



This article appeared in a journal published by Elsevier. The attached copy is furnished to the author for internal non-commercial research and education use, including for instruction at the authors institution and sharing with colleagues.

Other uses, including reproduction and distribution, or selling or licensing copies, or posting to personal, institutional or third party websites are prohibited.

In most cases authors are permitted to post their version of the article (e.g. in Word or Tex form) to their personal website or institutional repository. Authors requiring further information regarding Elsevier's archiving and manuscript policies are encouraged to visit:

<http://www.elsevier.com/copyright>



Relationships between the microstructural evolution and the rheology of talc at elevated pressures and temperatures

J. Escartín ^{a,*}, M. Andreani ^b, G. Hirth ^c, B. Evans ^d

^a CNRS, Laboratoire de Géosciences Marines (CNRS FRE2316), case 89, Institut de Physique du Globe de Paris, 4 Place Jussieu, 75252 Paris, France

^b Laboratoire de Géosciences, Université Montpellier 2, Case 49, Place, Eugène Bataillon, 34095 Montpellier Cedex 5, France

^c Brown University, Geological Sciences, 324 Brook Street, Box 1846, Providence, RI 02912, USA

^d Earth, Atmospheric and Planetary Sciences, Massachusetts Institute of Technology Cambridge, Massachusetts 02139-4307, USA

Received 11 October 2007; accepted 3 February 2008

Available online 16 February 2008

Editor: G.D. Price

Abstract

We conducted triaxial compression experiments to investigate the effects of pressure, temperature and strain rate on the rheology of talc rocks. The tests were carried out at T from 25 to 860 °C, P from 0.1 to 300 MPa, and $\dot{\epsilon}$ from 0.3 to $30 \times 10^{-5} \text{ s}^{-1}$. In addition to being very weak, there are other important differences between the mechanical behavior of talc and that of most other silicates. Deformation at all temperatures up to dehydration was accommodated by a combination of crystal plasticity, frictional sliding, and cataclasis. A transition from localized to distributed deformation occurred at $P=300$ MPa and $T=400$ °C, but this transition is ill-defined as both distributed and localized deformation were observed at $P=300$ MPa and $T=600$ °C. Unlike most silicates, both the coefficient of internal friction and the coefficient of (sliding) friction are very low and nearly equal to each other. Temperature enhances plastic deformation (kinking), and inter- and intra-granular microcracks are observed parallel to the (001) planes at all conditions. Voids are created by delamination along (001) planes at kink-band boundaries. Full crystal plasticity was not achieved under any of the conditions tested, i.e., the von Mises criterion was never satisfied. However, the strength of the aggregates remained much less than the confining pressure. These unusual properties are likely a manifestation of the pronounced mechanical anisotropy of talc at the grain scale. At $T \geq 750$ °C dehydration is observed, but only along shear zones. These results suggest that feedbacks between deformation and reaction kinetics could control fluid flow in fault zones. The presence of talc can promote significant weakening and strain localization in the oceanic lithosphere, the subducting plate and the overlying mantle wedge.

© 2008 Elsevier B.V. All rights reserved.

Keywords: rock mechanics; subduction; talc; rheology; dehydration

1. Introduction

Talc is a metamorphic alteration product commonly found along long-lived oceanic transform and detachment faults (MacLeod et al., 2002; Escartín et al., 2003; D'Orazio et al., 2004; Schroeder and John, 2004; Boschi et al., 2006). It has also

been identified on active slip surfaces of major faults (e.g., San Andreas (e.g., ICDP, 2005; Moore and Rymer, 2007) and Moresby faults (e.g., Floyd et al., 2001; Taylor and Huchon, 2002)). Talc is stable up to temperatures of 800 °C at pressures of 5 GPa and, consequently, is one of the last phyllosilicates to dehydrate during prograde metamorphism associated with plate subduction. Because it contains ~4.8 wt.% H_2O , talc could be a significant source of water percolating into the mantle wedge overlying the subducting slab (e.g., Peacock and Hyndman, 1999).

Previous experiments on intact cores (e.g., Edmond and Paterson, 1971a,b; Summers and Byerlee, 1977) demonstrate

* Corresponding author.

E-mail address: escartin@ipgp.jussieu.fr (J. Escartín).

