Subduction megathrust creep governed by pressure solution and frictional-viscous flow

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Subduction megathrust slip speeds range from slow creep at plate convergence rates (centimetres per year) to seismic slip rates (metres per second) in the largest earthquakes on Earth. The deformation mechanisms controlling whether fast slip or slow creep occurs, however, remain unclear. Here, we present evidence that pressure solution creep - fluid-assisted, stress-driven mass transfer - is an important deformation mechanism in megathrust faults. We quantify megathrust strength using a laboratory-constrained microphysical model for fault friction, involving viscous pressure solution and frictional sliding. We find that at plate-boundary deformation rates, aseismic, frictional-viscous flow is the preferred

deformation mechanism at temperatures above 100 ˚C. The model thus predicts aseismic creep at temperatures much cooler than the onset of crystal plasticity, unless a boundary condition changes. Within this model framework, earthquakes may nucleate when a local increase in strain rate triggers velocity-weakening slip, and we speculate that slip area and event magnitude increase with increasing spacing of strong, topographically derived irregularities in the subduction interface.

Understanding why some megathrust segments accommodate displacement by earthquake slip versus aseismic creep is a major challenge. Geophysically observed variation in seismic style along active subduction megathrusts, involving a continuum of slip speeds from plate boundary creep rates to earthquake slip[1], arises from processes within a fault zone in subducting sediments on top of potentially rugged ocean floor[2-6]. Dominantly creeping margins are characterised by low seismic coupling coefficients - the observed seismic moment release rate over that required by plate motion vectors - and lack of earthquake moment magnitudes ≥ 8.0 (Supplementary Figure S1)[7,8]. Thus, some margins produce small to medium magnitude earthquakes, but the total moment of these earthquakes is insufficient to explain total geodetically observed displacement, and they must therefore be accompanied by aseismic creep[9].

The megathrust interface is commonly inferred as seismogenic to a depth where temperature exceeds the 350˚C required for crystal plasticity in quartz, or to the intersection with the hanging wall Moho, whichever is shallower[10]. However, geodetic inversions[8,11-14] reveal aseismic creep shallower than both the 350˚C isotherm and the hanging wall Moho. The question thus arises: how do some megathrust segments, such as north Hikurangi[11], the southern Japan Trench[12], southern New Hebrides[13], southern Kermadec Arc[13], and the Manila Trench[14] accommodate detectable displacement by aseismic creep in addition to moderate size earthquakes, both originating at a similar depth range? This observation requires average creep rates of centimetres per year at temperatures less than 350°C. Identifying the associated creep mechanism is critical for recognising where megathrust displacement can occur without great earthquakes, and by contrast, interpret where great earthquakes may occur.

The mechanism of creep at seismogenic zone conditions

Tectonic mélanges comprising sheared trench-fill and ocean floor sediments have been interpreted as megathrust fault rocks (Fig. 1a)[3-6,15]. Deformation structures developed at seismogenic pressure-temperature ($P - T$) conditions include both discontinuities, such as faults and tensile fractures, and continuous structures such as folds, boudins and foliations. One possible interpretation is that faults and associated fractures represent seismic deformation styles, whereas continuous features characterise slower, distributed, aseismic mechanisms recorded as creep[5,6,15]. In this interpretation, the mechanism accommodating deformation in continuous structures is responsible for aseismic creep.

In exhumed subduction thrusts, cleavage defined by fine-grained phyllosilicates wraps around rigid quartz clasts (Fig. 1b). Comparable microstructures are reported in borehole samples from the creeping segment of the continental San Andreas transform fault[16,17]. Mass balance calculations on San Andreas samples indicate pressure solution, involving fluid-assisted, stress-driven mass transfer, as the cleavage-forming process[16].

If empirical rates can be extrapolated, pressure solution is fast enough to account for aseismic sliding[16,18].

Pressure solution is also widely inferred as the dominant cleavage-forming process in mudrocks and phyllites sampled from exhumed subduction thrusts[6,19-21]. As an example, we consider a sample representative of sheared, cleaved mudstone from an inferred exhumed megathrust in the Chrystalls Beach Complex, New Zealand[21](Fig. 1a-e), where cleavage defined by illite-muscovite developed at $T < 300^{\circ}$C (ref. *22*). In this sample, cleavage seams are depleted in Si and enriched in Al (Fig. 1c; Supplementary Figure 2). If cleavage develops by pressure solution, more soluble elements, such as Si, are dissolved, whereas less soluble elements, such as Al, are retained. Thus, the observations in the Chrystalls Beach sample are consistent with cleavage formation by pressure solution. Stress shadows around quartz clasts lack evidence for opening of pore space (Fig. 1d), and are sites of local silica enrichment (Fig. 1e). In addition to formation of phyllosilicate cleavage, mass-transfer processes are therefore illustrated by silica enrichment and clast elongation through mineral growth in pressure shadows (Fig. 1d,e).

Microphysical model for fault gouge strength

The observations on exhumed megathrust rocks indicate that one of the microscopic processes that controls macroscopic frictional behaviour is viscous pressure solution. Indeed, microphysical modelling studies have shown that experimental observations on shear deformation at low strain rates in rocks comprising rigid clasts in a phyllosilicate matrix can be explained by *frictional-viscous flow*: frictional sliding along cleavage planes coupled to viscous (time-dependent) pressure solution of intervening rigid clasts[23,24,25]. The microstructures reported in these experimental studies are essentially identical to those seen in samples from the exhumed Chrystalls Beach Complex (Fig. 1b-e). Frictional-viscous flow is restricted to low strain-rates (and/or high *T*); at higher strain rates (or lower *T*), slip is activated on anastomosing phyllosilicates, and microphysical models predict an importance of compaction by pressure solution[24,25]. Here, we use the model by Den Hartog and Spiers[25], coupled to analytical thermal gradients[26](Methods), to predict megathrust shear strength. This microphysical model is based on friction experiments performed on materials and at conditions representative for subduction megathrusts. Following this model, we assume a matrix-supported megathrust shear zone where frictional sliding occurs on aligned phyllosilicates, accommodated by pressure solution shear of intervening quartz grains or dilatation (Fig. 1f). In this model, the relation between shear strain rate and shear stress is derived by considering stress balances at the microscale for a unit cell defined in Fig. 1f. The megathrust shear strain rate ($\dot{\gamma}$) is related to the slip velocity (V) assuming strain is distributed through a shear zone thickness (w) and therefore $\dot{\gamma} = V/w$.

Each unit cell consists of quartz clasts, which are uniformly distributed such that horizontal rows overlap, and phyllosilicate foliations that are on average parallel to the shear plane, but locally curve around rigid clasts (ref. *25*; Fig. 1f; Supplementary Figure 3), resembling the natural microstructure (Fig. 1b). Slip along foliation is assumed to be a frictional process governed by the frictional resistance of phyllosilicates, which varies with temperature and normal stress according to experimental data for illite and muscovite[27-31](Methods). Depending on the conditions (e.g. slip velocity, temperature, normal stress), the frictional resistance predicted by the model either decreases (velocity-weakening) or increases (velocity-strengthening) as slip accelerates. Whereas velocity-weakening behaviour is potentially unstable, and can promote fast earthquake slip, velocity-strengthening behaviour is inferred to lead to stable sliding, recorded as aseismic creep[32-34]. In the microphysical model used here[25], velocity-strengthening flow occurs when easy shear of quartz clasts by thermally activated pressure solution, in series with rate-independent slip on planar phyllosilicates, leads to non-dilatant deformation (frictional-viscous flow). Velocity-weakening slip occurs when difficult pressure solution shear of quartz results in increased shear stress and slip is activated on curved phyllosilicate cleavages. This slip along curved foliation results in dilatation at the clast-matrix interface under extension (Fig. 1f), which at steady state is balanced by compaction via pressure solution.

Application of flow law to natural subduction zones

We apply boundary conditions appropriate for the northern Hikurangi margin, a megathrust shown to deform predominantly by aseismic creep, at least over the last few decades[11]. Pore fluid factors ($\lambda = P_f/\sigma_v$, where P_f is pore fluid pressure and σ_v is vertical stress) of 0.8 and 0.95 are imposed to test variations between moderate and high fluid pressure conditions. We distribute a steady creep rate of 40 mm yr^{-1} over a 1 - 100 m thick subduction thrust shear zone, a range representing strain rates from 10^{-11} to 10^{-9} s^{-1}, and a range in deforming thickness typical of exhumed mélanges and drilled subduction megathrusts[15]. Quartz grain size varies from 10 to 100 µm, based on Fig. 1b-e. All model parameters are listed in Supplementary Table S1.

