300 nD, seven times lower than the initial value measured at σ'=5.6 MPa. After exposing the sample to humidity for 72 hours, the permeability, measured at σ'=3.7 MPa, decreased to 67 nD.

CT-scans visual inspection and analyses reveal the variation of B_m that varied from 190+/-110, to 72+/-43 and 84+/-45 µm during the stages S1, S2, and S3, respectively (Fig 3A,B,C). During the same stages, the number of voxels counted within the fracture varied respectively from ~934,000 to ~581,000 and ~797,000. Microphotography of sample MT07 at stage S3, shows that in several loci, the fracture collapsed, and a fine-grained amorphous mass infilled the fracture (Fig 3D,E). These observations suggest that varying confining pressure and humidification caused clay minerals plastic deformation and swelling, partially closing the fracture and reducing the permeability.

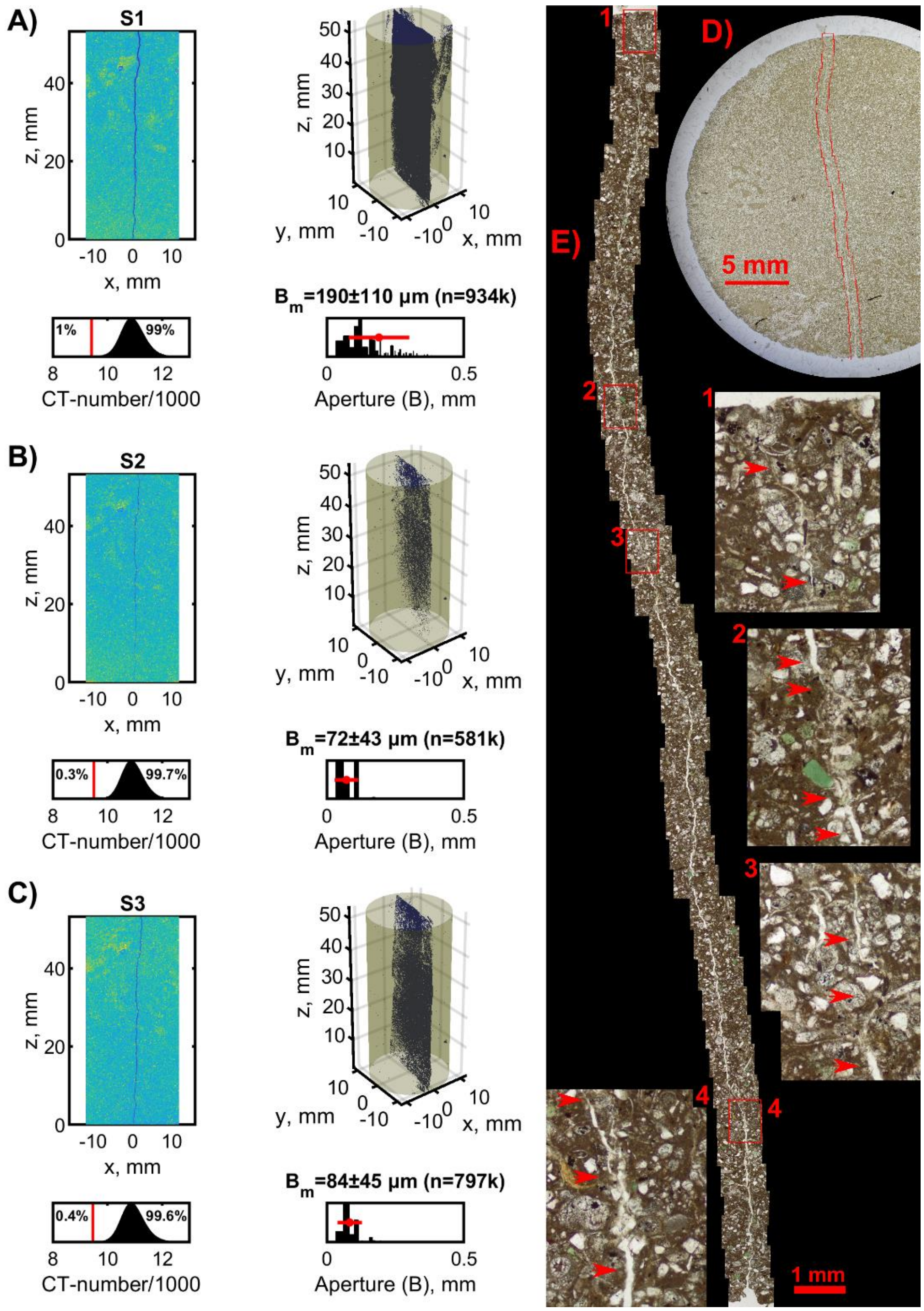

Figure 3. CT-scan and transmitted light microphotography of sample MT07 after fracturing. A, B, and C are CT-scans at stages S1, S2, and S3, respectively. Each top-left inset in these panels reports a section of the CT-scan model after normalization. CT-number distribution is shown in the bottom-left inset. The red vertical line indicates t_x (eq. S5). The percentages on the left and right of the red line indicate the relative quantity of voxels representing air and solid rock, respectively. The right top inset in each panel shows the binarized 3D model, where voxels collected within the fracture are blue. The bottom right inset show the aperture distribution (B), the calculated average and standard deviation (B_m and horizontal red bar), and the total count of voxel within the fracture (n). Panel D is a

microphotography of the thin section at stage S3. Panel E reports zooms from panel D. Insets 1 to 4 show fracture infill, which are highlighted by red arrows along with open fractures.

