Seismological data from recent subduction earthquakes suggest that megathrust earthquakes induce transient stress changes in the upper plate that shift accretionary wedges into an unstable state. These stress changes have, however, never been linked to geological structures preserved in fossil accretionary complexes. The importance of coseismically induced wedge failure has therefore remained largely elusive. Here we show that brittle faulting and vein formation in the palaeo-accretionary complex of the European Alps record stress changes generated by subduction-related earthquakes. Early veins formed at shallow levels by bedding-parallel shear during coseismic compression of the outer wedge. In contrast, subsequent vein formation occurred by normal faulting and extensional fracturing at deeper levels in response to coseismic extension of the inner wedge. Our study demonstrates how mineral veins can be used to reveal the dynamics of outer and inner wedges, which respond in opposite ways to megathrust earthquakes by compressional and extensional faulting, respectively.
The high relief of about 2,600 m provides unique insights into the internal architecture of this orogenic wedge. The area comprises an exhumed ~4-km-thick sequence of hemipelagic shales, marls and limestones, as well as syn-orogenic turbidites (flysch) including volcanic detritus. The sediments were deposited in an underfilled foreland basin and show a similar sequence of thrusts on top of the IFUs since the Oligocene. Peak and overlying nappes that were transported along out-of-plane thrusts ceased in the early Miocene when the coupling along the part of the wedge (Fig. 1b). The prograde evolution of the IFU transport to greater depths due to continuous plate convergence come from active convergent margins. Geodetic and seismological data recorded along the Chilean and Japanese forearc documented a significant increase in upper-plate seismicity after the 2010 Mw 8.8 Maule and the 2011 Mw 9.0 Tohoku earthquakes, and revealed that coseismically induced stress changes activate splay and normal faults in the upper plate. This suggests that megathrust earthquakes affect the mechanical state of the overlying wedge and are expected to leave a widespread signature in the seismic record of rocks.

Here we present a detailed field study that documents the long-term conditions of brittle rock failure in a domain of a fossil accretionary complex that evolved from an outer into an inner wedge, overlying the aseismic and seismogenic part of the basal detachment, respectively. Our combination of structural information with isotope data (Sr, C and O) and predictions of a Mohr–Coulomb wedge model enables us to better understand how accretionary wedges respond to megathrust earthquakes. We conclude that coseismically induced stress changes can trigger wedge internal deformation and leave a discernable signature in the seismic record of rocks.

Results

Geological background. The study area is located in the Palaeogene accretionary complex of the central European Alps, which is one of the best documented fossil systems worldwide. The high relief of about 2,600 m provides unique insights into the internal architecture of this orogenic wedge. The area comprises an exhumed about 300-m-thick rock unit (Globotruncana marl) to minimize potential variations in the isotopic composition of the host rock. During the transport of sediments towards greater depths and associated diagenetic and low-grade metamorphic reactions, the Sr isotopic signature of the pore fluid will evolve towards that of the host rock. The Sr/Sr ratios of mineral veins formed at different depths will therefore record the evolution of the pore fluid.

Temperature and depth of vein formation. We studied the strontium (Sr/Sr), carbon (δ13C, % VPDB) and oxygen (δ18O, % VSMOW) isotope systematics of mineral veins and host rock to further constrain the setting and relative timing of vein formation during the prograde evolution of the Alpine wedge (Fig. 3; Supplementary Tables 1 and 2). Our approach is based on the following two assumptions: (1) during sedimentation the pore fluid is derived from sea water, which is not in isotopic equilibrium with the host rock. During the transport of sediments towards greater depths and associated diagenetic and low-grade metamorphic reactions, the Sr isotopic signature of the pore fluid will evolve towards that of the host rock. The Sr/Sr ratios of mineral veins formed at different depths will therefore record the evolution of the pore fluid. (2) C and O isotopes bear information on carbonate diagenesis, fluid sources, open- vs closed-system behaviour and temperature-controlled isotope fractionation. All samples were taken from a homogenous, 300-m-thick rock unit (Globotruncana marl) to minimize potential variations in the isotopic composition of the host rock. The unit was deposited in a calcite sea (that is, marine conditions at ~320–350 °C and ~0.3 GPa in the southern part of the study area) (Supplementary Fig. 1).

Field observations. Deformation related to accretion involved sediment compaction, folding, boudinage, dissolution-precipitation creep and faulting. Faults and fractures form mineral veins, indicating the presence of fluids during fracturing. The veins are present in all three thrust slices of the IFU and can be traced on the metre to decametre scale. Cross-cutting relationships indicate a successive formation of three groups of mineral veins (Fig. 2), which we used to decipher the history of brittle rock failure within the accretionary complex. The first group (G1) consists of calcite shear veins that are widespread in fine-grained lithologies, such as slates, marlstones and siltstones, but are absent in coarse-grained sandstones. The majority of G1 veins were formed incrementally along bedding planes during reverse faulting, documenting horizontal contraction within the wedge (Fig. 2a; Supplementary Fig. 2). Conjugate shear fractures are sometimes developed within marlstones at an angle of ~60° ± 10° to bedding-parallel shear veins. Oblique shear veins are intensively affected by sediment compaction and shortened by ~30–60% (Fig. 2d), suggesting a formation at a shallow level within the wedge. G1 veins and bedding planes were folded, which is most evident where interlayer folding occurs (Fig. 2e). Veins of the second and third group (G2 and G3) cross-cut G1 veins, folds and cleavage (Fig. 2f,g) and consist mainly of quartz and calcite with minor amounts of sulfides, plagioclase and white mica. They are less common than G1 veins and occur commonly in clusters, predominantly in the southern domain of the study area (Supplementary Fig. 1). G2 veins represent mineralized fault cores of steep normal faults and include centimetre-sized clasts of brecciated host rock (Fig. 2b). Similar to G1 veins, they are well developed within fine-grained lithologies. In contrast, G3 veins formed by steep-dipping extension fractures (Fig. 2c) are present in all lithologies. Together with G1 veins they record extensional brittle deformation within the wedge.