The frictional-viscous flow strength of quartz-phyllosilicate mixtures as a function of depth is compared to frictional strengths of mono-mineralic quartz and illite-muscovite faults (Fig. 2a,b). At all considered conditions, frictional sliding in quartz requires higher shear stress than any slip mechanism in phyllosilicates or quartz-phyllosilicate mixtures; we therefore note that frictional sliding in quartz is an unlikely deformation mechanism in phyllosilicate-rich megathrust shear zones. For both high and moderate fluid overpressure, there is a depth below which frictional-viscous flow requires a lower shear stress than that required for frictional sliding in mono-mineralic phyllosilicate fault gouges (Fig. 2a,b). For deforming zones of 100 m thickness, frictional viscous flow becomes favourable at 8 – 10 km depth in moderate fluid pressure conditions (Fig. 2a), and at 12 – 16 km depth under high fluid pressure (Fig. 2b). In both cases, frictional-viscous flow becomes favourable at $T \geq 100 \pm 20$ ºC (Fig. 2c), where the corresponding shear stress, τ, is ≤ 10 MPa at high fluid pressure, and $\leq$ 20 MPa at moderate fluid pressure (Fig. 2a,b). For a 1 m thick deforming zone, higher strain rates make frictional-viscous flow less favourable; at high fluid pressure, frictional sliding of phyllosilicates remains favourable until a depth of ~ 26 km ($T < 200$ ˚C, $\tau < 20$ MPa), whereas at lower fluid pressures, frictional sliding also requires higher stresses and frictional-viscous flow becomes favourable from 16 km depth ($T < 150$ ˚C, $\tau \sim 40$ MPa).

Calculated temperatures define low thermal gradients, partly because very low stresses reduce temperatures relative to models with Byerlee friction (Fig. 2c). In our warmest model, where $\lambda = 0.8$, shear zone width is 1 m, and quartz grain size 100 µm, shear heating makes up approximately 30 % of the heat budget; for the coldest model, with λ = 0.95, shear zone width of 100 m, and quartz grain size 10 μm, less than 10 % of the heat budget is contributed by shear heating. Hikurangi is also a cool margin in the global spectrum of subduction zone thermal models, where model temperatures[35] compare to Fig. 2c. Compared to a recent numerical model[36], calculations here with λ = 0.8 are cooler at depths below ~ 10 km, whereas λ = 0.95 gives consistently lower temperatures.

Aseismic frictional-viscous flow is the predicted deformation style at $T \geq 100$ ˚C, when average plate boundary shear strain rates are accommodated in a hundreds of metres thick shear zone (Fig. 2a,b). Generation of run-away earthquake slip requires a change in these boundary conditions. This is because, at low strain rates, pressure solution of quartz clasts accommodates local finite strain around the rigid clasts created by slip on surrounding, planar phyllosilicate cleavages (Figs. 1d,e, 3a,b)[25]. At higher strain rate, pressure solution requires greater driving stress, bulk fault zone strength increases, and eventually dilatant, velocity-weakening behaviour occurs, allowing potentially unstable slip[25] (Fig. 3a,b). At each depth increment in Fig. 2, we calculate the friction coefficient as a function of strain rate, as shown for a depth of 30 km in Fig 3b. The strain rate required for a change from velocity-strengthening to velocity-weakening behaviour increases with depth (Fig. 3c). At depths greater than 15 km, where frictional-viscous flow generally becomes favourable (Fig. 2), velocity-strengthening behaviour occurs at strain rates slower than 10^{-12} s^{-1} and shear zone widths greater than tens of metres at 40 mm yr^{-1} slip rates (Fig. 3c). At a depth of 30 km, where frictional-viscous flow is preferred for all our considered conditions with a plate boundary slip rate (Fig. 2a,b), the shear strain rates required for velocity-weakening behaviour range from 10^{-9} to 10^{-4} s^{-1} (Fig. 3b,c).

At shallow depths, although commonly interpreted as a velocity-strengthening region[10,34], potentially seismic slip is predicted at strain rates as low as 10^{-12} s^{-1} at 5 km depth, and 10^{-16} s^{-1} at the surface (Fig. 3c). This is because shear deformation by pressure solution of quartz is difficult at low temperature, yielding dilatant behaviour. At greater depths, where $T \geq 100 \pm 20$ ºC, low strain rate frictional-viscous flow is the predicted deformation mechanisms (Fig. 2), because a high quartz solubility yields efficient dissolution and re-precipitation at this temperature (Ref. *37*, Supplementary Figure 4). This potential change in deformation mechanism is reflected in exhumed accretionary prisms, where mélange deformation at $T < 100$ ˚C is dominated by distributed cataclasis, whereas a pressure solution cleavage and localised slip surfaces are prevalent in rocks deformed at $T > 150$˚C (Refs. *6,21,38,39*). In central and northern Hikurangi, the margin we used for our thermal calculations, it is uncertain whether a near-surface velocity-strengthening zone and updip limit of seismicity is present, as slow slip events may propagate to the trench[40]; the downdip limit of the interseismically locked zone is here at less than 10 km depth[11]. This downdip limit of the locked zone is in agreement with the onset of velocity-strengthening frictional-viscous flow at 10 km depth and $T \leq 100$˚C, in a margin of moderate fluid overpressure and distributed shear (Fig. 2a,c).

Following Den Hartog and Spiers[25], we conclude that frictional-viscous flow involving pressure solution is a viable mechanism of velocity-strengthening, stable creep. We consider the recently discovered phenomenon of slow slip along subduction megathrusts[41,42], defined as geodetically observed displacement that is faster than plate convergence rates but too slow to generate seismic waves, as a form of unstable slip[43]. Shallow slow slip, as observed near the trench in northern Hikurangi[40], may therefore be a manifestation of unstable, dilatant shear at $T < 100^{\circ}$C (the 'potentially seismic slip' in Fig. 3c). Deeper slow slip events occurring down-dip of the locked zone and at depths ≥ 30 km, such as in Cascadia, are either independent of, or possibly load, the seismogenic region[44]. The application of the microphysical model predicts velocity-strengthening behaviour at such depths; thus, as for earthquakes, slow slip faster than steady-state plate convergence rates requires a local change in conditions, possibilities of which we discuss in the next section. Under the local triggering conditions, slow slip likely reflects competition between deformation modes within a heterogeneous fault zone[45], but may be an expression of either localised frictional sliding or distributed shearing flow; differentiating between these basic geometries requires currently missing knowledge of the deforming thickness during slow slip events.

Relating creep to subduction of rugged vs. smooth slab topography

Large earthquakes ($M_w \geq 8.0$) have been associated with subduction of smooth sea floor, because a lack of barriers to slip – such as local topography, seamounts, and horst-and-graben structures – allows for large rupture areas[2,46]. By comparison, subduction of rugged ocean floor has been suggested to lead to smaller earthquakes because rupture areas are geometrically constrained[2,12,36]. We therefore consider the implications of the model results for two end-member subducting plates, with (i) smooth and (ii) rugged topography[2,36,46].

(i) Smooth subducting slabs lack geometrical barriers to rupture propagation and the fault zone has similar thickness and strain rate at all depths (Fig. 4a). However, small-scale heterogeneities may locally elevate strain rates, causing velocity-weakening behaviour (Fig. 3b,c), and triggering rupture propagation over a large area without hindrance by large-scale barriers[4]. (ii) Rugged subducting ocean floor also deforms predominantly via creep by frictional-viscous flow, and small-scale heterogeneities may again lead to local velocity-weakening behaviour. However, in this case, strong, topographically derived irregularities on the interface create barriers to earthquake propagation, constraining earthquakes to smaller slip areas and therefore moderate magnitudes (Fig. 4b). At and around such barriers, local brittle deformation occurs to accommodate subduction of the topographic feature[2,12]. Extrapolating from continental strike slip faults[47], we suggest that geometrical barriers - such as deformed, subducting seamounts - that result in a discontinuity of potential slip surfaces by more than ~ 4 km, are likely to arrest rupture propagation. Moreover, because of numerous stress and strain-rate peaks, megathrusts associated with rugged subducting topography may appear strong in stress calculations from heat flow measurements[36] or Coulomb wedge mechanics[48], relative to fault segments where smooth subducting slabs allow large slip areas on a through-going weak surface or a system of anastomosing slip surfaces. A caveat to this broad, end-member interpretation is that subducting topography and megathrust structure may evolve with depth. For example, subducting seamounts may be progressively destroyed if they are indeed areas of increased brittle deformation[2], and the microscale geometry within the fault zone can change through development of through-going fault surfaces, mineral precipitation and reactions, and evolving grain shapes and sizes, through progressive deformation, metamorphism and fluid flow[49]. As such, the subduction thrust is a dynamic structure, displacing a footwall with inherently complex geometry, and accurate predictions require high-resolution subsurface data.

In summary, our model offers an explanation for why megathrusts creep in some places, and slip seismically in others. It implies that creep by frictional-viscous flow is the preferred deformation mechanism of most if not all subduction thrust interfaces, below some depth determined by thermal structure, strain rate, and fluid pressure (Figs. 2,3; Supplementary Figure 4). However, earthquakes may nucleate at local heterogeneities where the behaviour is velocity-weakening[50]. Slip area and earthquake magnitude should then depend on the spacing of strong, topographically derived irregularities in the subduction interface, with giant earthquakes requiring this spacing to be large.