4 Discussion

We provide porosity-permeability relationships for rock samples from the subaerial northern HM under a range of confining pressures. Ultrasonic velocities of dry samples are similar to the seismic velocities estimated offshore New Zealand by the SHIRE project (Gase et al., 2021). The seismic reflectivity imaged along the transect MC10 shows the decollement along the prism base and several splay faults that may partly accommodate the convergence (Fig 4A). Inside the prism, Vp increases gradually from 2.0 km/s near the surface to 4.5 km/s at the prism base ~7 km below sea level. In Fig 4B, the comparison between the seismic and ultrasonic velocities suggests that sample FB12, and possibly also sample GB13, represent the modern slope basins on the outer prism, which is consistent with their depositional environment. The ultrasonic velocities of sample MT07 of the Tinui Group correspond well to the velocities of the deep part of the prism, where Vp reaches 4.5 km/s. Compaction and diagenesis must contribute to the increase of Vp with depth (Dvorkin & Nur, 1996; Saxena & Mavko, 2014). We measured an ultrasonic Vp of 4.8 km/s at 150 MPa in the Torlesse basement sample MO02, which is higher than what we imaged in the deep prism on Line MC10 (Fig 4), suggesting that there may not be a deep offshore portion of the Torlesse basement offshore northern HM (Bassett et al., 2022; Gase et al., 2021).

Our comparison between seismic and ultrasonic velocities in Figure 4 is semi-quantitative as uncertainty is introduced by microcracks produced during sample preparation – see SI for details (Eberhart-Phillips et al., 1989; Tsuji & Iturrino, 2008), and by frequency differences. Velocities in section MC10 and on our samples have been measured at frequencies around 20 Hz and 800 kHz, respectively. Considering the frequency range, a typical P-wave quality factor ranging from 30 to 150, and a nearly-constant Q model (Liu et al., 1976; Tisato et al., 2021), we should expect a velocity dispersion between 2.3 and 12%. Conversely, SHIRE and laboratory data were collected on saturated and dry samples, respectively. Saturation increases P-wave velocities of sample GB13 by 5 to 15%, suggesting that the effects of fluid saturation and anelasticity on velocities should counteract each other. Given the similarity in P-wave velocities and depositional environment, we suggest that the Tinui and Tolaga group rocks (samples MT07 and GB13) are good lithological proxies for the deep and shallow offshore Hikurangi prism, respectively.

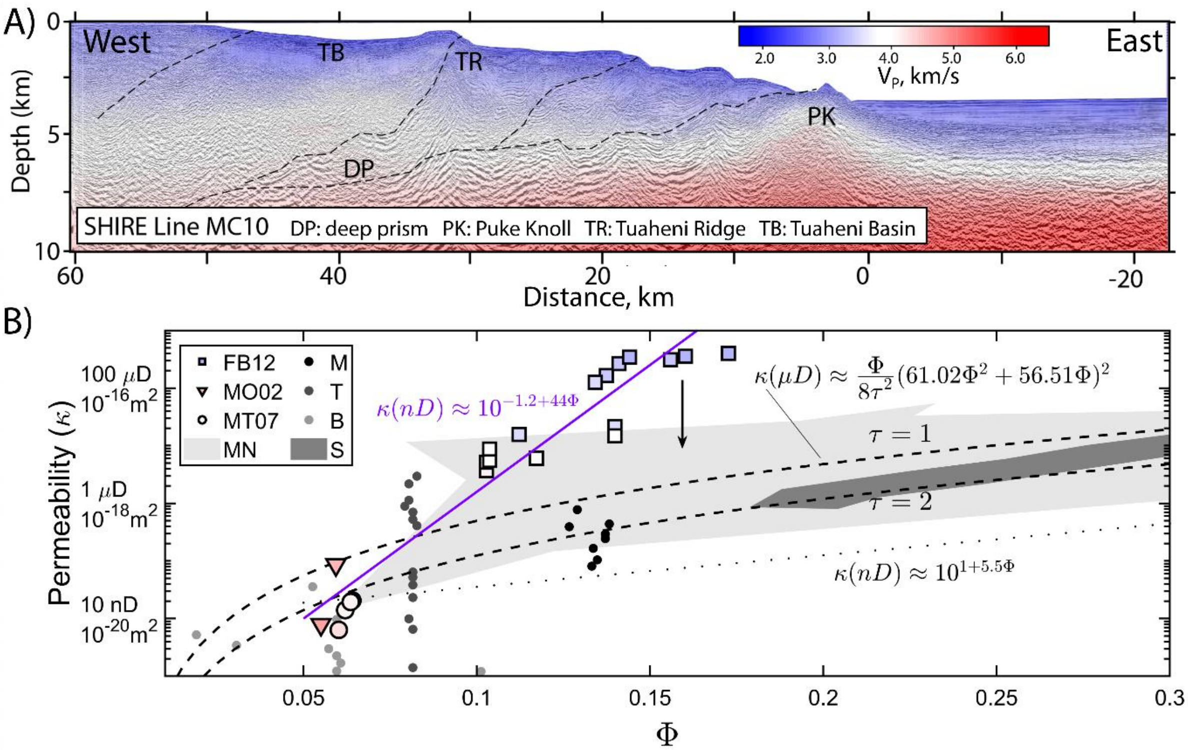

Figure 4. A) Velocity model along the SHIRE Line MC10 (Gase et al., 2021). B) Summary of laboratory result: permeabilities vs porosity and color-coded markers (colorbar in panel A) as a function of ultrasonic Vp for samples FB12, MO02, and MT07 (Tables S1 and S2). The arrow indicates in which direction the permeabilities vary when tests are performed using water rather than helium gas. Dashed lines indicate empirical permeability vs porosity according to eq. 3. The dotted line represents an average permeability for unconsolidated clays and possibly a lower bound for the permeability of HM sediments (Neuzil, 1994). S data (dark-gray area) are for siltstones (Reece et al., 2012). The continuous line fits our data and agrees with measured mudstone permeabilities indicated by the MN gray-shaded area (Magara, 1978; Neglia, 1979). Such a line also represents an upper bound for the permeability of HM rocks. M, T, and B data are permeabilities measured in boreholes: M by Reisdorf et al. (2016), Yu et al. (2017); T by Boisson et al. (2001); B by Intera Eng. Ltd. (2011), Roberts et al. (2011), Walsh (2011).