G1 veins. Calcite of G1 veins show Sr/Sr ratios of 0.7075–0.7079, which are similar to sea water at the time of sediment
To bulk carbonate leachate obtained from a limestone bed and to the pore fluid during calcite recrystallization. The high Sr content of the calcite will effectively buffer the isotopic signature of contemporaneous sea water. Such a Sr isotopic signature in our data set, we exclude any substantial post-diagenetic calcite recrystallization (for example, during low-grade metamorphism) and conclude that the formation of G1 veins took place over a temperature range of 40–70 °C. These values agree with previous temperature estimates for carbonate diagenesis obtained from deep-sea drilling projects and a recent clumped-isotope study on cementation and matrix recrystallization, which yielded temperatures of ~14–65 °C for calcite diagenesis. Given the pristine and marine-equilibrated isotopic signature in our data set, we exclude any substantial post-diagenetic calcite recrystallization (for example, during low-grade metamorphism) and conclude that the formation of G1 veins took place over a temperature range of 40–70 °C. Assuming a geothermal gradient of 20–30 °C/km, this corresponds to ~1–4 km depth and suggests a formation in the shallow outer wedge.
Figure 2 | Examples of mineral veins and cross-cutting relationships. (a) G$_1$ veins (red arrows) were formed by bedding-parallel slip and record contraction within the wedge. (b) G$_2$ veins represent mineralized cores of steep normal faults, contain clasts of brecciated host rock and are surrounded by a damage zone. (c) G$_3$ veins comprise of mineralized extension fractures. Both, G$_2$- and G$_3$ veins record pulses of extensional brittle deformation within the wedge. The orientation of the principal stresses $\sigma_1$ and $\sigma_3$ during vein formation is depicted in the upper left corner of the sketches. (d) Bedding-parallel G$_1$ vein with conjugate shear fracture. The conjugate shear fracture was intensively shortened during sediment compaction. (e) G$_1$ vein folded during interlayer folding. (f) Two G$_3$ veins cross-cutting G$_1$ veins. The G$_3$ vein on the right is offset along a minor, not mineralized normal fault. (g) G$_3$ veins cross-cutting cleavage. Scale bar, 25 cm (a–c,e,f). Scale bar, 5 cm (d,g).
**Analyses of whole-rock samples and leached silicate fractions of the Globotruncana marl yield relatively high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7088–0.7097 and 0.7174–0.7207, respectively (Supplementary Table 3). The pore fluid is therefore expected to evolve towards higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios during progressive burial. In fact, carbonates of G2- and G3 veins show $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.7080–0.7089, which is distinctly higher than the $^{87}\text{Sr}/^{86}\text{Sr}$ of G1 veins (0.7077 ± 0.0002). Similar to G1 veins, the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of calcite from G2- and G3 veins lie between 22.2 and 23.2‰ and 1.5 and 2.2‰, respectively, indicating that also during later stages of vein formation the C and O-isotopic composition of the pore fluid was buffered by the host rock (Supplementary Fig. 3). This enables us to constrain the formation temperatures of G2- and G3 veins based on the oxygen isotope fractionation between quartz and calcite, which is buffered by the host rock (Supplementary Table 4). The pore fluid is therefore expected to evolve towards higher $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values during the prograde evolution of the wedge.**

**Dynamic Mohr–Coulomb wedge.** To investigate how the two modes of brittle deformation are related to the large-scale tectonic setting at convergent margins, we apply the dynamic Mohr–Coulomb wedge theory33 to the Alpine prism. This theory expands the original Coulomb wedge model by considering stress changes during the subduction earthquake cycle, and predicts both the orientation of the maximum principal stress ($\sigma_1$) and the stress conditions under which the wedge fails. This enables us to constrain the timing of vein formation within the subduction earthquake cycle. The model is constrained by the wedge geometry, the strength of the wedge and the basal detachment, as well as the pore fluid pressure ratio $\lambda$ (where $\lambda$ is the ratio of pore fluid pressure and lithostatic pressure). The parameters were estimated from previous wedge analyses33,34 and balanced cross-sections35 and are given in Fig. 4. We further explored the parameter space within the range of reasonable values (Supplementary Fig. 4). The model wedge is divided into an outer and an inner wedge (Fig. 4a; Supplementary Table 4). The inner wedge overlaps the unstable, seismogenic zone of the basal detachment where earthquake nucleation occurs. This part of the plate boundary is characterized by velocity-weakening behaviour, which is modelled by a coseismic decrease in basal friction ($\mu_b$). The outer wedge overlaps the (conditionally) stable zone of the basal detachment, which shows velocity-strengthening behaviour and is modelled by a coseismic increase in $\mu_b$. The division of the basal detachment into a velocity-strengthening and a velocity-weakening part assumes a common behaviour for the subduction fault. Exceptional situations, in which the earthquake rupture propagates through the upper seismogenic zone, suggesting that the shallow part of the detachment experience only little strengthening or even weakening36, are not further considered here.