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ACKNOWLEDGEMENTS: Initial work on this project was funded through NRF Incentive Funding for Rated Researchers to Å.F. while at the University of Cape Town. We thank Christel Tinguely (Cape Town) and Duncan Muir (Cardiff) for microprobe/SEM assistance, and Andrew Cross and three anonymous reviewers for constructive reviews that significantly improved the manuscript.

AUTHOR CONTRIBUTIONS: Both authors contributed to designing the study, making the calculations, and writing the paper. Å.F. provided the microstructural observations.

COMPETING INTERESTS: The authors declare that they have no competing financial interests.

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Figure captions

Figure 1: Example of pressure solution microstructures in a sample from the Chrystalls Beach Complex, New Zealand.

a, Photograph of outcrop-scale mélange shear zone with sandstone lenses in cleaved mudstone matrix. **b,** Photomicrograph (plane-polarised light) of sample from mélange matrix, cleavage wraps around quartz clasts. **c,** Close-up of cleavage seams, rectangle shows location of element maps of Si and Al; warm and cold colours show high and low relative abundance, respectively. **d** Backscatter electron image of quartz clasts in phyllosilicate matrix, accompanied by composite element map in **e**. **f,** Model microstructure where matrix (grey) deforms by frictional sliding along foliations (dashed lines), and clasts (black) deform by pressure solution[25]. All panels show dextral sense of shear.

Figure 2: Strength curves calculated along a subduction thrust interface with properties representative of the northern Hikurangi margin.

The pore fluid factor $\lambda = P_f/\sigma_v$, where P_f is pore fluid pressure and σ_v is vertical stress, is moderate (0.8) in **a** and high (0.95) in **b**. The curves labelled 'microphysical model' represent the strength of a fault where deformation occurs by slip on phyllosilicate surfaces and pressure solution of intervening quartz. Microphysical model predictions depend on grain size, D, and shear zone thickness, w, as shown in the legend, and thermal profiles as shown in **c**, including initial thermal structure where the frictional coefficient, μ, is 0.6. Supplementary Table S1 reports the full list of parameters.

Figure 3: Relations between slip velocity and frictional behaviour

a, Schematic relationship between friction coefficient and strain rate in the microphysical model used here, indicating a change from velocity-strengthening to velocity-weakening at high strain rate[25]. **b,** quantifies friction coefficient at a fixed depth, as a function of strain rate or shear zone width at a fixed slip velocity, whereas **c** combines calculations of friction coefficient vs. strain rate at all considered depths, to depict the depth-dependent strain rates where a change from velocity-strengthening to velocity-weakening is predicted (parameters as in Fig. 2).

Figure 4: Representations of the effect of frictional-viscous flow on megathrust seismic style.

Subduction interfaces related to smooth (**a**) and rugged (**b**) topography on subducting oceanic crust. Inferred transient and steady-state strain rate variations along these interfaces are shown below, as is the inferred depth vs. creep strength profile as based on Fig. 2 and down-dip variation in shear zone width.

Method

Element Maps

Element maps (Fig. 1c,d,e; Supplementary Figure 2) were plotted from energy dispersive spectroscopy (EDS) data, which give relative abundance of elements, measured on a carbon-coated, 30 μm thick sample. Maps in Fig. 1c were collected using an electronprobe microanalyser at the University of Cape Town, with beam conditions of 15 kV, 18.5 nA, 12 ms dwell time, and spot size of 1 μm. Electron backscatter images in Fig. 1d, and map in Fig. 1e were acquired using a Zeiss Sigma HD scanning electron microscope in the School of Earth & Ocean Sciences at Cardiff University. The EDS data for these element maps were acquired with an beam accelerating voltage of 20 kV, nominal beam current of 4.7 nA, and a 20 ms dwell time. Resulting pixels are approximately 1 μm.

Pressure-temperature estimates

To calculate the shear stress predicted by the microphysical model as a function of depth, approximations of temperature, T, and effective normal stress, σ_n', as functions of depth are required. Because the subduction thrust interface is gently dipping, σ_n' is approximated as the effective vertical stress[51,52] so that

$$\sigma_n' = \rho g z(1 - \lambda) \tag{1}$$

where ρ is the average density of overlying rock, taken as 2650 kg/m typical of quartzofeldspathic rocks, g is gravitational acceleration, λ is pore fluid factor as defined in the main text, and z is depth.

Temperature (in ˚C) is calculated according to the analytical derivation of Molnar and England[26], as also applied to the Hikurangi margin by McCaffrey et al.[53], that sums advective, radiogenic, and shear heating terms where

$$T = \frac{K_m}{SK_s}\frac{T_0 z}{\sqrt{\pi\kappa\left(t_0 + t_s\right)}} + \frac{A_r z^2}{2SK_s} + \frac{\tau V z}{SK_s} \quad (2)$$

in which the dimensionless parameter S is defined as

$$S = 1 + b\frac{K_m}{K_s}\sqrt{\frac{Vz\sin\delta}{\kappa}} \quad (3)$$

In these formulations, K_m and K_s are mantle and accretionary prism conductivities, respectively, T_0 is temperature at the base of the lithosphere, κ is thermal diffusivity, t_0 is the age of the subducting oceanic crust at the trench, A_r is average radioactive heat production rate in the forearc materials, τ is shear stress, and b is a geometrical factor. t_s is the time to subduct the slab to depth z, approximated as $t_s = z/\left(V\sin\delta\right)$ where V is slip velocity, assuming the megathrust accommodates the trench-normal component of the plate convergence vector, and δ is the average dip angle of the subduction thrust interface. Values for all the above parameters are listed in Supplementary Table 1. To obtain the shear heating term in the initial thermal structure, τ is estimated as σ_n' multiplied by a frictional coefficient of μ = 0.6, estimating the lower end of the Byerlee range[54]. After calculating shear stress according to the microphysical model, the calculated shear stress as a function of depth is used to re-calculate the thermal structure, which is then used to re-calculate shear stress. The change in thermal structure from the first calculation to calculations involving shear stresses from the microphysical model can be seen in Fig. 2c.

Temperature-dependent mono-mineralic friction

Values for the friction coefficient for phyllosilicates were determined assuming the dominant phyllosilicate mineral to be illite and muscovite at temperatures below and above 300 °C, respectively, and by assuming that temperature rather than effective normal stress dominantly affects the friction coefficient. The friction coefficient of illite as a function of temperature was determined by fitting a linear trend line to a combination of the data by Tembe et al.[55] at 20°C and the data of Den Hartog et al.[27] at 200, 350 and 500°C, all representing final friction values (at 9.21 and ~40 mm shear displacement, respectively) at a sliding velocity of 1 μm/s. Note that the sliding velocity at ~40 mm shear displacement in the experiments of Den Hartog et al.[27] was 10 μm/s, and we thus recalculated it to 1 μm/s using the value for $\Delta\mu/\Delta\ln V$, or (a-b), for a velocity step from 10 to 1 μm/s obtained in the same experiment. Similarly, the friction coefficient of muscovite as a function of temperature was determined by fitting two linear trendlines (joining at 600°C) to the data by Den Hartog et al.[28] at 200, 400 and 600°C and the data by Van Diggelen et al.[29] at 400, 500 and 700°C. These data represent close to final friction coefficients, those by Den Hartog et al.[28] taken at a shear strain of 50 and recalculated for 1 μm/s by the method described for illite and those by Van Diggelen et al.[29] reported for the 0.5 μm/s step, which occurred at near steady state friction. The resultant empirical function for phyllosilicate friction coefficient, μ_{ph}, becomes

$$\mu_{ph}\begin{cases} 0.320+9.10\times10^{-4}T, & T<300^{\circ}\mathrm{C} \\ 0.300+6.18\times10^{-4}T, & 300^{\circ}\mathrm{C}\leq T<600^{\circ}\mathrm{C} \\ 1.997-2.24\times10^{-3}T, & T\geq 600^{\circ}\mathrm{C} \end{cases} \tag{4}$$

This definition for the phyllosilicate friction coefficient was used to construct the strength profiles for pure phyllosilicates and as input to the microphysical model.

The friction coefficient of quartz, for plotting the frictional strength of mono-mineralic quartz aggregates in Fig. 2a,b, is estimated based on the room temperature data of Tembe et al.[55], data at 140 ˚C of Den Hartog and Spiers[30] and the data at 400-600 ˚C of Niemeijer et al.[31]. Based on similar arguments as for creating an empirical function of phyllosilicate friction as a function of temperature, we obtain a function for quartz friction, μ_{qtz}:

$$\mu_{qtz}\begin{cases} 0.750-1.04\times10^{-4}T, & T<500^{\circ}\mathrm{C} \\ 1.41-1.43\times10^{-3}T, & T\geq 500^{\circ}\mathrm{C} \end{cases} \tag{5}$$

Microphysical model by Den Hartog and Spiers[25]

The microphysical model used to calculate the strength profiles (Fig. 2) was derived by Den Hartog and Spiers[25]. The model describes the steady state frictional behaviour of sheared illite-quartz mixtures, and assumes a matrix-supported shear zone consisting of phyllosilicates and quartz clasts (Supplementary Figure 3). The quartz clasts are uniformly distributed, arranged such that horizontal rows of clasts overlap. On average, the phyllosilicates are aligned parallel to Y-shear bands, but locally anastomose around the rigid clasts. Note that the Y-shear bands considered in the model by Den Hartog and Spiers[25] will on average be parallel to the megathrust interface, which implies that foliation that is parallel to these shear bands, described as "horizontal" in the model, will be gently dipping in the megathrust setting.