The permeability of our samples ranges from 1 nD to 1 mD, with the samples representing the deep part of the prism being the tightest. Neuzil (1994, 2019) compiled data from several studies on unconsolidated clays with a maximum porosity of 80%, and a few consolidated mudstone-siltstones with porosities (Φ) <35%. Saffer & Bekins (1998) followed Neuzil's work and described the permeability (κ) of the Nankai accretionary complex as:

$$\boldsymbol{\kappa(nD) \approx 10^{1+5.5\phi}} \qquad \text{eq. 1}$$

Equation 1 fits the porosity-permeability relationship of unconsolidated sediments and is a lower bound for the permeability of mudstones that are similar to our samples (Magara, 1978; Neglia, 1979; Reece et al., 2012). On the other hand, we found that:

$$\boldsymbol{\kappa(nD) \approx 10^{-1.2+44\phi}} \qquad \text{eq. 2}$$

fits our results and is an upper bound for the permeability of mudstones. We suggest that the permeabilities calculated from equations 1 and 2 (Fig 4B) overestimate permeabilities in the Northern Hikurangi accretionary prism at depths >1 km because helium gas is not as efficient as seawater in activating swelling clays whose expansion lowers the effective permeabilities (Villar et al., 2005); At burial depths >1-2 km, the porosity of clay-bearing sediments, mudstones, siltstone, and shales drops below 35% (Griffiths & Joshi, 1989; Magara, 1978; Skempton, 1969); Permeabilities measured in boreholes are typically orders of magnitude higher than those measured in the laboratory due to the presence of fractures (Fig 4B lines M,T,B) (Neuzil, 2019), and numerical models of permeability in microfractured claystones agree with the mudstone porosity-permeability in Fig 4B (Vora & Dugan, 2019). We also propose that the permeability of rocks in the Northern Hikurangi accretionary prism can be described by a Kozeny-Carman relation (dashed lines in Fig 4B):

$$\boldsymbol{\kappa = \frac{\phi}{8\tau^2} R^2} \qquad \text{eq. 3}$$

Where $\boldsymbol{\tau}$ is tortuosity, and *R* is the median pore diameter (Carman, 1997). We obtained $\boldsymbol{R(nm) = 61.02\phi^2 + 56.51\phi}$ from data reported by Hunt (1996) for similar lithologies.

Every 1-2 years, the northern HM experiences an SSE that lasts several weeks (Wallace, 2020). Recent analyses of the APG data offshore Gisborne have shown that the 2014 SSE may have experienced up to 30 cm of slip in the center of a ~100 km wide patch, though less displacement is expected along the edges (Yohler et al., 2019). Some authors have suggested that SSEs that originate along the decollement at the base of the wedge are accompanied by slip diverted to thrust faults in the Hikurangi accretionary wedge (Shaddox & Schwartz, 2019). We expect SSEs to deform and fracture the rocks along these thrust faults (Morgan et al., 2022). Our laboratory measurements before and after rock failure for sample MT07 show that the deeper prism, where Tinui Group equivalent rocks may be present, may experience large increases in permeability during an SSE.

In a few weeks, the fractured sample MT07 regained its pre-fracturing permeability. Between stages S1 and S2, the permeability recovery was achieved in dry conditions. Although sample MT07 and sample FB12 have different compaction levels and grain sizes, they share similar mineralogy. Thus, although limited, we expect plastic deformation also in sample MT07, likely concentrated near clays (Mondol et al., 2008). Between stages S2 and S3, the permeability decreased by a factor of 5 while B_m increased, suggesting clay expansion. Once confined, we expect that the hydrated clays would deform plastically, clogging the fracture more efficiently than dry clays and justifying the permeability loss. We propose that permeability healing is also present along HM faults, given the presence of clays at depth, especially above the 5-7 km deep temperature-controlled smectite-illite transition **(**Antriasian et al., 2018; Freed & Peacor, 1989; Pecher et al., 2017; Tisato & Marelli, 2013).

In the Hikurangi subduction zone, fluids expelled from pore space and fluids released by dehydration reactions travel along the plate interface or through the accretionary wedge (Ellis et al., 2015). As the fluid pressure increases near the decollement and inside the accretionary wedge, conditions may become favorable for an SSE (Burgreen-Chan et al., 2016; Kobayashi & Sato, 2021). Though this mechanism has been proposed for several subduction zones where SSE occur at larger depths (Audet et al., 2009; Kodaira et al., 2004), the analysis of Warren-Smith et al. (2019) on the northern HM, is also compatible with the sealing of fluid pathways after an SSE. The expansion and plastic deformation of clays may provide an efficient mechanism to reduce permeability over weeks or months after an SSE.

Permeability healing, favoring the development of overpressures, reconciles with the poor mechanical healing shown by Shreedharan et al. (2023), hindering elastic energy accumulation, because both set conditions conducive to SSEs.