¹ Temporarily at: Earth, Atmospheric & Planetary Sciences, Massachusetts Institute of Technology, 77 Mass. Av., Cambridge MA 02139, USA, & EPS, Harvard University, Cambridge, MA 02138, USA.

that talc is extremely weak at laboratory conditions, consistent with its well-known position on the Moh's hardness scale. Thus, if present along a fault, talc could play an important role in the rheology of oceanic rocks and in the dynamics of subduction. However, the grain-scale deformation mechanisms are not well-studied, and therefore an understanding of how to apply the experimental deformation results is somewhat limited. For example, the style of brittle deformation is important for constraining relationships between faulting and fluid transport. Friction experiments on talc gouge show that the coefficient of sliding friction (μ) decreases from 0.36–0.24 under dry

conditions, to about 0.2 under wet conditions (Morrow et al., 2000; Moore and Lockner, 2004). The low coefficient of friction is attributed to weak bonding along basal (001) planes (Morrow et al., 2000).

In this study, we report results of triaxial deformation experiments on intact cores of talc; the tests investigated the effects of confining pressure (P), temperature (T), and strain rate ($\dot{\epsilon}$) on rheology. Samples were subsequently examined using optical, scanning electron (SEM) and transmission electron microscopy (TEM) to characterize the micromechanisms of deformation and to compare them with those of other

Table 1
Summary of experimental conditions and results

Run# ¹	T (°C)	P_{eff}^2 (Mpa)	Strain rate ($\times 10^{-5} \text{ s}^{-1}$)	σ_{max}^3 (Mpa)	$\sigma_{1\%}^4$	Sliding mode ⁵	Stress evolution ⁶	Fault angle, β^7 °	$\Delta\epsilon_v^8$ (%)
TM-4.1	25	100	3	55		–	W	45	–0.10
TM-4.3 ⁹	25	50	3	13		–	W	Split	–
TM-3.1	25	100	3	58		–	W	51	–0.05
TM-3.2	25	74	3	61		–	W	50	0.15
TM-4.2	25	112	3	54		–	W	50	–
TM04	600	305.2	0.3, 1, 3, 1	90.5		U	W	45	–
TM05	400	292.4	3	–	71.7	S	H	D	–
TM06	400	290.0	3, 10, 30, 3	–	56.0	S	H	D	–
TM07	400	50.2	3, 9, 3	48.8	37.8		W	25	–
TM08a	400	49.8	3	33.5	22.3	S	W	43	–
TM08b	400	198.2	3	–	64.1	S	H	–	–
TM08c	400	294.8	3	–	80.6	S	H	–	–
TM09	400	49.7	3	26.5	19.6	S	W	45	–
TM10	400	295.7	6	–	49.8	S	H	D(42)	–
TM11 ¹⁰	400	51.0	3	–		–	–	–	–
TM20	600	299.2	3	–	55.6	S	H	D(43)	–
TM21 ¹⁰	400	99.3	3	41.1		S	W	45	–
TM23	600	101.0	3	57.1	45.3	U	W	47	–
TM24a	600	99.6	3	49.5	44.5	U	W	43	–
TM24b	600	199.7	3	–	61.7	U	H	–	–
TM24c	600	298.6	3	–	78.8	S	H	–	–
TM25	600	100.3	3, 30	42.1	–	U	W	40	–
TM26	500	99.6	3	46.4	33.1	U	W	49	–
TM27	750	101.1	3	56.2	50.3	U	H	D(45)	–
TM28	860	99.2	3	48.2	36.5	–	W	33	–
TM30	600	50.1	3	35.1	14.3	S	W	30	–
TM31	600	285.7	3	–	93.8	U	H	D	–
TM32	600	52	3, 9, 3	61.7	31	S	W	45	–
TM33	600	302	3, 9, 3, 1	–	72.9	S,U	H	D	–
TRA	727	100.1	3	50.1	22.8	S	W	45	–
TRB	600	100.0	3	57.7	34.8	S	W	45	–
TRD	600	301.8	7	–	86	S	H	D	–
TRF	875	100	3	30.2	13.8	S	W	32	–

¹Experiment run. Runs in italics were conducted on a stiff apparatus, and the rest on an internally heated 'Paterson Instruments' apparatus. P -stepping runs are indicated by the run number and a letter for each P step.

² P_{eff} : Effective confining pressure. $P_{\text{eff}} = P - P_f$. P_f is the pore fluid pressure. All runs had $P_f = 0$, except for TM09, run at $P_c = 107.3$ and $P_f = 59$ MPa with a sinusoidal ± 1 and ± 2 MPa variation for permeability measurement.