**Figure 3 | Sr and O isotope systematics.** The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios ($n = 41$) and $\Delta^{18}\text{O}_{\text{Qz–Cc}}$ values ($n = 14$) of mineral veins record the development of the pore fluid during the prograde evolution of the wedge. (a) Calcite veins (G1, blue) show $^{87}\text{Sr}/^{86}\text{Sr}$ ratios similar to sea water and limestone matrix and document reverse faulting within shallower levels of the wedge during carbonate diagenesis (~40–70 °C). Quartz-calcite veins (G2, red and G3, purple) show higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, which develop towards the Sr signature of the host rock (marlstone). The veins indicate extensional faulting and fracturing within deeper levels of the wedge. $\pm$ uncertainties are smaller than the symbols. (b) The oxygen isotope fractionation data between quartz and calcite ($\Delta^{18}\text{O}_{\text{Qz–Cc}}$) from G2- and G3 veins show a good correlation (R = 0.82) with the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. The respective temperatures suggest a formation of G2- and G3 veins over similar time intervals during the prograde evolution of the wedge between ~210 and 290 °C.
fluid pressure within the wedge to near-lithostatic values ($\lambda > 0.92$ and $> 0.96$ for the outer and inner wedge, respectively). For the outer wedge, we consider this as unlikely, as such high fluid pressures are not supported by seismological and borehole data\textsuperscript{37,38}. In the inner wedge, near-lithostatic fluid pressures persisting during the interseismic period would result in the formation of subhorizontal hydrofractures, which we do not observe in the field. Strong earthquakes, however, can induce stress variations, which will affect the mechanical state of the entire wedge. Coseismic strengthening of the basal detachment leads to a compression of the outer wedge, pushing it towards the compressively critical state. In this process, $\sigma_1$ rotates slightly towards the horizontal (Fig. 4b). The outer wedge fails when the basal friction $\mu'_b$ increases to $\sim 0.07$, which results in the formation of thrust faults (Fig. 4c). Taking a coseismic increase in fluid pressure into account\textsuperscript{33} (for example, from $\lambda = 0.8$ to $\lambda = 0.85$), rock failure occurs at an even lower increase of the basal friction. The inner wedge behaves in a converse manner to the outer wedge. Coseismic weakening of the basal detachment brings the wedge towards an extensionally critical state, and $\sigma_1$ rotates towards the vertical (Fig. 4b). Wedge instability occurs at fluid pressures of $\lambda \sim 0.90–0.95$ and associated drops in $\mu'_b$ of 0.02–0.03 (Fig. 4d). Depending on the cohesional strength of the rocks, the inner wedge fails either by normal faulting (for example, along weak planes such as cleavage) or extensional fracturing (Fig. 4d; Supplementary Fig. 5). In summary, our wedge analysis indicates that instabilities occur during transient stress changes, which are induced by megathrust earthquakes and persist over the co- to early postseismic periods (see also ref. 33). Moreover, the predicted stress fields and fracture patterns for the critical outer and inner wedge match remarkably well with the geometries and fracture modes inferred from the mineral veins in our study area (Figs 2 and 4; Supplementary Fig. 5).
Figure 5 | Different stages of vein formation along a particle path. (1) New sediments enter the subduction zone and are transported towards greater depth. (2) Coseismic compression of the outer wedge results in the formation of G1 veins. (3) The sediments are compacted and folded, pore water is expelled and remaining pore fluids are mixed with diagenetic-metamorphic fluids. (4) Coseismic extension within the inner wedge leads to the formation of G2- and G3 veins.

Discussion
We have investigated the formation of mineral veins during the structural evolution of an accretionary complex, covering the range from initial frontal accretion to deep burial and low-grade metamorphism of the rocks (Fig. 5). By combining field data with geochemical and mechanical constraints, we learn that vein formation in the Alpine prism was restricted to distinct structural positions within the wedge. Vein formation in the outer wedge coincided with carbonate diagenesis at 1–4 km depth and at temperatures of ~40–70 °C, that is, the same interval at which sediments experience the highest consolidation rates39. The combined effects of compaction and calcite cementation are likely to lead to a pronounced decrease in porosity, which promotes elevated pore fluid pressures in the accreted units. We suppose that, under these conditions the overpressured and mechanically weak rocks fail during coseismic compression of the outer wedge, resulting in the formation of G1 veins, as predicted by the Mohr–Coulomb wedge analysis.