Within the model microstructure, shear deformation occurs either within the "clast body" zones containing a horizontal phyllosilicate foliation and quartz clasts (Type B zones, Supplementary Figure 3) or in the "clast overlap" regions containing anastomosing phyllosilicates and overlapping quartz clast edges (Type O zones, Supplementary Figure 3). The horizontal foliation in the Type B zones abuts against the quartz clasts, so that sliding on this foliation requires serial simple shear of the clast "bodies". Shear of the clasts is assumed to occur by thermally activated deformation. By contrast, in the Type O zones, the foliation anastomoses around the clast "overlaps". In these zones, deformation can occur either by slip on the phyllosilicates at the zone margins accommodated by shearing of the clast overlaps, or by slip on the curved foliation accompanied by dilatation at extensional clast-matrix interface sites. Sliding on the foliation is assumed to be a purely frictional process, which implies that slip on the curved foliation will not occur unless a critical value of the macroscopic shear stress, τ_{dil}, is attained. When slip is activated, it will cause dilatation and porosity development. Den Hartog and Spiers[25] assumed that developing porosity concentrates at the extensional quartz-illite interfaces (Supplementary Figure 3), resulting in a decrease in the clast overlap distance, and hence in the mean inclination of the curved foliation. This in turn causes a decrease in the rate of dilation per unit horizontal displacement on the inclined foliation, i.e. a decrease in the dilatation angle ψ_{dil}, with increasing porosity. Den Hartog and Spiers[25] assumed that the appearance of porosity, via clast/matrix debonding, initiates compaction by thermally activated deformation of the clasts, which accelerates as porosity increases. At steady state, dilation due to slip on the curved foliation and compaction by the thermally activated mechanism must balance. This competition between dilatation and compaction is of key importance since it will lead to higher steady state porosities, a flatter foliation and lower frictional strength as sliding velocity increases, and hence to velocity-weakening slip. This as opposed to non-dilatant deformation, where the serial nature of deformation implies that the velocity-dependence of friction is governed by thermally activated deformation of the quartz clasts which is by definition velocity-strengthening. Dilatation, when active, is assumed to continue until a limiting or critical state porosity is reached.

The model by Den Hartog and Spiers[25] does not strictly apply to muscovite. However, in the absence of a microphysical model for the steady state frictional behaviour of muscovite-quartz fault gouge, and since muscovite-quartz gouge shows broadly similar behaviour to illite-quartz gouge[28], we have applied this model also at temperatures >300°C where muscovite is expected to be the dominant phyllosilicate.

Model calculations

The reader is referred to Den Hartog and Spiers[25] for the derivation of the equations governing the steady state frictional behaviour of the model microstructure shown in Supplementary Figure 3 and described below.

Den Hartog and Spiers[25] derive their equations for the unit cell shown in Supplementary Figure 3b, which has a horizontal dimension equal to horizontal clast spacing:

$$L = \frac{k_f \pi D^2}{\left(D - x_0\right) f_{qtz}} \qquad (6)$$

where k_f is a factor accounting for clast shape, D is grain size (clast diameter), f_{qtz} is the volume fraction of quartz clasts, and x_0 is the vertical overlap of the clasts at zero porosity defined by Den Hartog and Spiers[25] as

$$x_0 = D\left(1 - \sqrt{\frac{k_f \pi}{2 f_{qtz}}}\right) \qquad (7)$$

As porosity, ϕ, increases due to dilatational slip on the curved foliation, this overlap decreases from x_0 to an instantaneous value x according to the relation $x = (x_0 - \phi D]/(1-\phi)$. The decrease in overlap in turn leads to a decrease in the width, d, of overlapping clast segments (Supplementary Figure 3b), given $d = 2\sqrt{(Dx-x^2)}$.

During non-dilatant deformation at low slip velocities and/or high temperatures, thermally activated shear deformation of the quartz clasts will be easy. The total resistance to slip on the horizontal foliation will then be lower than the shear stress to activate slip and dilatation on the anastomosing foliation. Under these conditions, Den Hartog and Spiers[25] assumed that non-dilatant deformation takes place by the parallel processes of (i) slip on the horizontal foliation with serial shear of the clast bodies in the B zones of the microstructure plus (ii) slip on the horizontal phyllosilicates with serial shear of clast overlaps at the margin of the O zones. Equilibrium between the shear stresses supported by the B and O zones (τ_B and τ_O, respectively) requires $\tau_m = \tau_B = \tau_O$ where τ_m is the macroscopic shear stress. The shear stresses in the B and O zones were derived by Den Hartog and Spiers[25]:

$$\tau_B = \tau_{ph}\left(1 - \frac{A_{qtz-b}}{LD}\right) + \tau_{qtz-b}\frac{A_{qtz-b}}{LD} \quad (8)$$

$$\tau_O = \tau_{ph}\left(1 - \frac{A_{qtz-o}}{LD}\right) + \tau_{qtz-o}\frac{A_{qtz-o}}{LD} \quad (9)$$

where τ_{ph} is the shear stress needed to drive frictional slip on the horizontal phyllosilicate foliation and τ_{qtz-b} and τ_{qtz-o} are those needed to drive thermally activated clast body and overlap deformation, respectively. A_{qtz-b} represents the average horizontal area occupied by a single clast body within zone B of the unit cell, and is given $A_{qtz-b} = [(\frac{1}{4}\pi D^2 - 2A'_{seg})D]/(D - 2x)$, where $A'_{seg} = [16x^2(D - x) + 3x^3]/[12\sqrt{(Dx - x^2)}]$ (Ref. *56*) is the area of an individual clast segment located in the overlap zone of the cell in the plane of Supplementary Fig. 3 and $A_{qtz-o} = dD = 2D\sqrt{(Dx - x^2)}$ is the area over which the overlap is displaced by slip at its base. Note that $\tau_{ph} = \mu_{ph}\sigma_n'$ where μ_{ph} is defined by equation (4).

The parallel shear processes (i, ii) operating in the O and B zones mean that the total, measured shear strain rate during non-dilatant deformation is $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O$, where $\dot{\gamma}_B$ and $\dot{\gamma}_O$ denote the shear strain rate contributed to the unit cell by each zone respectively (i.e. $\dot{\gamma}_B$ and $\dot{\gamma}_O$ are determined by taking into account the thickness of the B or O zone relative to the unit cell thickness). Note that the serial coupling of rate-independent slip on the phyllosilicates with thermally activated deformation of clasts implies that $\dot{\gamma}_B = \dot{\gamma}_{qtz-b}$ and $\dot{\gamma}_O = \dot{\gamma}_{qtz-o}$, where, $\dot{\gamma}_{qtz-b}$ and $\dot{\gamma}_{qtz-o}$ are the shear strain rate contributions to the unit cell due to thermally activated deformation of the clast bodies and clast overlaps, respectively. Thermally activated deformation was assumed to occur via pressure solution by Den Hartog and Spiers[25], yielding:

$$\dot{\gamma}_{qtz-b} = \frac{AI\tau_{qtz-b}\Omega}{RT}\frac{D-2x}{D(D-x)} \tag{10}$$

$$\dot{\gamma}_{qtz-o} = \frac{2I\tau_{qtz-o}\Omega}{RT}\frac{1}{\sqrt{Dx-x^2}} \tag{11}$$

Where A is a shape factor, I is the product of the dissolution rate coefficient k_+ and molar volume Ω of quartz, and R is the gas constant.

Following Den Hartog and Spiers[25], we obtained τ_m as a function of $\dot{\gamma}_m$, by first imposing $\dot{\gamma}_m$, defined as $\dot{\gamma}_m = V / w$ where w is the shear zone width. We next solved $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O$ together with $\tau_m = \tau_B = \tau_O$ to obtain $\dot{\gamma}_B$ or $\dot{\gamma}_{qtz-b}$. We subsequently used $\dot{\gamma}_{qtz-b}$ to determine τ_{qtz-b} via equation (10). The value of τ_{qtz-b} obtained, then yielded $\tau_B = \tau_m$ through equation (8). Note that in the current calculations we prevented $\dot{\gamma}_O$ from taking a negative value in the non-dilatant regime[57].