³ σ_{max} : For stress–strain curves showing weakening, this corresponds to the peak stress. Jacket correction using the flow law of copper (Frost and Ashby, 1982) is applied to reported values.

⁴ $\sigma_{1\%}$: For strain hardening curves, we report $\sigma_1 - \sigma_3$ at 1% non-elastic strain ϵ_n . ϵ_n is calculated from the measured strain ϵ , and the elastic strain ϵ_e , $\epsilon_n = \epsilon - \epsilon_e$; $\epsilon_e = \sigma_1 - \sigma_3 / E$, and E is the Young's Modulus. Jacket correction using the flow law of copper (Frost and Ashby, 1982) is applied to reported values.

⁵Sliding mode. S: stable; U: unstable.

⁶Mode. W: Strain weakening after σ_{max} and localization; H: Strain hardening.

⁷Fault angle: β corresponds to the angle between σ_1 and the fault plane as measured on polished cut surfaces. For samples showing distributed deformation, the angle of visible shear zones is given in parenthesis.

⁸ $\Delta\epsilon_v$: Total change in an elastic volumetric strain at σ_{max} after the onset of dilatancy or compaction.

⁹Run TM4.3 developed a pressure leak during the experiment.

¹⁰Runs stopped at low strain, after yield prior to peak stress (TM11) or immediately after it (TM20).

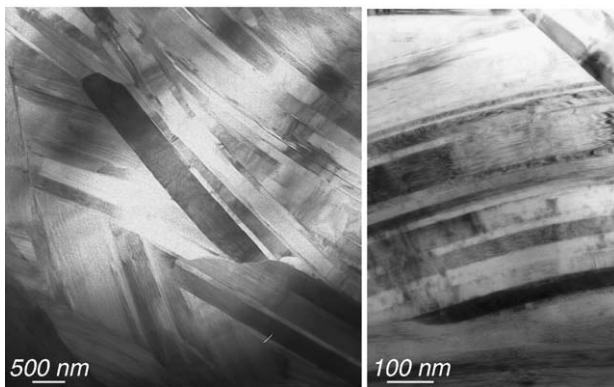


Fig. 1. Bright field TEM images of the undeformed TM starting material. The sample shows grains $<10\ \mu\text{m}$ in length and $<300\ \text{nm}$ in width and no apparent porosity.

phyllosilicates, such as serpentine (Morrow et al., 2000; Escartín et al., 1997a; 2001). We discuss the implications of these results for the mechanisms of strain localization along oceanic faults and for processes operating at subduction zones.

2. Experimental details

2.1. Starting materials

Most experiments were done on intact cylindrical samples cored from a nominally pure talc rock, provided by Mondo Minerals, which has a weak and variable schistosity. A few experiments were performed on talc rocks with no apparent schistosity, provided by E. Rutter (University of Manchester, England). These runs are named TM and TR, respectively in Table 1. In both rocks, talc grains range from 1–10 μm in length, and 50–300 nm in width (Fig. 1). When observed in TEM, TM samples exhibit domains of strong lattice preferred orientation (Fig. 1), but these domains are only continuous at scales of 1 to 2 μm . The density of the material is $\sim 2780 \pm 20\ \text{kg/m}^3$, at the higher end of the 2600–2800 kg/m^3 range reported for talc (Deer et al., 1962), and consistent with both the apparent lack of porosity (Fig. 1), and the presence of trace amounts of higher-density apatite (3160–3220 kg/m^3 (Deer et al., 1962)).

2.2. Experimental design and sample preparation

Deformation experiments were conducted from room temperature to 860 $^{\circ}\text{C}$, P from 50 to 300 MPa, and $\dot{\epsilon}$ from 0.3 to $30 \times 10^{-5}\ \text{s}^{-1}$ (Table 1; Figs. 2–4). Most experiments were conducted within the stability field of talc, i.e., $<700\ ^{\circ}\text{C}$ for the pressures we employed (Pawley, 1998; Pawley and Wood, 1995). Four samples were deformed outside the stability field at $P=100\ \text{MPa}$ (TM27, TM28, TRA and TRF, Table 1); these samples showed trace amounts of dehydration reaction products in TEM (Andreani et al., 2006).