5 Conclusions

We provided relationships between porosity, permeability, and confining pressure for rocks that make up the accretionary prism of the northern HM. We suggest an empirical porosity-permeability relationship to model fluid transport and estimate effective stress in shallow subduction zones. Mechanical failure of these rocks enhances permeability, but over the course of several weeks, healing reduces the permeability again, suggesting that after an SSE, sediments deep in the northern HM accretionary prism can recover permeability efficiently within the time frame of an SSE as a mechanism explaining the regular recurrence of these events.

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Open Research

Data are publicly available upon publication at https://doi.org/10.18738/T8/RMXMIQ or can be requested to the corresponding author.

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Permeability and elastic properties of rocks from the northern Hikurangi margin: Implications for slow-slip events

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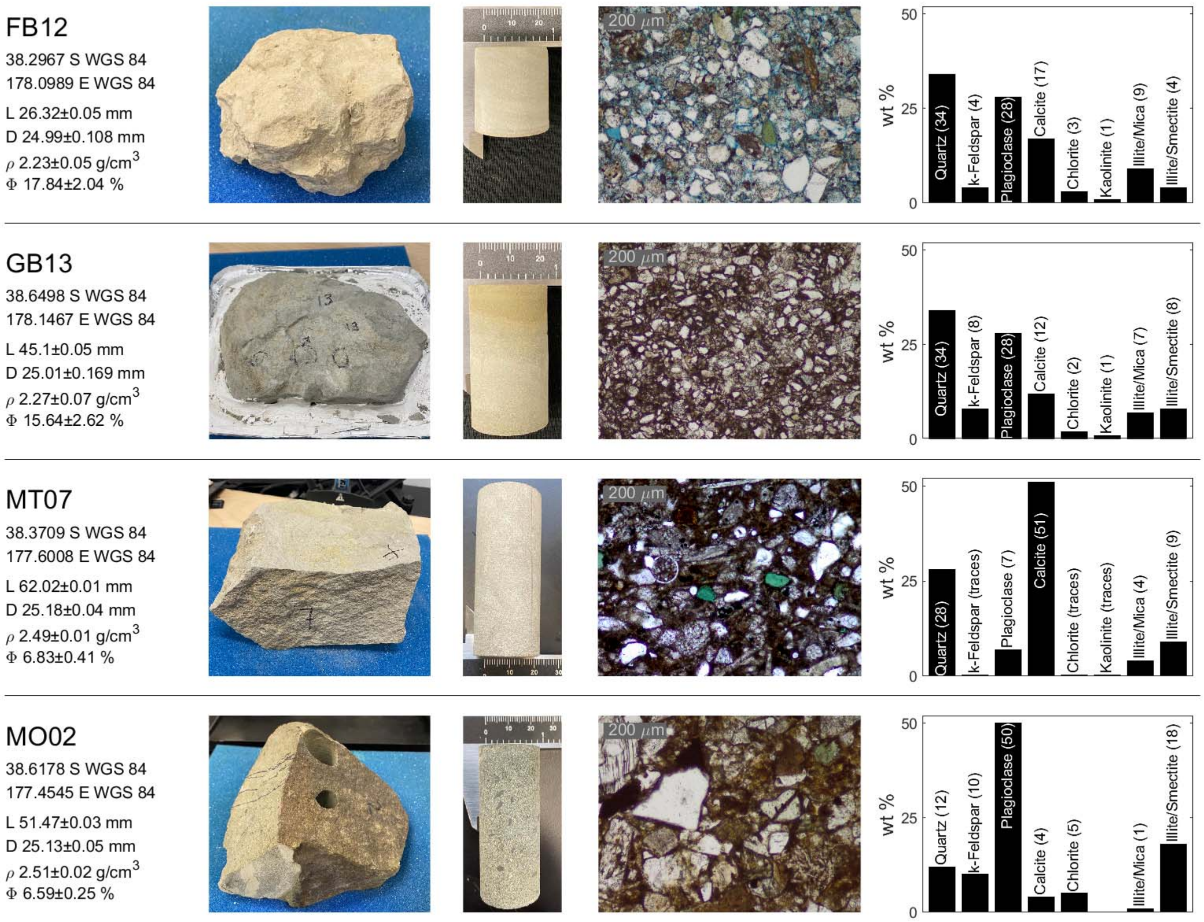

Figure S1. For each sample, the left column reports the geographic coordinates, length (L), diameter (D), density (ρ), and porosity (Φ). The three center columns are pictures of hand samples and transmitted light microphotographs. The right column reports mineral compositions according to X-ray diffraction analyses (XRD).

Sample preparation details

The end faces of each core plug were smoothed to parallel using a rock saw and a lathe equipped with an angular grinder. Parallelism was checked with a 0.01 mm resolution caliper. Each core was oven-dried at ~333 K for several days to reduce absorbed water. We then calculated the total volume and density of each core by measuring its mass and dimensions using a scale and a caliper to accuracies of 0.001 g and 0.02 mm, respectively. A helium pycnometer (Micromeritics AccuPyc II 1340) was used to measure the solid volume and porosity of each core.

To evenly distribute the saturating water or the helium gas to test permeability across the sample end-face, we placed 3.175 mm thickness, 10 μm grain size, AISI 316 stainless steel porous frits between each sample holder and the adjacent sample end-face.

Sample MT07 at stage S3 - i.e., fractured after being exposed to humidity - was epoxy impregnated before removing the rubber jacket to avoid offsetting the fracture.

Preparation of the saturating water for sample GB13

Water chemically equilibrated with sample GB13 was prepared and injected as follows: For several weeks before saturation, we submerged a few grams of GB13 granules in deionized water. Then, the injection of such aqueous fluid was performed using a high-pressure syringe pump (ISCO 260HP), recording – via a Matlab script - the injected volume and injection pressure. The latter was maintained constant to a value of 3 MPa lower than the confining pressure that varied between 20 and 50 MPa.