At high slip rates or low temperatures, thermally activated shear deformation of the quartz clasts is difficult, leading to an increase in the total resistance to shear on the horizontal foliation. In the model microstructure of Den Hartog and Spiers[25] this would ultimately activate slip on the curved phyllosilicates in the overlap (O) zones of the microstructure. The measured shear strength in that case is equal to that required to activate slip on the anastomosing foliation, τ_{dil}, derived by Den Hartog and Spiers[25] to be

$$\tau_{dil} = \left\{ \frac{\mu_{ph}\left(1+\tan^2 \Psi_{fr}\right)}{1-\mu_{ph}^2 \tan^2 \Psi_{fr}} \right\} \sigma_n' \tag{12}$$

where $\tan\Psi_{fr}$ is a straight line approximation of the curved foliation, i.e.

$$\tan \Psi_{fr} = \frac{2\left(D-x_0\right) f_{qtz}}{k_f \pi D^2} x \tag{13}$$

Stress equilibrium between B and O zones means that in the dilatant case $\tau_m = \tau_{dil} = \tau_B = \tau_O$. The total shear strain rate $\dot{\gamma}_m$, in turn, is given $\dot{\gamma}_m = \dot{\gamma}_B + \dot{\gamma}_O + \dot{\gamma}_{dil}$, or equivalently $\dot{\gamma}_m = \dot{\gamma}_{qtz-b} + \dot{\gamma}_{qtz-o} + \dot{\gamma}_{dil}$, where $\dot{\gamma}_{dil}$ is the shear strain rate contribution to the unit cell by dilatant slip on the curved phyllosilicates. This mechanism produces an associated dilational strain rate, $\dot{\varepsilon}_{dil}$, which Den Hartog and Spiers[25] defined following the classical soil mechanics approach to granular flow, i.e.

$$\dot{\varepsilon}_{dil} = \left(\frac{\mathrm{d}\varepsilon_{dil}}{\mathrm{d}\gamma_{dil}} \right) \frac{\mathrm{d}\gamma_{dil}}{\mathrm{d}t} = \left(\tan \Psi_{dil}\right) \dot{\gamma}_{dil} \tag{14}$$

Den Hartog and Spiers[25] defined the dilatation angle Ψ_{dil} as the steepest portion of the curved, i.e. sinusoidal, foliation:

$$\tan \Psi_{dil} = \sqrt{\frac{\pi f_{qtz}}{2k_f}} - \frac{\pi}{2(1-\phi)} \tag{15}$$

This angle (Ψ_{dil}) decreases with increasing porosity, reaching zero at a limiting or "critical state" porosity, defined $\phi_c = x_0/D$ when $x = 0$.

The porosity generated by dilatant slip will induce compaction by thermally activated deformation of the quartz clasts at a rate $\dot{\varepsilon}_{comp}$. Taking compaction as positive, the total, measured compaction strain rate is therefore given $\dot{\varepsilon}_m = \dot{\varepsilon}_{comp} - \dot{\varepsilon}_{dil}$. At steady state, dilatation and compaction must balance, resulting in a steady state porosity corresponding to the condition that $\dot{\varepsilon}_m = 0$ or $\dot{\varepsilon}_{comp} = \dot{\varepsilon}_{dil}$. Following Den Hartog and Spiers[25], $\dot{\varepsilon}_{comp}$ is given by:

$$\dot{\varepsilon}_{comp} = \frac{2I\sigma_n'\Omega}{RT}\frac{A_{pore}}{(D-x)DL} \tag{16}$$

Compaction occurs by pressure solution transfer from compressively stressed illite-quartz interfaces to debonded (dilated) interfaces (pore walls) with surface area A_{pore}, written $A_{pore} = (A_{pore\text{-}c}/2)(\phi/\phi_c)^n$ where ϕ_c and $A_{pore\text{-}c}$ are the porosity and pore area per clast at the critical state. Den Hartog and Spiers[25] derived that $A_{pore\text{-}c} = (\pi D^2)/2$.

To calculate τ_m as a function of $\dot{\gamma}_m$ in the dilatant regime, we followed the procedure by Den Hartog and Spiers[25] and incremented the porosity from 0 to ϕ_c and calculated the corresponding values of $\tan\Psi_{dil}$ and $\tan\Psi_{fr}$ using equations (15) and (13). Using $\tan\Psi_{fr}$, equation (12) gives τ_{dil}. The corresponding shear strain rate $\dot{\gamma}_m$ is calculated via $\dot{\gamma}_m = \dot{\gamma}_{qtz-b} + \dot{\gamma}_{qtz-o} + \dot{\gamma}_{dil}$ and using the flow laws in equations (10) and (11). Here, $\dot{\gamma}_{dil}$ is obtained via equation (14) and using the steady state condition $\dot{\varepsilon}_{comp} = \dot{\varepsilon}_{dil}$, where $\dot{\varepsilon}_{comp}$ is calculated using equation (16).

In our calculations, we assumed cylindrical quartz clasts (k_f= 0.25) of either 10 or 100 μm in diameter, taking up a volume fraction of 0.45. Following Den Hartog and Spiers[25], we assume that pressure solution is controlled by the interfacial reactions of dissolution and precipitation and can be described using the empirical equation for the dissolution rate coefficient provided by Tester et al.[37]:

$$k_+ = 276 \exp\left(\frac{-90100}{RT}\right) \quad (17)$$

with T in Kelvin. We used a shape factor A of π in our calculation of the clast body shear strain rate, while a factor of 2 was used in the original model. We also follow the assumption that the porosity can be characterised by an exponent n of 0.3 (Ref. *25*).

To determine our shear strength versus depth profiles predicted by the microphysical model, we selected σ_n', T and the corresponding μ_{ph} at each depth. Using this input, we obtained τ_m as a function of $\dot{\gamma}_m$ (incorporating both non-dilatant and dilatant deformation) following the above procedure. We next used the assumed subduction velocity of 40 mm/yr and shear zone thickness (1 to 100 m in the current calculations) to select relevant $\dot{\gamma}_m$ and determined τ_m at that shear strain rate.

The result of our calculations, shown in Fig. 2, yield dilatant deformation at shallow depths and low temperatures for the shear strain rates explored in this study.

With increasing depth and temperature shear deformation of the quartz clasts by pressure solution becomes easier, resulting in a transition to non-dilatant deformation. To illustrate this effect, we show the calculated values of τ_{ph}, $\tau_{qtz\text{-}b}$ (for non-dilatant shear), and the inferred shear stress as a function of depth, for the scenario where D is 100 μm, w is 100 m, and λ is 0.95, in Supplementary Figure 4. For any given set of conditions, the transition to non-dilatant deformation depends on strain rate, and we plot the strain rate at which the transition occurs, as a function of depth, in Fig. 3c.

Code and data availability

Code and additional data are available from the authors on request.