Room temperature experiments were conducted on talc cores ($\sim 12.5\ \text{mm}$ in diameter and $\sim 25.5\ \text{mm}$ in length) using a stiff, gas-medium apparatus and jacketed in thin ($<0.1\ \text{mm}$) copper foil. Radial and axial strains were measured using strain gauges

glued onto the copper jackets (see Escartín et al. (1997a) for details; Fig. 5). Higher temperature experiments, using a servo-controlled, internally heated, gas-medium apparatus from Paterson Instruments (see Paterson (1990) and Xiao and Evans (2003)) were conducted on cores 10 mm in diameter and 20 mm in length, also jacketed in copper ($\sim 0.3\ \text{mm}$ thick). Argon was the confining medium in all experiments. Temperature was monitored 3 mm above the sample; furnace calibrations show that T varied by less than 3 $^{\circ}\text{C}$ along the sample length. In all experiments except TM09 (Table 1) solid spacers were used at the upper and lower ends of the samples. Split spacers were used in experiment TM09 so that we could measure permeability.

Raw data from the axial displacement and load transducers were converted to axial strain and differential stress. The permanent inelastic strain was calculated by subtracting the elastic strain, i.e., $\epsilon_e = (\sigma_1 - \sigma_3)/E$, from the total sample strain; E was calculated from the linear portion of the stress–strain curve. For experiments where strain-gauge data were available, a similar calculation was done to obtain the inelastic volumetric strain (ϵ_v). In calculating stress, we corrected the data in Table 1 for the strength of the copper jackets (Frost and Ashby, 1982), based on calibration experiments on jacketed aluminum samples. In calculating the jacket correction we assume

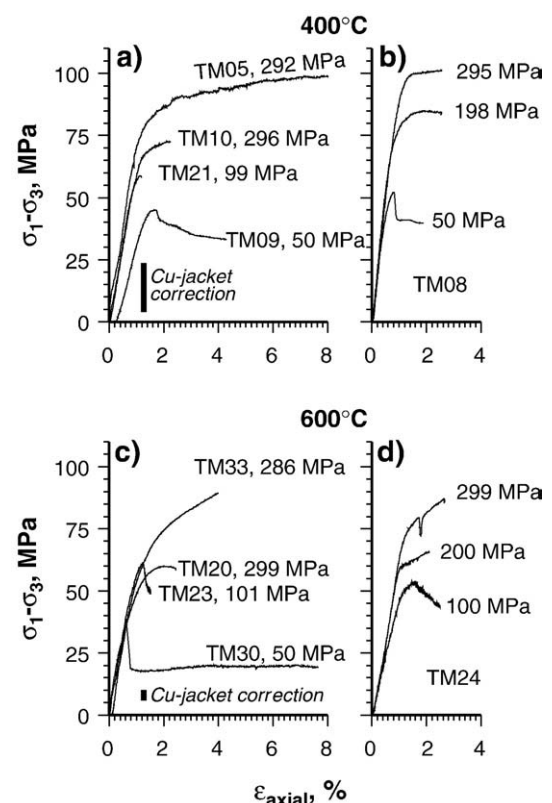


Fig. 2. Differential stress ($\sigma_1 - \sigma_3$) vs. axial strain (ϵ_a) curves of selected experiments at constant P (left), and with increasing P steps (right). All experiments were conducted at constant $\dot{\epsilon}$ (see Table 1). Note the variability of the differential stress in experiments conducted at $\sim 300\ \text{MPa}$ both at 400 $^{\circ}\text{C}$ (a–b) and 600 $^{\circ}\text{C}$ (c–d). The magnitude of the copper jacket correction for is shown for each T at $\dot{\epsilon} = 3 \times 10^{-5}\ \text{s}^{-1}$; no strength correction is actually applied to the curves shown in the figure.