Ultrasonic and mechanical testing details

Our samples have a maximum ultrasonic velocity of ~6 km/s and considering the testing frequency of 800 kHz, we estimated a maximum wavelength (λ_M) of 7.5 mm and, to avoid nearfield effects, we prepared cores with a length (L) > 3 λ_M. Velocities were estimated with the transmission method by measuring the time of travel of the elastic wave along the core plug (Birch, 1960). We corrected the first arrival by the delay introduced by the sample holders that was determined by a standard calibration procedure (e.g., Prelicz, 2005). A pulser-receiver apparatus (JSR Ultrasonics DPR300) generated a negative spike pulse with a typical duration of ~40 ns feeding the source ultrasonic transducer. We used a pulsing rate of 100 pulses/sec (PRF RATE=1), pulse amplitude of ~194 V (PULSE AMPLITUDE = 4, and PULSE ENERGY = HIGH Z 4), and damping of 331 Ohms (DAMPING = 1). In addition, the pulser-receiver produces a trigger signal (5 V in amplitude) to synchronize the pulser and the oscilloscope (Rigol DS1104Z-S) collecting the signals generated by the receiving transducer and amplified by the receiver. The latter has a gain of 66 dB (REL. GAIN = 79), a high-pass filter corner frequency of 1 MHz, and a low-pass filter corner frequency of 3 MHz. Two data transfer switches allow selecting the recording of the V_P, V_{S1} or V_{S2} signal. To improve the signal-to-noise ratio the oscilloscope collects and stacks 1024 signals and transmits the digitized wavelets to a computer via a USB port. Typically, the signal, comprising 1200 samples, is digitized every 0.2 µs or less and saved as a comma-separated-value (CSV) file. Shear velocities were calculated as the average of V_{S1} and V_{S2}.

All velocities (V) as a function of σ_M were fit according to Eberhart-Phillips et al., 1989:

$$V = a + k\,\sigma_M - b\,e^{-d\,\sigma_M} \qquad \text{eq. S1}$$

Where a, k, b, and d are fitting parameters. Table S1 reports the fitting parameters for all the measurements reported in Figure 2A. As σ_M increases, especially above ~50 MPa, the effect of the non-linear part of eq. S1 decreases, and V tends to be equal to:

$$V = a + k\,\sigma_M \qquad \text{eq. S2}$$

The exponential increase of velocity (e.g., $-b\,e^{-d\,\sigma_M}$) is controlled by crack closure (e.g., Eberhart-Phillips et al., 1989; Tsuji & Iturrino, 2008). Cracks are naturally occurring, but some of our sample cracks were probably produced during preparation. Therefore, the measured velocities and those modeled with eq. S1 possibly underestimate the velocities of the undisturbed rocks. On the other end, the velocities calculated according to eq. S2 represent an upper bound for the undisturbed rock velocities. Therefore, to provide a range of possible velocities, table S2 reports values calculated according to eqs. S1 and S2, and we used their average to color code the symbols in Figure 4B, which compares ultrasonic and seismic velocities in section MC10 (fig. 4A).

We estimated the ultrasonic wave velocities of the saturated sample GB13 (wet) using the Gassmann fluid substitution (Gassmann, 1951). We obtained the dry bulk and shear modulus from the measured ultrasonic velocities and density. We used a porosity of 15.64% and estimated the effective bulk modulus of the mineral material making up the rock (K_0=41.9 GPa) using the Voigt-Reuss-Hill average (Hill, 1952). Such an average was calculated considering the mineral abundances and bulk moduli in Table S3.

Samples compaction was measured to 1 µm accuracy with a Linear Variable Displacement Transducer connected to the axial piston, whose signal was acquired along with the confining pressure and vertical force.

Sample	Vp a, km/s	Vp k, km/(s MPa)	Vp b, km/s	Vp d, 1/MPa	Vs a, km/s	Vs k, km/(s MPa)	Vs b, km/s	Vs d, 1/MPa
MT07	4.259	0.00040	0.5508	0.01559	2.287	0.00040	0.0979	0.0403
MO02	4.833	0.00048	0.9625	0.01998	2.671	0.00107	0.3333	0.0469
FB12	3.411	0.00149	0.8080	0.01892	1.935	0.00052	0.6075	0.0280
FB12 compacted	3.655	0.00221	0.6489	0.01697	2.087	0.00048	0.5253	0.0135
GB13 dry	3.198	0.00040	0.6979	0.05500	1.925	0.00040	0.3755	0.0530
GB13 wet	3.120	0.00453	0.4441	0.30994	1.598	0.00370	0.0215	0.0301

Table S1: Fitting parameters for the samples ultrasonic velocities according to eqs. S1 and S2.