In contrast, the formation of G2- and G3 veins took place in the inner wedge at elevated temperatures of ~210–290 °C. Interestingly, the smectite–illite transformation and associated K-feldspar consumption40, which commonly occur at temperatures of 60–150 °C (ref. 41), must have happened before the formation of these veins. We suppose that the release of radiogenic 87Sr from clays and feldspars caused a first increase in the 87Sr/86Sr signature of the pore fluid, as recorded in the mineral veins. Moreover, the release of silica and subsequent quartz cementation is thought to facilitate the change from a velocity-strengthening to a velocity-weakening behaviour of the rocks, which is necessary for the onset of seismogenesis42. A striking feature of G2- and G3 veins is that they record stages of extensional faulting in an overall convergent setting, which has also been documented for other active and fossil accretionary complexes. For example, submarine structural mapping of the Nankai accretionary prism revealed the formation of normal faults along the crest of anticlines in the outer wedge43. Stratal extension and normal faulting have also been found along thrust faults in the Makran accretionary prism44. The occurrence of extension fractures in accretionary mélanges further suggests the persistence of tensile stress conditions during tectonic mélangé formation45,46. Tension gashes in the palaeo-accretionary wedge of the South-Central Chilean forearc indicate extensional fracturing during exhumation of previously underplated rock units47. The latter case indicates that accretionary prisms may undergo protracted stages of regional extension, which could have produced the observed superposition of G2- and G3 veins in the Alpine wedge, albeit without the envisioned dynamic effects outlined above. However, the extensional structures described in our study are neither associated with other large-scale tectonic structures indicative of an extensional stage within the Alpine wedge, nor are they related to specific deformation processes such as folding, thrusting or mélangé formation. Moreover, the G2- and G3 veins developed on the prograde path and record a prominent switch from reverse faulting at shallow depths to extensional faulting at greater depths and temperatures. We therefore suggest that vein formation in the Alpine wedge illustrates the response of the wedge to megathrust earthquakes on the plate interface, which repeatedly shifted the wedge into a critical state. In this model, bedded-parallel shear veins (G1) originate from coseismic compression of the outer wedge,
whereas normal faults and steep extension fractures (G2 and G3) were formed during short periods of extension within the inner wedge (Figs 4 and 5). For the latter, the strength of the basal detachment has to decrease coseismically from 0.04 to 0.01–0.02. This decrease is equivalent to stress drops of 50–75% and may suggest the need of great megathrust earthquakes. Average stress drops, however, have been found to be constant and independent of earthquake magnitude and commonly range between 1 and 10 MPa, with a mean value of around 3 MPa. The shear stress on the megathrust is given by the friction law $t_s = \mu_b n$, where $\mu_b$ is the normal stress and approximately the weight of the overburden. At 11 km depth (that is, the average depth for the formation of G2- and G3 veins) $\sigma_b$ is about 270 MPa and $t_s \sim 11$ MPa during the interseismic period. A coseismic decrease of $\mu_b$ to 0.01–0.02 corresponds to stress drops of 5–8 MPa, indicating that the apparently high stress drops fall within the range of average values. Moreover, stress drops can vary significantly along the fault surface. While some patches experience complete stress drops others may even undergo stress increases.

We conclude that faulting and vein formation in accretionary complexes can be triggered by megathrust earthquakes, which shift the wedge into an unstable state. This implies a coseismic increase in fracture permeability within the hanging wall of megathrusts, followed by a stable phase of interseismic fault sealing. Our work helps to understand how fractures are generated throughout the subduction earthquake cycle, which is essential to better-constrain the nature of postseismic fluid flow and to assess the seismic hazard of hydraulically driven aftershocks.

Methods

Field investigations. The study area is located in the UNESCO World Heritage Swiss Tectonic Arena Sardona, in the northeastern part of the central European Alps (Supplementary Fig. 1). Structural elements (faults, mineral veins, bedding, cleavage and folds) were investigated along a 30–km-long northwest–southwest transect, comprising all three IFUs (that is, the central to southern part of the cross-section in Supplementary Fig. 1b). To approximate the orientation of structures at the time of formation, all structural measurements were corrected for exhumation-related block rotation, as constrained by apatite fission-track data. The trend of the rotation axis is 70° (orogen-parallel); the magnitude of rotation is 10° (Supplementary Fig. 1c). We note that this correction does not affect the results or interpretations presented in this study. All samples for geochemical and isotope analyses were taken from two areas within the Globotornacrina marl in the southern part of the study (marked by A1 and A2 in Supplementary Fig. 1a).

Geothermal gradient. The modern average surface heat flow Qs in the study area is ~80 mW m$^{-2}$ (ref. 54). For the upper continental crust the temperature at a given depth can be calculated by (for example, see ref. 55)

$$T(z) = T_s + Q_b / k - A_{\text{rad}} c^2 / 2 k$$

(1)

where $T_s$ is the surface temperature (273 K), $k$ the thermal conductivity (2.5 W m$^{-1}$ K$^{-1}$) and $A_{\text{rad}}$ the radiogenic heat production rate within the upper crust. Following Pollack and Chapman, we assume that $A_{\text{rad}}$ in the upper crust accounts for 40% of the surface heat flow, that is, $A_{\text{rad}} = 0.4Q_b / D_w$, where $D_w$ is the thickness of the upper crust (taken to be 20 km). Solving equation (1) gives an average geothermal gradient of ~27 °C km$^{-1}$ for the upper 15 km of continental crust. As the surface heat flow may have been lower before the onset of continental collision and associated thickening of the continental crust, we assume a conservative range for the geothermal gradient in our study area of 20–30 °C km$^{-1}$.