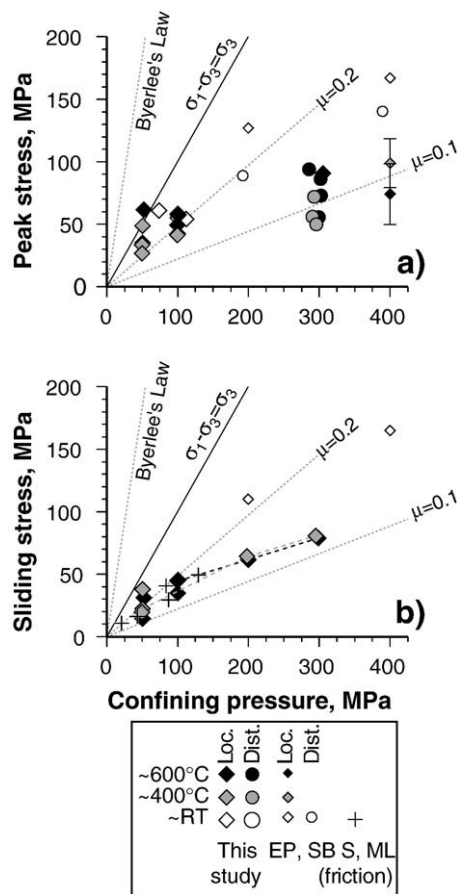


Fig. 3. Pressure dependence of the strength of intact cores (a) and of the sliding stress (b), at different T . Data are corrected for the strength of the copper jacket (Table 1). Samples that show localized deformation (Loc.) display strain weakening, while those that show distributed deformation (Dist.) display constant stress or strain hardening. Byerlee's Law, Goetze's criterion ($\sigma_1 - \sigma_3 = \sigma_3$) and lines for $\mu=0.1$ and 0.2 are plotted for reference. EP: Edmond and Paterson (1971a,b); SB: Summers and Byerlee (1977); S: Scruggs (1997); ML: Moore and Lockner (2004).

homogenous deformation of the jacket, thus the correction may be overestimated for samples that exhibit strain localization. We did not apply any geometric corrections to the load data, as might be based on assumptions of changes in cross-sectional area of the sample or in fault surface area.

After deformation, selected samples were impregnated with epoxy, cut perpendicular to the macroscopic shear zone, and made into double-sided, polished thin sections. Ion-milled samples were observed using a JEOL 2000fx TEM operating at 200 kV, at the CRMN (Marseille, France). SEM observations were performed on a LEICA S440 at LISE (CNRS/UPMC, Ivry, France).

3. Mechanical results

3.1. Stress–strain curves

Table 1 summarizes the conditions and results of 29 triaxial deformation experiments. At low confining pressures, samples display an upper yield point, i.e., a fracture strength, followed by an interval of strain at lower stresses (Fig. 2). As is common in other brittle rocks, strain weakening was associated with

localization of deformation, followed by subsequent sliding along a fault. We used the latter portion of the stress versus strain curve to characterize the sliding strength (see Section 4.1). With increasing pressure, the samples undergo a transition from strain weakening to strain hardening. Strain hardening was observed in all experiments conducted at 400 °C and $P \approx 300$ MPa. By contrast, at 600 °C we observed strain weakening at $P < 300$ MPa, but both hardening and weakening were observed at $P \approx 300$ MPa (compare TM33 and TM20 in Fig. 2). In all experiments, the slope of the elastic portion of the stress–strain curves is similar, with no resolvable dependence on P or T .

The stress–strain curves (Fig. 2a and c) illustrate the variability in mechanical behavior of the TM samples, which likely arises from heterogeneity in sample microstructure. To mitigate the influence of sample variability on the interpretation of mechanical behavior, we also conducted pressure-stepping experiments at 400 and 600 °C (TM08 and TM24, Table 1 and Fig. 2b and d). In these tests, samples were initially loaded past failure at either $P=50$ or 100 MPa (ϵ of about 2%), unloaded, and then reloaded at the same displacement rate, but at higher confining pressures. During the pressure-stepping experiments, any damage caused by strain localization did not significantly modify the elastic slope during subsequent loading cycles.

3.2. Pressure and temperature dependence of strength and sliding stress

The pressure dependence of strength at different temperatures is shown in Fig. 3a. We report the maximum differential stress, $\sigma_{\max}$, for samples that localized, and the differential stress at 1% permanent strain, $\sigma_{1\%}$, for samples that showed distributed deformation (see Table 1). In Fig. 3b we plot the sliding stress after strain localization, picked as $\sigma_{1\%}$; in all cases 1% permanent strain is achieved after the stress drop associated with localization. A modest P dependence of both strength and sliding stress is evident at all temperatures, despite the ~ 30 MPa scatter in the data (Fig. 3). The strength versus confining pressure data were fit to a linear Mohr failure envelope to obtain