Sample	Φ, %	κ, m2	σM, MPa	Vp (meas.) km/s	Vp (EP89) min, km/s	Vp (EP89) max, km/s	Vp (EP89) mean, km/s
FB12	17.3	3.95E-16	10	2.788	2.757	3.426	3.092
FB12	16.0	3.52E-16	20	2.937	2.887	3.441	3.164
FB12	15.6	3.04E-16	30	2.986	2.998	3.456	3.227
FB12	13.7	1.63E-16	50	3.055	3.172	3.486	3.329
FB12	13.4	1.24E-16	70	3.251	3.301	3.515	3.408
FB12	14.1	2.60E-16	30	2.968	2.998	3.456	3.227
FB12	14.4	3.38E-16	20	2.895	2.887	3.441	3.164
FB12 compacted	14.0	2.13E-17	30	3.333	3.331	3.721	3.526
FB12 compacted	14.0	1.46E-17	70	3.543	3.611	3.809	3.710
FB12 compacted	11.7	5.97E-18	150	3.866	3.935	3.986	3.961
FB12 compacted	10.3	3.64E-18	200	4.008	4.075	4.097	4.086
FB12 compacted	10.3	5.10E-18	150	3.863	3.935	3.986	3.961
FB12 compacted	10.4	5.56E-18	100	3.745	3.757	3.876	3.816
FB12 compacted	10.4	8.67E-18	70	3.618	3.611	3.809	3.710
FB12 compacted	11.2	1.54E-17	30	3.298	3.331	3.721	3.526
MO02	5.9	8.47E-20	30	4.353	4.320	4.848	4.584
MO02	5.5	7.80E-21	50	4.479	4.502	4.857	4.680
MT07	6.4	2.03E-20	30	3.838	3.926	4.271	4.098
MT07	6.2	1.39E-20	50	3.913	4.026	4.279	4.153
MT07	6.0	6.29E-21	70	3.995	4.102	4.287	4.194
MT07	6.4	1.92E-20	20	3.843	3.864	4.267	4.065

Table S2: Porosity, permeability, mean stress, and Vp for our sample data that are reported in Figure 4B. 'Vp (meas.)' indicate the measurements, 'Vp (EP89) min' is the velocity estimated using eq. S1, 'Vp (EP89) max' is the velocity estimated according to eq. S2. 'Vp (EP89) mean' is the average between 'Vp (EP89) min' and 'Vp (EP89) max'. The latter is used to color-code the symbols of samples MT07, MO02, and FB12 in Figure 4B.

Mineral	Fraction	Bulk Modulus
Quartz	34%	37.0 GPa
K-feldspar	8%	37.5 GPa
Plagioclase	28%	76.0 GPa
Calcite	12%	77.0 GPa
Clays	18%	15.0 GPa

Table S3. Parameters used to calculate the effective bulk modulus of the minerals making up sample GB13 (K_0). Fractions are estimated from XRD (see Figure S1), and bulk moduli are taken from (Carmichael, 1989).

Permeability testing

The two reservoirs connected to the sample end-faces have volumes V_1=58.725 ml and V_2=162.53 ml, and at the beginning of the test, we connected the reservoirs to a high-pressure helium gas bottle to raise their internal pressures to two different values $P_{1i} > P_{2i}$. While P_{1i} is greater than P_{2i}, helium flows through the sample until pressure equilibrium is reached. Two digital manometers (Keller LEO3) connected to a computer and a Matlab code record P1 and P2 over time (t). The two manometers also measure temperature (T). Permeability is then calculated as:

$$\kappa = -\frac{\beta\, \eta\, L}{\left(\frac{1}{V_1}+\frac{1}{V_2}\right) K\, A}, \qquad \text{eq. S3}$$

Where η and K are Helium viscosity and bulk modulus, respectively; L and A are the lengths and cross-section area of the sample; β is the exponent of the pressure decay:

$$P_1 = (P_{1i} - P_{2i})\, e^{\beta\, t} + P_f, \quad \text{eq. S4}$$

Where P_f is the equilibrium pressure, i.e., P_1 and P_2 at time infinite. We assume helium properties as a function of pressure and temperature from the national institute for standards and technology (NIST) fluid thermophysical properties (Arp et al., 1998; Ortiz-Vega et al., 2020). P_f and β were estimated by means of a non-linear least absolute residuals fit implemented in Matlab.

XRD and CT-scanner setup

Mineralogical X-ray diffraction analyses were conducted at the Geomaterials Characterization and Imaging Facility (GeoMatCl) at The University of Texas at Austin. Whole rock samples were manually homogenized, ground, and sieved to a 250 µm mesh size. XRD analyses were performed using a Bruker D8 diffractometer instrument equipped with Cu Kα radiation and a nickel filter, along with a LYNXEYE solid-state detector. The analyses were carried out at a voltage of 45 kV and a current of 40 mA, employing a 2θscan axis ranging from 3° to 70°, with step increments of .0195° (2θ) and a duration of 1 s per step. Whole rock X-ray patterns (Fig S2) were determined through Rietveld refinement utilizing Bruker TOPAS 4.2 software.

For clay speciation analyses (Fig S3), we followed the modified methods based on Hillier (2000) and Moore & Reynolds (1997). $CaCO_3$ rich samples were subjected to a modified HCl-Na_2CO_2 treatment (5% diluted HCl) to disseminate clay minerals following the method of Komadel et al. (1990) and Meredith E. Ostrom (1961). Disaggregated material was separated into a <2-micron clay fraction suspension using sodium hexametaphosphate, enabling the acquisition of clay speciation by excluding heavier non-phyllosilicate minerals. The <2-micron clay suspension was vacuum-filtered through a millipore filter and subsequently oriented onto a glass slide. The oriented clay mounts were subjected to ethylene glycol vapors for 24 hours, followed by heating (1 hour) to 400°C to identify swelling clays. Clay speciation X-ray patterns with a 2θ scan axis ranging from 3° to 70°, with step increments of 0.195° (2θ) and a duration of 1 s per step were evaluated using reference intensity ratios (RIR), and mineral intensity factors (MIF) with the MDI Jade software.

For CT-scanning we used an NSI scanner equipped with a Fein Focus High Power source, at 120 kV voltage and 0.14 mA current. CT scans were acquired at 33.3 μm per voxel resolution. The X-ray source was filtered using aluminum foil. The CT scanner is equipped with a Perkin Elmer detector, with 0.5 pF gain, and the 1800 projections were collected at 1 fps and 1x1 binning. The source-to-object distance was 150.566 mm, and the source to detector 963.799 mm. We performed a continuous CT scan by averaging 2 frames and by skipping 0 frames. We applied a beam-hardening correction of 0.25 and a post-reconstruction ring correction using the following parameters: oversample = 2, radial bin width = 21, sectors = 32, minimum arc length = 2, angular bin width = 9, angular screening factor = 4. The final reconstructed volume had a voxel size of 33.3 μm and 1873 slices.