Dynamic Mohr–Coulomb wedge analysis. The dynamic Mohr–Coulomb wedge theory provides exact stress solutions for an elastic-perfectly Coulomb plastic rheology and considers temporal variations of stresses within the wedge and along the basal detachment in subduction earthquake cycles. We will briefly summarize the general concept of the theory and explain the main parameters constraining the model and refer the reader to refs 35,49,57 for further reading. The Coulomb wedge theory postulates that, the geometry of a wedge, which is given by the upper slope angle $\beta$ and the basal dip $\delta$, is related to the strength of the wedge and the strength of the basal detachment. The strength of the wedge is defined by the coefficient of internal friction $\mu_b$, the cohesion gradient $q$ and the pore fluid pressure ratio $\lambda$. The strength of the basal detachment is given by the effective coefficient of basal friction $\mu_b' = \mu_b (1 - \lambda_b)$, where $\mu_b$ and $\lambda_b$ denote the coefficient of basal friction and the basal pore fluid pressure ratio, respectively. The model is divided into an outer wedge (for the outer wedge, to account for differences in wedge strength and in seismogenic behaviour of the basal detachment. The outer wedge represents the near-trench part of accretionary wedges, which overlies the updip velocity-strengthening (aseismic) segment of the basal detachment and comprises young and weak sediments. The inner wedge overlies the velocity-weakening (seismic) part of the detachment and consists of well-consolidated older accreted units. The subduction earthquake cycle is modelled by varying the effective coefficient of basal friction $\mu_b'$. For the interseismic period $\mu_b'$ is set to a fixed reference value, which is equal for both parts of the basal detachment. Megathrust earthquakes are considered by increasing $\mu_b'$ during the seismic and seismogenic part of the detachment fault, respectively. Given the strength and geometry of the wedge, the model calculates the effective stress ratio $m$ for every value of $\mu_b'$ determining whether the wedge is stable (elastic deformation) or critical (at Coulomb failure). The results are pictured in a stability field diagram (Fig. 4, Supplementary Fig. 4), illustrating the critical values of basal friction $\mu_b'$ as a function of pore fluid pressure ratio $\lambda$. In addition, the model computes the maximum and minimum principal stresses ($\sigma_1$ and $\sigma_3$), as well as the angle of $\sigma_1$ with the wedge surface $\gamma$ as a function of basal friction $\mu_b'$ (Supplementary Fig. 5b).

All model parameters for the interseismic reference state are listed in Supplementary Table 4. The geometry of the Alpine accretionary wedge was constrained from geological cross-sections, which are based on surface information, borehole data and p-wave tomography models of the deep structure of the European Alps. The errors on $\lambda$ and $\beta$ due to uncertainties in the construction of cross-sections and velocity to depth conversion are estimated to be 1° and 0.01, respectively. $\mu_b'$ for the interseismic period was set to 0.04, according to values reported for the basal detachment in the central European Alps, as well as for the Nankai and Cascadia subduction zones. Following the above discussion, we defined the outer wedge to be weaker ($\mu_b' = 0.4$, $\lambda = 0.3$) than the inner wedge ($\mu_b' = 0.7$, $\lambda = 0.6$). The reference pore fluid pressure ratios for the outer and inner wedges were fixed at $\lambda = 0.8$ and $\lambda = 0.9$, respectively (Supplementary Table 4).

We verified the validity of our results by testing the effect of different model parameters (Supplementary Fig. 4). For example, the transition from the outer to the inner wedge is often associated with a break in slope, which cannot be assessed for exhumed palaeo-accretionary wedges. To take this into account, we examined the effects of a steeper outer and a flatter inner wedge and found no restrictions to the principle findings presented in this study (Supplementary Fig. 4a,d). The same applies to the other parameters (Supplementary Fig. 4). Finally, we constrained the precise conditions for normal faulting and extensional fracturing within the inner wedge (Supplementary Fig. 5). We calculated the interseismic part of the deformation and its orientation with a CO2-laser in 50 mbar of pure F2. Liberated O2 was purified through an extraction line and the gas passing over heated KCl salt (150 °C). The extracted O2 was collected on a molecular sieve (0.5 nm) and subsequently expanded into the mass spectrometer. Replicate oxygen isotope analyses of the standards used during the runs ($n = 19$) yielded an external reproducibility of ≤ 0.2%.

Stable isotopes and temperature estimates. $^{13}$C values of quartz from G2- and G3 veins were obtained using a laser fluorination method and a Finnigan MAT 253 isotope ratio mass spectrometer at the University of Lausanne. Quartz samples were decarbonated in 1 M HCl for 4h. Between 0.5 and 2 mg of sample material was loaded onto a Pt-sample holder and pumped out to a vacuum of about 10$^{-6}$ mbar. After preflourination of the sample chamber, the samples were heated with a CO2 laser in 50 mbar of pure F2. Liberated O2 was purified through an extraction line and the gas passing over heated KCl salt (150 °C). The extracted O2 was collected on a molecular sieve (0.5 nm) and subsequently expanded into the mass spectrometer. Replicate oxygen isotope analyses of the standards used during the runs ($n = 19$) yielded an external reproducibility of ≤ 0.2%.