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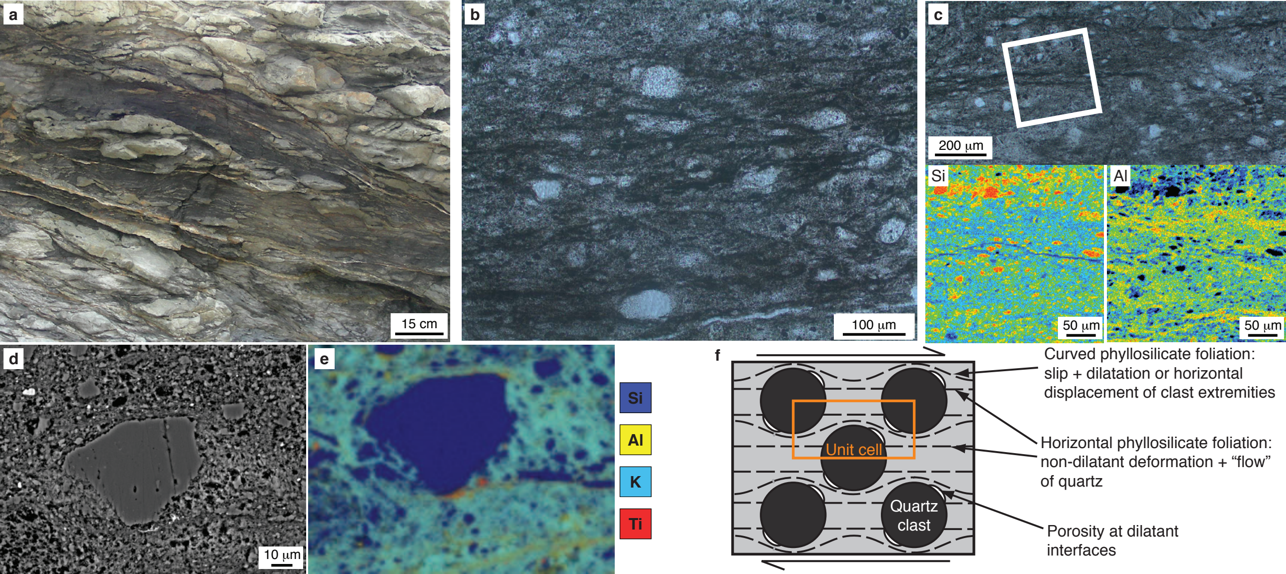
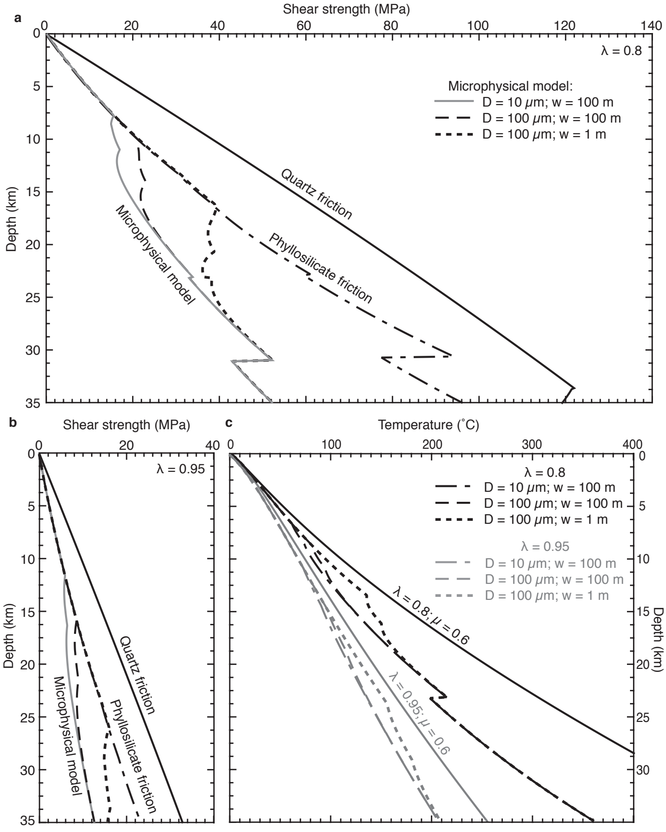
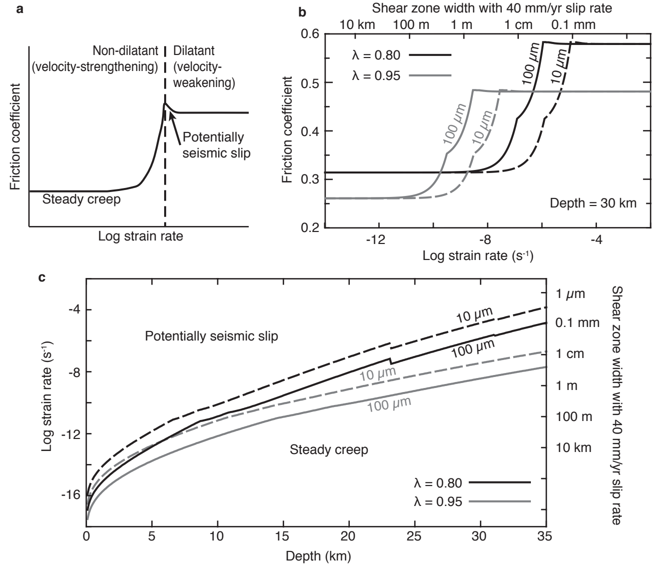
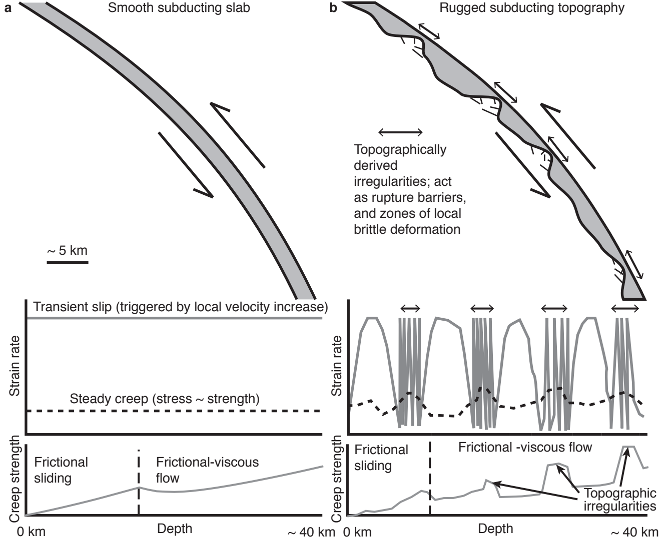

a
15 cm
b
100 μm
c
200 μm
Si
50 μm
Al
50 μm
d
10 μm
e
Si
Al
K
Ti
f
Unit cell
Quartz clast
Curved phyllosilicate foliation: slip + dilatation or horizontal displacement of clast extremities
Horizontal phyllosilicate foliation: non-dilatant deformation + "flow" of quartz
Porosity at dilatant interfaces


a
Shear strength (MPa)
Depth (km)
λ = 0.8
Microphysical model:
D = 10 μm; w = 100 m
D = 100 μm; w = 100 m
D = 100 μm; w = 1 m
Quartz friction
Phyllosilicate friction
Microphysical model
b
Shear strength (MPa)
λ = 0.95
Quartz friction
Phyllosilicate friction
Microphysical model
c
Temperature (˚C)
λ = 0.8
D = 10 μm; w = 100 m
D = 100 μm; w = 100 m
D = 100 μm; w = 1 m
λ = 0.95
D = 10 μm; w = 100 m
D = 100 μm; w = 100 m
D = 100 μm; w = 1 m
λ = 0.8; μ = 0.6
λ = 0.95; μ = 0.6



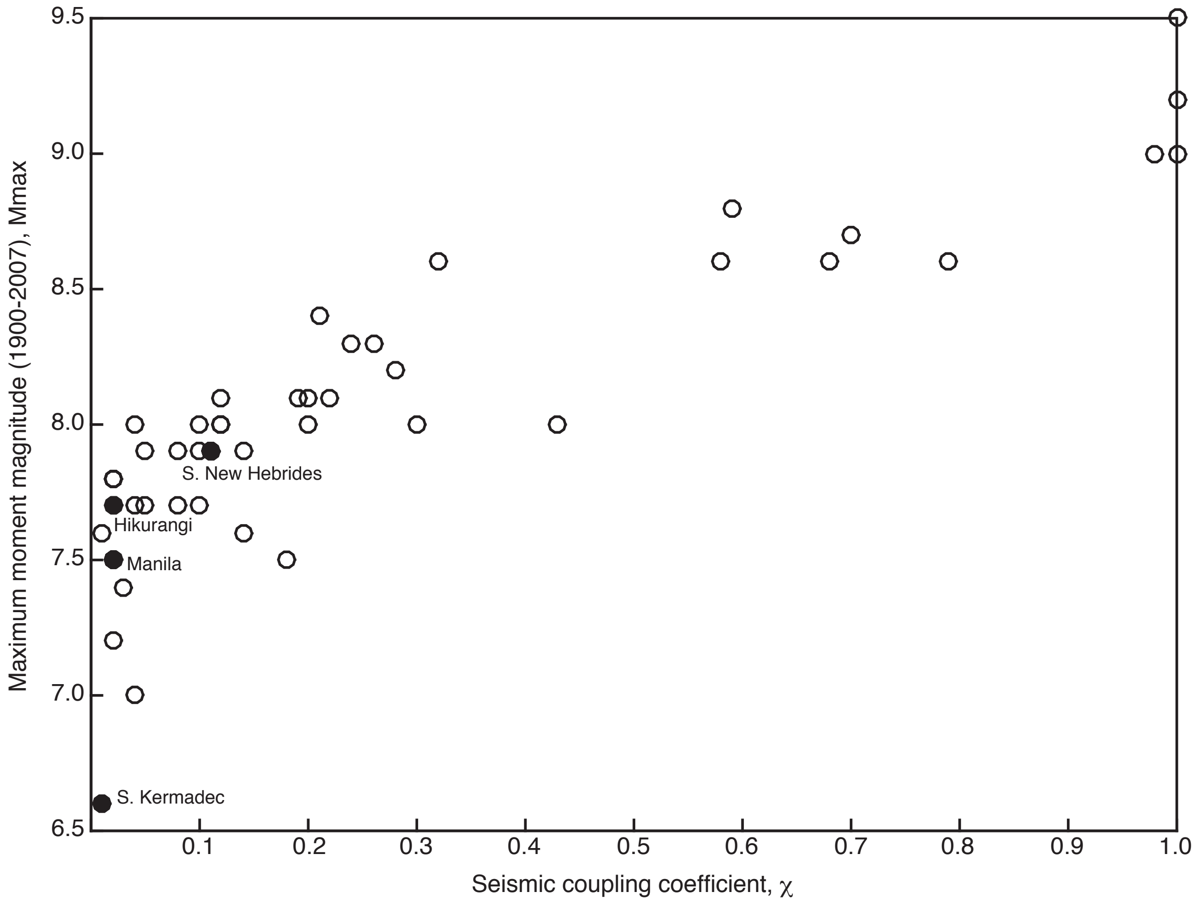
a
Non-dilatant (velocity-strengthening)
Dilatant (velocity-weakening)
Friction coefficient
Potentially seismic slip
Steady creep
Log strain rate
b
Shear zone width with 40 mm/yr slip rate
10 km
100 m
1 m
1 cm
0.1 mm
λ = 0.80
λ = 0.95
100 μm
10 μm
Friction coefficient
Depth = 30 km
Log strain rate (s-1)
c
Potentially seismic slip
Steady creep
Log strain rate (s-1)
Shear zone width with 40 mm/yr slip rate
1 μm
0.1 mm
1 cm
1 m
100 m
10 km
10 μm
100 μm
λ = 0.80
λ = 0.95
Depth (km)



a
Smooth subducting slab
~ 5 km
Transient slip (triggered by local velocity increase)
Strain rate
Steady creep (stress ~ strength)
Creep strength
Frictional sliding
Frictional-viscous flow
0 km
Depth
~ 40 km
b
Rugged subducting topography
Topographically derived irregularities; act as rupture barriers, and zones of local brittle deformation
Strain rate
Creep strength
Frictional sliding
Frictional -viscous flow
Topographic irregularities
0 km
Depth
~ 40 km