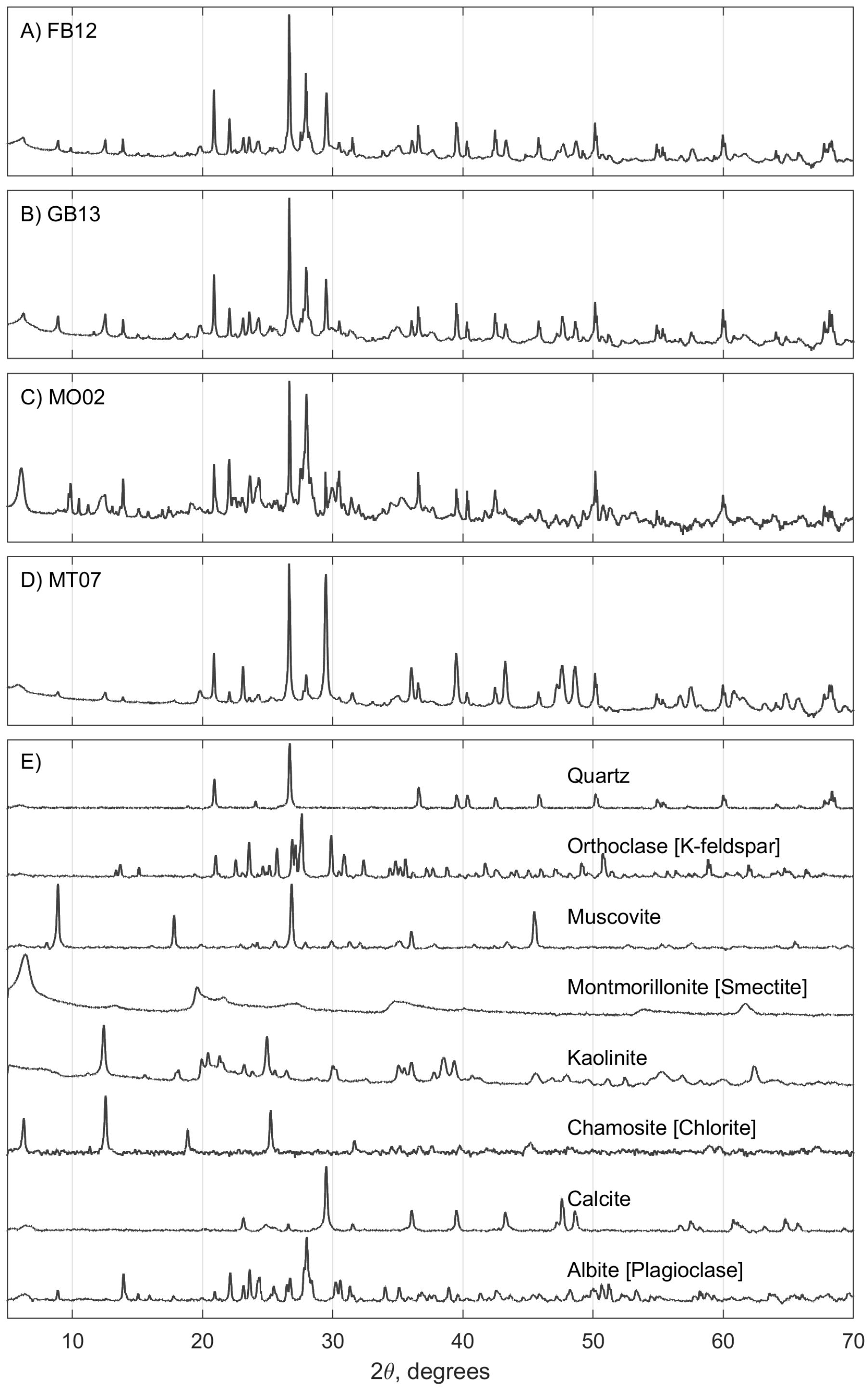

Figure S2. A-D) XRD spectra of the four samples. E) Standard spectra for the mineral comprising our samples. Data have been taken from the RRUFF database (Lafuente et al., 2015): Talc URL=rruff.info/R040137; Quartz URL=rruff.info/R040031; Orthoclase URL=rruff.info/R040055; Muscovite URL=rruff.info/R040104; Montmorillonite URL=rruff.info/R110052; Kaolinite URL=rruff.info/R140004; Chamosite URL=rruff.info/R060188; Calcite URL=rruff.info/R040070; Albite URL=rruff.info/R040068.