Stable isotopes and temperature estimates. $^{18}$O values of quartz from G2- and G3 veins were obtained using a laser fluorination method and a Finnigan MAT 253 isotope ratio mass spectrometer at the University of Bern with a Gasb dichromat connected to a Finnigan MAT Delta Plus XL isotope mass spectrometer, using a He-carrier gas system. About 250–1,000 µg of sample material was dissolved in concentrated H3PO4. All results are normalized using an in-house standard calibrated against $^{18}$O and $^{13}$C values of NBS-19 and NBS-28, respectively. The external reproducibility of replicate analyses estimated from replicate analyses of the in-house standard ($n = 16$) is ± 0.05% for $^{13}$C and ± 0.04% for $^{18}$O.
The temperature-dependent O isotope fractionation between quartz and calcite was calculated using empirically derived fractionation equations for O₂-H₂O and \( \text{CO}_2-\text{H}_2\text{O} \), which were determined for a temperature range of 200–500 °C and 0–500 °C, respectively.31,32 Kim and O’Neill4 reported small differences in the oxygen isotope fractionations between calcite and water at temperatures <25 °C, which do not affect our calculations and are not further considered here. The 1σ error of the vein formation temperature was calculated by assuming a total 1σ error of ±0.24 in the δD measurements, that is, the sum of the external reproducibilities (±0.20% for δD_H2O) and ±0.04% for δ18O_C.

Radioactive isotopes. Rb-Sr isotopes were obtained for three samples of Globotrunca marl, comprising two samples with 50–60% calcite and one sample of shale containing no calcite. Aliquots of the two calcite-rich samples were treated with 1 M HCl for 5 minutes and subsequently centrifuged for 10 minutes at 3,000 rpm. The solutions were analyzed for the Rb-Sr isotope system.