Subduction megathrust creep governed by pressure solution and frictional-viscous flow

Åke Fagereng and Sabine A.M. den Hartog



Supplementary Figure S1: Plot of maximum moment magnitude, *Mmax*, against seismic coupling coefficient, χ. The data are from the compilation of Heuret et al. (2011), but limiting the maximum seismic coupling coefficient to 1.0. Examples in the main text are highlighted in solid circles. Note that Hikurangi in this plot includes both northern and southern Hikurangi, and that the M_w7.7 event in the Heuret et al. (2011) compilation may have included significant slip on a splay fault in the overlying accretionary prism (Wallace et al., 2009). Thus, both the coupling coefficient and the maximum magnitude may be overestimated. The southern Japan Trench is not highlighted despite being mentioned in the main text, as the area referred to is relatively small, but described in detail by Mochizuki et al. (2008); it is not added to retain consistency in the plotted data.

Data in Fig. S1 were compiled by Heuret et al. (2011) using data from the Harvard CMT catalogue for $M_w \geq 5.5$ earthquakes from the 1976-2007 time period, including 1900-1975 for $M_w \geq 7.0$ events in the Centennial catalogue of (Engdahl and Villasenor, 2002). Earthquake locations were, if possible, relocated from the EHB catalogue of Engdahl et al. (1998). Thus, Heuret et al. (2011) extracted earthquakes with locations and, if available, nodal planes that align with the subduction thrust interface. From this data set, they defined the seismogenic zone of a number of megathrust interfaces, 49 for which they provide both *Mmax* and χ.

For the 49 subduction interfaces plotted in Fig. S1, *Mmax* is the largest subduction thrust earthquake identified in the Heuret et al. (2011) compilation, i.e. that occurred between 1900 and 2007, and fell on the inferred megathrust interface. We note that overestimates may occur, through inclusion of poorly located events, that were not actually megathrust events, particularly events prior to 1964 that were not relocated in the EHB catalogue.

To calculate χ, the amount of seismic slip and the rate of plate convergence must be estimated for each region. Defining seismic moment of a single earthquake as $M_0 = GLWu$, where G is shear modulus (50 GPa), L and W are the length and width of the rupture area, and u is average slip, the seismic slip rate for a time period T is $v_s = Su/T = SM_0/(GLWT)$ (Brune, 1968). The seismic coupling coefficient, χ, can be defined as the ratio of v_s to the subduction velocity as defined by global plate kinematic models, and was calculated accordingly by Heuret et al. (2011).

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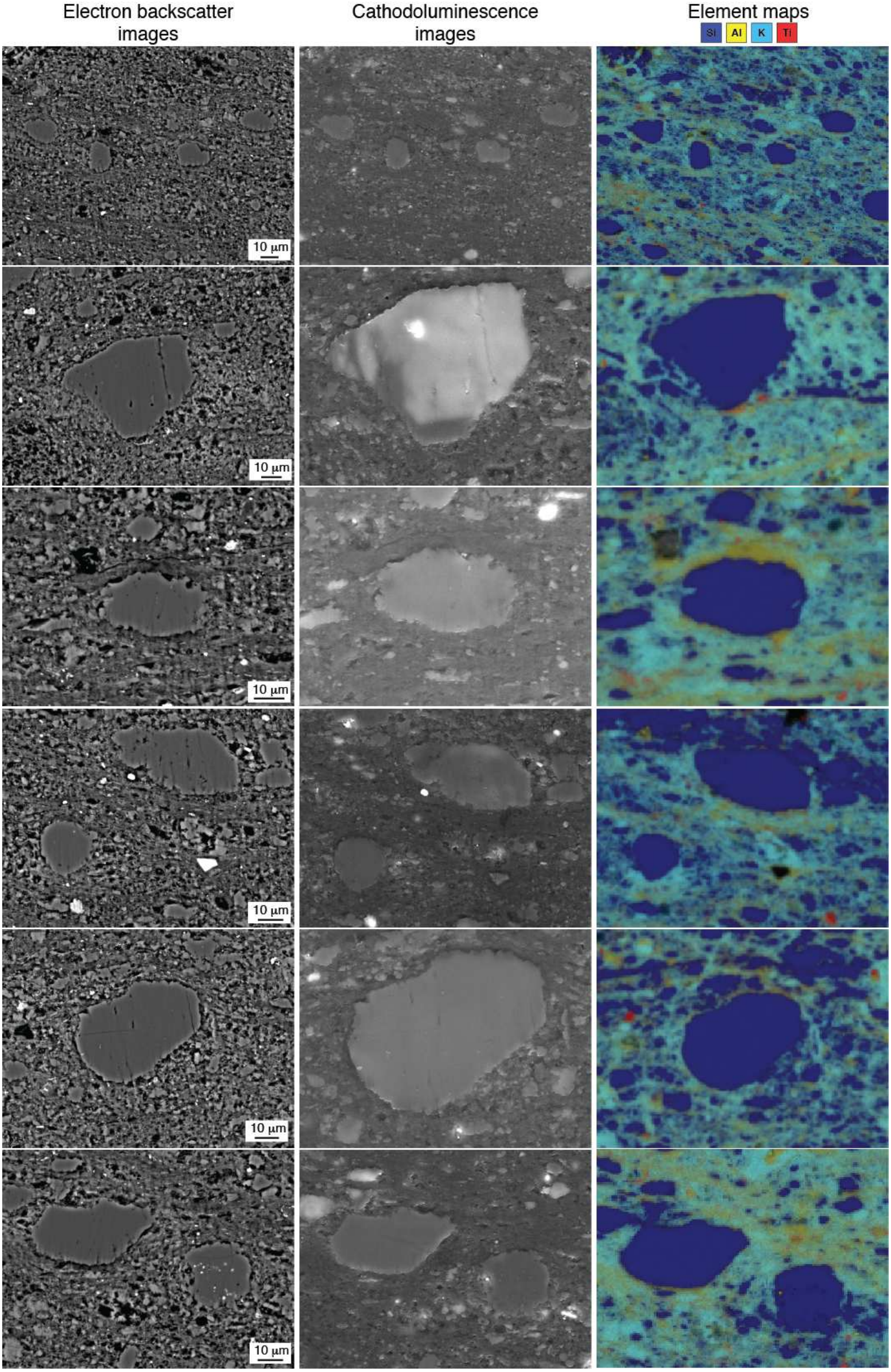
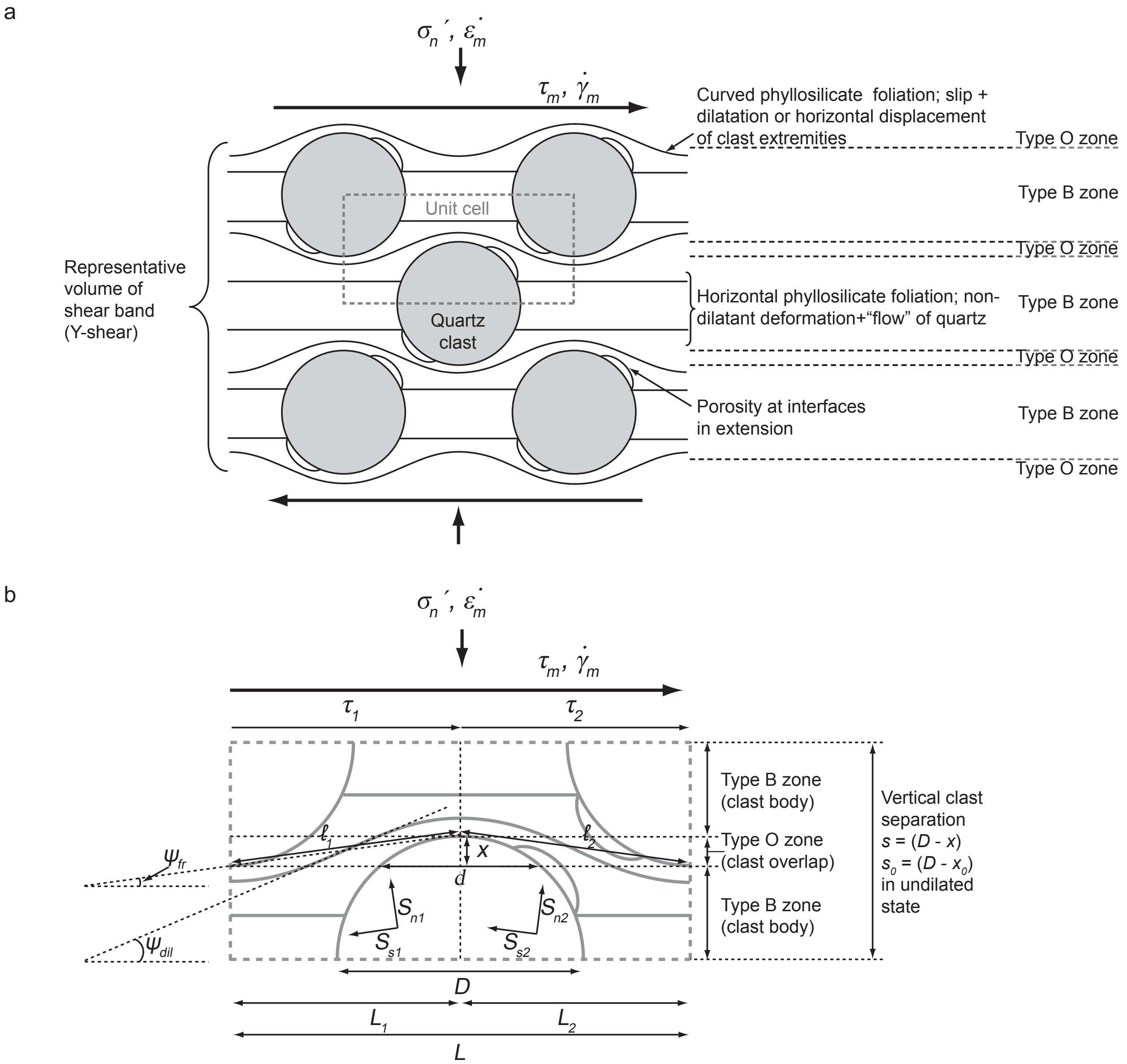
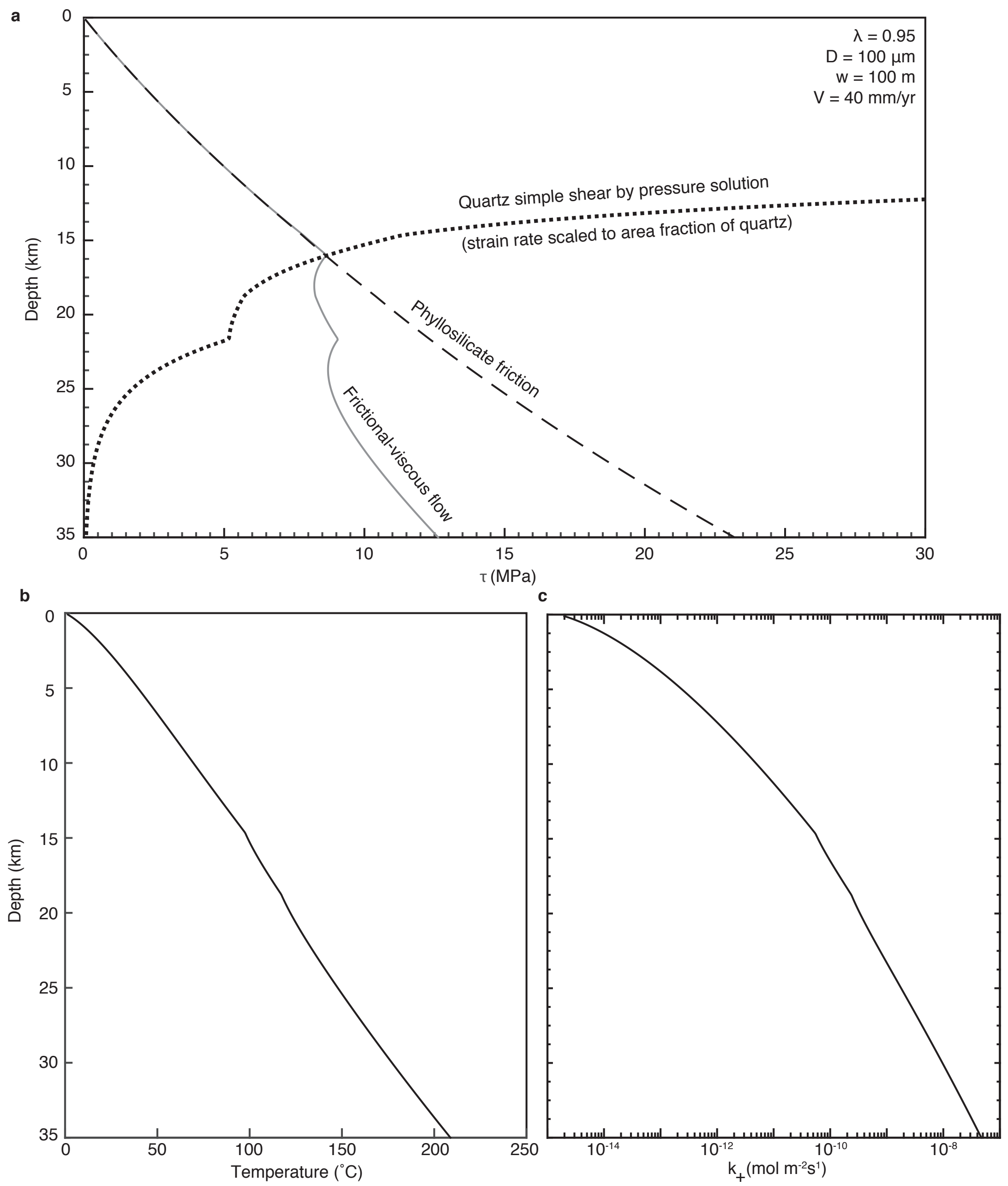