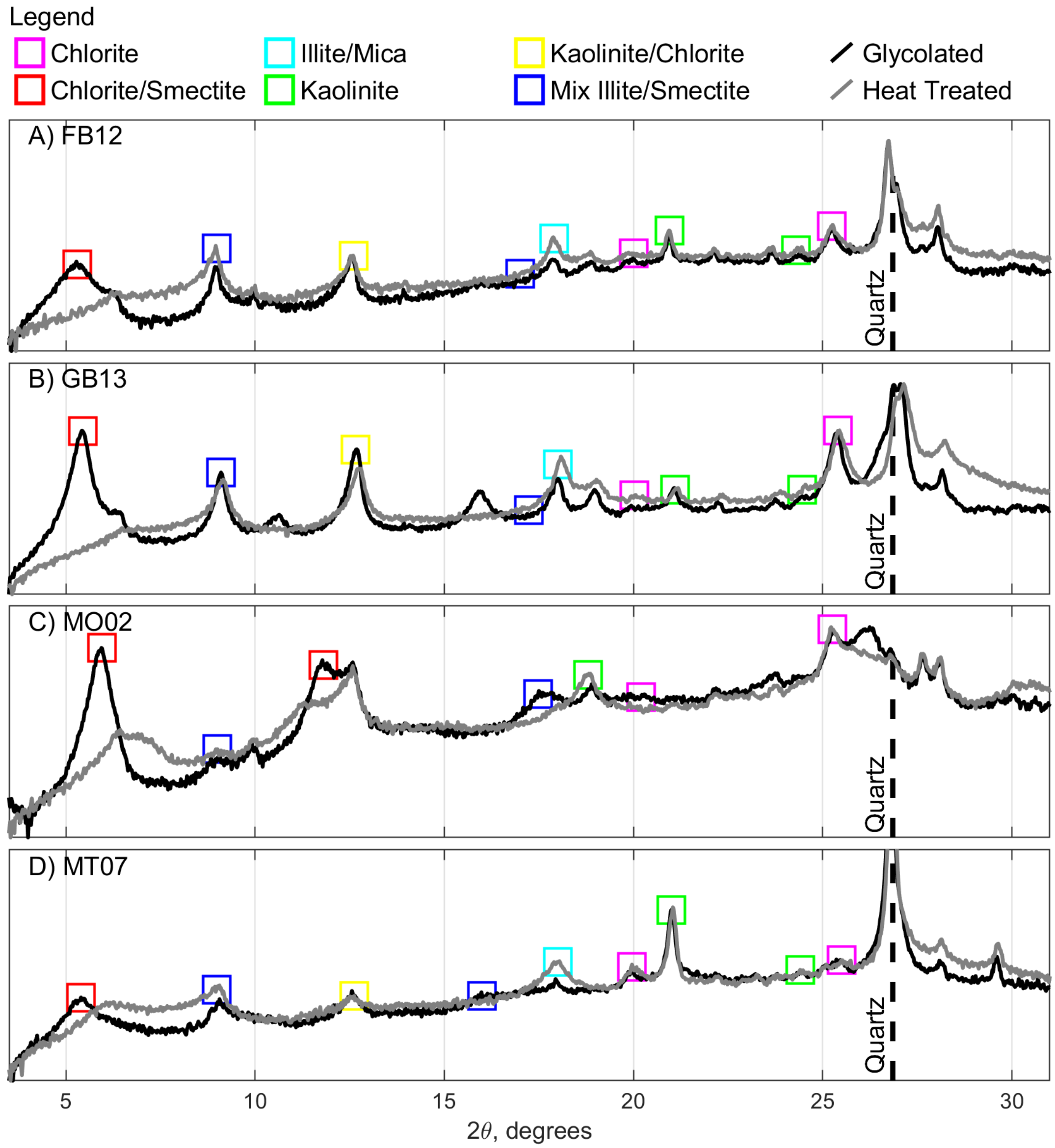

Figure S3. XRD clay patterns (oriented, glycolated, heat-treated at 400°C) for Illite/Mica, Mix Illite/Smectite, Kaolinite, and Chlorite minerals. Squares indicate peaks and portions of spectra used to speciate and estimate clay fractions for each sample.
A) Sample FB12 is dominated by Illite/Mica, followed by Mix Illite/Smectite, with minor quantities of Chlorite and Kaolinite. B) Sample GB13 exhibits an abundance of Mix Illite/Smectite and Illite/Mica, along with trace amounts of Chlorite and Kaolinite. C) Sample MO02 is notably rich in Mix Illite/Smectite, with a significant presence of Chlorite and minor content of Illite/Mica.
D) Sample MT07 is primarily rich in Mix Illite/Smectite, featuring a notable abundance of Illite/Mica, and minor quantities of Chlorite and Kaolinite.

Fracture aperture calculation

To normalize CT-scan datasets, we fit a Gaussian function to the distribution of CT numbers to obtain a CT-number mean (m_x) and standard deviation (s_x), where x is either S1, S2, or S3. To compare datasets acquired at different stages, we shifted the CT-numbers of datasets S2 and S3 by m_{S1}-m_{S2} and m_{S1}-m_{S3}, respectively. We added a value of 1 to each voxel, cropped each image to 718x718 pixels around the sample center, and assigned a value of 0 to pixels with a distance >718/2 from the sample center. We binarized the datasets to assign each voxel to either solid rock or air by applying a threshold calculated as:

$$t_x = m_x - 2.5\ s_x \qquad \text{eq. S5}$$

Voxels with CT-number equal to or greater than t_x were assumed to represent rock and assigned a value of 255. Voxels with CT-number lower than t_x and greater than zero were assumed to be air and assigned a value of 128.

To obtain a FADP of a binarized dataset, we calculated: 1) The Euclidian distance of each voxel in the fracture. This is achieved by a) performing an iterative image morphological erosion assigning approximated distances of each fracture voxel from the fracture rim, and b) calculating the Euclidian distance of each voxel within the fracture from the closest voxel representing rock; 2) The skeleton of the fracture (SK) comprises the voxels that are within the fracture and have the maximum Euclidian distance from the fracture rim into respect the 26 surrounding voxels. Such a device extracts the center surface while preserving the topology and Euler number, also known as the Euler characteristic of the objects (Kerschnitzki et al., 2013; Lee et al., 1994). Finally, the FADP was calculated at each SK location by doubling the Euclidian distance recorded in such voxels.

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