References
1. Barr, T. D., Dahlen, F. A. & McPhail, D. C. Brittle frictional mountain building 3. Lower-grade metamorphism. J. Geophys. Res. 96, 319–30,338 (1991).
2. Fischer, D. & Byrne, T. Structural evolution of underthrustsediments, Kodiak Islands, Alaska. Tectonics 6, 775–793 (1987).
3. Moore, J. C. & Saffer, D. Updip limit of the seismogenic zone beneath the accretionary prism of southwest Japan: an effect of diagenetic low-grade metamorphic processes and increasing effective stress. Geology 29, 183–186 (2001).
4. Tobin, H. J. & Saffer, D. M. Elevated fluid pressure and extreme mechanical weakening of a plate boundary thrust, Nankai Trough subduction zone. Geology 37, 679–682 (2009).
5. Sample, J. C. Stable isotope constraints on vein formation and fluid evolution along a recent thrust fault in the Cascadia accretionary wedge. Earth Planet. Sci. Lett. 293, 300–312 (2010).
6. Rowe, C., Moore, J. C., Meneghini, F. & McKeiman, A. W. Large-scale Precambrian tectonophases and fluidized cataclasites from an ancient subduction thrust fault. Geology 33, 937–940 (2005).
7. Vannucci, P., Remitti, F. & Bettele, G. Geological record of fluid flow and serpentinization along an erosive subducting plate boundary. Nature 451, 699–703 (2008).
8. Bachmann, R. et al. Exposed plate interface in the European Alps reveals fabric styles and gradients related to an ancient seismogenic coupling zone. J. Geophys. Res. 114, B05002 (2009).
9. Fagereng, Å. & Sàbin, R. H. Melange rheology and seismic style. Geology 38, 753–754 (2010).
10. Aron, F., Allmendinger, R. W., Cembrano, J., González, G. & Yáñez, G. Permanent fore-arc extension and seismic segmentation: insights from the 2010 Maule earthquake, Chile. J. Geophys. Res. 118, 724–739 (2013).
11. Lien, K. et al. Splay fault activity revealed by aftershocks of the 2010 Mw 8.8 Maule earthquake, central Chile. Geology 42, 823–826 (2014).
12. Kato, A. et al. Imaging the source regions of normal faulting sequences induced by the 2011 M9.0 Tohoku-Oki earthquake. Geophys. Res. Lett. 40, 273–278 (2013).
13. Piiffner, O. A. Evolution of the north Alpine foreland basin in the central Alps. Spec. Publs. Int. Ass. Sediment. 8, 219–228 (1986).
14. Lihou, J. C. & Saffer, D. M. Elevated fluid pressure and extreme mechanical weakening of the seismogenic zone beneath the Maule earthquake, central Chile. J. Geophys. Res. 118, 724–739 (2013).
15. Sinclair, H. D. Tectonostratigraphic model for underfilled peripheral foreland basins: An Alpine perspective. Geol. Soc. Am. Bull. 699–703 (2008).
16. Burkhard, M. A. A. X-ray study of Miocene deformation in the most external part of the Swiss Alps (Subalpine Molasse to Jura fold belt). Ecolge Geol. Helv. 83, 559–582 (1990).
17. Glotzbach, C. et al. Thermal history of the central Gotthard and Aar massifs, European Alps: evidence for steady state, long-term exhumation. J. Geophys. Res. 115, F03017 (2010).
18. Rahn, M. K., Hurford, A. J. & Frey, M. Rotation and exhumation of a thrust plane:apatite fission-track data from the Glarus thrust, Switzerland. Geology 25, 599–602 (1997).
19. Lihou, J. C. Structure and deformational history of the Infralithospheric flysch units, Glarus Alps, eastern Switzerland. Ecolge Geol. Helv. 89, 439–460 (1996).
20. Milliman, J. D. Recent sedimentary carbonates: Part 1 Marine carbonates (Springer, 1974).
21. Veizer, J. et al. 87Sr/86Sr, δ18O and δ34S evolution of Phanerozoic seawater. Chem. Geol. 161, 59–88 (1999).
22. Hudson, J. D. Stable isotopes and limestone lithification. J. Geol. Soc. Lond. 133, 637–660 (1977).
23. Machel, H. G., Cavell, P. A. & Patay, K. S. Isotopic evidence for carbonate cementation and recrystallization, and for tectonic expulsion of fluids into the Western Canada Sedimentary Basin. Geol. Soc. Am. Bull. 108, 1108–1119 (1996).
24. Voigt, J. et al. Variability of carbonate diagenesis in equatorial Pacific sediments deduced from radiogenic and stable Sr isotopes. Geochim. Cosmochim. Acta 148, 360–377 (2015).
25. Milliman, J. D. Recent Sedimentary Carbonates: Part 1 Marine Carbonates (Springer, 1974).
26. Veizer, J. et al. 87Sr/86Sr, δ18O and δ34S evolution of Phanerozoic seawater. Chem. Geol. 161, 59–88 (1999).
27. Hudson, J. D. Stable isotopes and limestone lithification. J. Geol. Soc. Lond. 133, 637–660 (1977).
28. Machel, H. G., Cavell, P. A. & Patay, K. S. Isotopic evidence for carbonate cementation and recrystallization, and for tectonic expulsion of fluids into the Western Canada Sedimentary Basin. Geol. Soc. Am. Bull. 108, 1108–1119 (1996).
29. Sample, J. C. The effect of carbonate cementation of underthrust sediments on deformation styles under upwelling. J. Geophys. Res. 95, 9111–9121 (1990).
30. Huntington, K. W. et al. Use of clumped-isotope thermometry to constrain the crystallization temperature of diagenetic calcite. J. Sediment. Res. 81, 656–669 (2011).
31. O’Neill, J. R., Clayton, R. N. & Mayeda, T. K. Oxygen isotope fractionation in diatomaceous eolian carbonates. J. Chem. Phys. 51, 5547–5558 (1969).
32. Clayton, R. N., O’Neill, J. R. & Mayeda, T. K. Oxygen isotope exchange between quartz and water. J. Geophys. Res. 77, 3057–3067 (1972).
33. Wang, K. & Hu, Y. Accretionary prisms in subduction earthquake cycles: the theory of dynamic Coulomb wedge. J. Geophys. Res. 111, B04100 (2006).
34. von Hagen, C., Ondre, O. & Evseev, S. Critical taper analysis reveals lithological control of variations in detachment strength: an analysis of the Alpine basal detachment (Swiss Alps). Geochem. Geophys. Geosyst. 15, 176–191 (2014).
35. Rosenberg, C. L. & Kissling, E. Three-dimensional insight into Central-Alpine collision: lower plate or upper plate indentation? Geology 39, 121–1222 (2013).
36. Kameyama, G. et al. Runaway slip to the trench due to the rupture of highly pressurized megathrust beneath the middle trench slope: the tsunamigenesis of the 2011Tohoku earthquake off the east coast of northern Japan. Earth Planet. Sci. Lett. 339–340, 32–42 (2012).
37. Seno, T. Determination of the pore fluid pressure ratio at seismogenic megathrusts in subduction zones: implications for strength of asperities and Andean-type mountain building. J. Geophys. Res. 114, B05050 (2009).