Supplementary Figure S2: Scanning electron microscope images of Chrystalls Beach sample. Minor pore space development in pressure shadows can be seen in BSE images in the left column, and silica enrichment in pressure shadows can be discerned from EDS maps in the right column. The middle column shows little CL variation within most quartz clasts.



Supplementary Figure S3: Model microstructure assumed for phyllosilicate-quartz gouge undergoing shear deformation. **a**. Shows the representative microstructure, and the definition of clast body zones (B), clast overlap zones (O), and the unit cell as referred to in the method, and originally defined by Den Hartog and Spiers (2014, *25* in main text). In this figure, σ_n' is the effective normal stress, τ_m is the macroscopic shear stress, $\dot{\gamma}_m$ is the macroscopic or imposed shear strain rate and $\dot{\varepsilon}_m$ is the macroscopic rate of compaction. **b.** Shows the definition of the microstructural variables; the key variables are referred to and defined in the Methods. Note that the curved foliation is drawn with an exaggerated amplitude for clarity. Figure taken from Den Hartog and Spiers (2014, Ref *25* in the main text).



Supplementary Figure S4: Figure showing details of calculated parameters, for the scenario where quartz clast size, D, is 100 µm, shear zone width, w, is 100 m, and the pore fluid factor, λ, is 0.95. **a.** calculated values of phyllosilicate frictional resistance, τ_{ph}, shear stress required for simple shear of quartz clast bodies, $\tau_{qtz\text{-}b}$ (for non-dilatant shear), and the inferred shear stress for frictional-viscous flow at the given strain rate of v/w = 1.3×10^{-11} s^{-1}, as a function of depth. **b.** Temperature as a function of depth. **c.** Dissolution rate constant for quartz in water, k_{+}, as a function of depth.

Supplementary Table S1 List of parameter values used in the calculations performed in this paper.

Parameter	Meaning	Value
	Temperature and stress	
A_r	Average radiogenic heat production in the forearc	10^{-6} W m^{-3}
b	Geometric constant	1.0
K_m	Mantle conductivity	3.3 W m^{-1} K^{-1}
K_s	Accretionary prism conductivity	2.55 W m^{-1} K^{-1}
t_0	Age of subducting oceanic crust at the trench	80 Ma
T_0	Temperature at the base of the lithosphere	1300 ˚C
V	Average fault slip rate	40 mm yr^{-1}
δ	Average dip of subduction thrust interface	15˚
λ	Pore fluid factor	0.8 or 0.95
κ	Thermal diffusivity	10^{-6}
ρ	Average density above the shear zone	2650 kg m^{-3}
	Microphysical model	
A	Shape factor in the clast body shear strain rate equation	π
D	Clast diameter (grain size)	10 or 100 μm
f_{qtz}	Volume fraction of quartz	0.45
k_f	Constant depending on the 3-D clast shape, ¼ for cylinder	0.25
k_+	Dissolution rate coefficient	= 276 × exp(-90100/[R x T (K)]) mol m^{-2} s^{-1}
n	Exponent in relation describing pore area, pore shape evolution parameter	0.3
R	Universal gas constant	8.31462 J mol^{-1} K^{-1}
w	Average shear zone thickness	1 - 100 m
μ_{ph}	Friction coefficient within phyllosilicates	0-300°C: = 0.3199 + 9.101 x 10^{-4} T (°C) 300-600°C: = 0.2997 + 6.180 x 10^{-4} T (°C) 600-700°C: = 1.9967 – 2.244 x 10^{-3} T (°C)
Ω	Molar volume of quartz	2.27×10^{-5} m^3 mol^{-1}