38. Becke, K., Fisher, A. T. & Davis, E. E. The cork experiment in hole 949C: long-term observations of dewatering processes. Proc. Ocean Drill. Program Sci. Results 156, 247–252 (1997).
39. Bray, J. C. & Karig, D. E. Porosity of sediments in accretionary prisms and some implications for dewatering processes. J. Geophys. Res. 90, 768–778 (1985).
40. Awasser, D. N. Illite/smectite formation and potassium mass transfer during burial diagenesis of mudrocks: a study from the Texas gulf coast Paleocene-Eocene. J. Sediment. Petrol. 10, 501–512 (1993).
41. Pytte, A. M. & Reynolds, R. C. In: *Thermal History Of Sedimentary Basins*. (eds Naeser, N. D. & McCulloh, T. H.) 133–140 (Springer, 1988).
42. Moore, J. C., Rowe, C. & Meneghini, F. In: *The Seismogenic Zone of Subduction Thrust Faults*. (eds Dixon, T. H. & Moore, J. C.) 288–315 (Columbus Univ. Press, 2007).
43. Anma, R. *et al.* In: *Accretionary Prisms And Converging Margin Tectonics In The Northwest Pacific Basin*. (eds Ogawa, Y., Anma, R. & Dilek, Y.) 169–196 (Springer, 2011).
44. Platt, J. P. & Leggett, J. K. Stratal extension in thrust footwalls, Makran accretionary prism: implications for thrust tectonics. AAPG Bull. 70, 191–203 (1986).
45. Vrolijk, P. Tectonically driven fluid flow in the Kodiak accretionary complex, Alaska. Geology 15, 466–469 (1987).
46. Kimura, G. *et al.* Tectonic mélangé as fault rock of subduction plate boundary. Tectonophysics 568–569, 25–38 (2012).
47. Gladny, J. *et al.* Internal dynamics of a paleoaccretionary wedge: insights from combined isotope tectonochemistry and sandbox modelling of the South-Central Chilean forearc. Earth Planet. Sci. Lett. 231, 23–29 (2005).
48. Allmann, B. P. & Shearer, P. M. Global variations of stress drop for moderate to large earthquakes. J. Geophys. Res. 114, B01310 (2009).
49. Dahlen, F. A. Noncohesive critical Coulomb wedges: an exact solution. J. Geophys. Res. 89, 70125–10,133 (1984).
50. Gao, X. & Wang, K. Strength of stick-slip and creeping subduction megathrusts from heat flow observations. Science 345, 1038–1041 (2014).
51. Lutrell, K. M. *et al.* Estimates of stress drop and crustal tectonic stress from the 27 February 2010 Maule, Chile, earthquake: implications for fault strength. J. Geophys. Res. 116, B11401 (2011).
52. Husen, S. & Kissling, E. Postseismic fluid flow after the large subduction earthquake of Antofagasta, Chile. Geology 29, 847–850 (2001).
53. Miller, S. *et al.* Aftershocks driven by a high-pressure CO₂ source at depth. Nature 427, 724–727 (2004).
54. Maps of Switzerland - Swiss Confederation. http://map.geo.admin.ch (2015).
55. Chapman, D. S. Thermal gradients in the continental crust. Geol. Soc. Lond. Spec. Publ. 24, 63–79 (1986).
56. Pollack, H. N. & Chapman, D. S. On the regional variation of heat flow, geotherms, and lithospheric thickness. Tectonophysics 38, 279–296 (1977).
57. Hu, Y. & Wang, K. Coseismic strengthening of the shallow portion of the subduction fault and its effects on wedge taper. J. Geophys. Res. 113, B12411 (2008).
58. Wang, K. & He, J. Mechanics of low-stress forearc: Nankai and Cascadia. J. Geophys. Res. 104, 15,191–15,205 (1999).
59. Si bson, R. H. Brittle failure mode plots for compressional and extensional tectonic regimes. J. Struct. Geol. 20, 655–660 (1998).
60. Cox, S. F. The application of failure mode diagrams for exploring the roles of fluid pressure and stress states in controlling styles of fracture-controlled permeability enhancement in faults and shear zones. *Geofluids* 10, 217–233 (2010).
61. Rumble, D. III & Hoering, T. C. Analysis of oxygen and sulfur isotope ratios in oxide and sulfide minerals by spot heating with a carbon dioxide laser in a fluorine atmosphere. Acc. Chem. Res. 27, 237–241 (1994).
62. Kasemann, S. *et al.* Boron and oxygen isotope composition of certified reference materials NIST SRM 610/612, and reference materials JR-2G and JR-2G. Geostandards Newsletter 25, 405–416 (2001).
63. Spoel, C. & Vennemann, T. W. Continuous-flow IRMS analysis of carbonate minerals. Rapid Commun. Mass Spectrom. 17, 1004–1006 (2003).
64. Kim, S. T. & O’Neil, J. R. Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. Geochim. Cosmochim. Acta 61, 3461–3475 (1997).
65. Nier, A. O. The isotopic constitution of strontium, barium, bismuth, thallium and mercury. Phys. Rev. 54, 275–278 (1938).
66. Hunziker, J. C. *et al.* The evolution of illite to muscovite; mineralogical and isotopical data from the Glarus Alps, Switzerland. Contrib. Mineral. Petrool. 92, 157–180 (1986).
67. Neumann, W. & Huster, E. The half life of 87Rb, measured as difference between the isotopes 87Rb and 87Rb. Z. Phys. 270, 121–127 (1974).

Acknowledgements
Funding for this project was provided by the Swiss National Science Foundation (No. 144381). We thank R.M. Danisi, I. Villa, A. Hausermann, K. Merger, D. Culatello, G. Schreurs and S. Lowick for discussion and support.

Author contributions
M.H. and A.B. designed the study. H.V. conducted the Sr analysis. T.V. supervised the O analyses, executed the dynamic wedge calculations and wrote the manuscript. All authors discussed the data set.

Additional information
Supplementary Information accompanies this paper at http://www.nature.com/naturecommunications.

Competing financial interests: The authors declare no competing financial interests.

Reprints and permission information is available online at http://npg.nature.com/reprintsandpermissions/

How to cite this article: Dielforder, A. *et al.* Linking megathrust earthquakes to brittle deformation in a fossil accretionary complex. Nat. Commun. 6:7504 doi: 10.1038/ncomms8504 (2015).

This work is licensed under a Creative Commons Attribution 4.0 International License. The images or other third party material in this article are included in the article’s Creative Commons license, unless indicated otherwise in the credit line; if the material is not included under the Creative Commons license, users will need to obtain permission from the license holder to reproduce the material. To view a copy of this license, visit http://creativecommons.org/licenses/by/4.0/