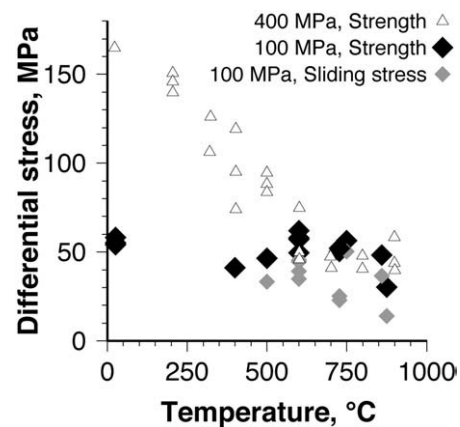


Fig. 4. Temperature dependence of strength and sliding stress of talc. At $P=100$ MPa (this study) talc displays no systematic T dependent weakening within the scatter of the strength data. At $P=400$ MPa (Edmond and Paterson, 1971a) talc shows a progressive decrease in strength with increasing T . At $T \geq 600$ °C the strength of talc is similar at both $P=100$ and $P=400$ MPa.

estimates of the uniaxial compressive strength (σ_0) and the slope of the envelope (μ'). These values were then used to calculate the coefficient of internal friction (μ_{int}) and the cohesive strength (τ_0) using the relationships (Paterson, 1978):

$$\mu_{\text{int}} = \frac{\mu'}{2\sqrt{1+\mu'}}; \tau_0 = \frac{\sigma_0}{2\sqrt{1+\mu'}}. \quad (1)$$

Similarly, Eq. (1) was used to calculate the coefficient of sliding friction (μ) and frictional cohesion (c_0) using the sliding stress data shown in Fig. 3b.

Both the strength (Fig. 3a) and the sliding stress data (Fig. 3b) show only modest effects of temperature between 400–600 °C. The P dependence of strength (Fig. 3a) corresponds to a decrease of μ_{int} from 0.14 at room T to 0.06 and 0.05 at 400 °C and 600 °C respectively. As shown in Fig. 3, our room T data agree reasonably well with previous studies (Edmond and Paterson, 1971a; Scruggs, 1997; Morrow et al., 2000; Moore and Lockner, 2004). With increasing T , μ decreases to 0.1 at both 400 °C and 600 °C, and to 0.11 and 0.08 at 400 °C and 600 °C respectively for the P -stepping experiments alone (Fig. 3b).

At $P \approx 100$ MPa we observe no systematic variation in strength or sliding stress with increasing T (Fig. 4). In contrast, experiments at $P=400$ MPa (Edmond and Paterson, 1971a) display a progressive reduction of strength from ~ 150 MPa at room T to ~ 50 MPa at $T \geq 600$ °C, a strength similar to that of talc deformed at 100 MPa in our study.

Linear fits to our strength data (Fig. 3) indicate a cohesive strength of $\tau_0 \approx 13$ at room T and 400 °C, and $\tau_0 \approx 20$ MPa at 600 °C. The apparent frictional cohesion (c_0) is lower, approximately 7 MPa at both 400 °C and 600 °C. Previous room T experiments indicate $c_0 \approx 0$ MPa at $P < 130$ MPa (Scruggs, 1997).

Small amounts of dehydration reaction products were observed in the samples deformed at 750 °C and 860 °C. However, as also noted by Edmond and Paterson (1971a), we did not observe a significant reduction in strength upon crossing the talc thermal stability limit. The approximately 10–25 MPa decrease in stress shown at 860 °C (Fig. 4) is comparable to sample-to-sample strength variability (Fig. 3).

3.3. Rate-dependence of frictional sliding and strength

We conducted experiments with imposed displacement rate (i.e., axial strain rate) changes to determine the velocity dependence of frictional sliding (Fig. 5a and b, Table 1). The nominal coefficient of sliding friction (μ_{nom}) is calculated from the sliding strength ($\sigma_1 - \sigma_3$), the confining pressure ($P = \sigma_3$) and the fault angle with respect to σ_1 (β) measured on cut polished surfaces of deformed cores (see Table 1). Assuming no cohesion within the fault (e.g., Jaeger and Cook, 1971):

$$\mu_{\text{nom}} = \frac{1/2(\sigma_1 - \sigma_3)\sin(2\beta)}{1/2(\sigma_1 + \sigma_3) + 1/2(\sigma_1 - \sigma_3)\cos(2\beta)}. \quad (2)$$

Displacement rate-stepping experiments on samples that displayed localized deformation suggest that talc is velocity strengthening (Fig. 5), although the velocity dependence of talc

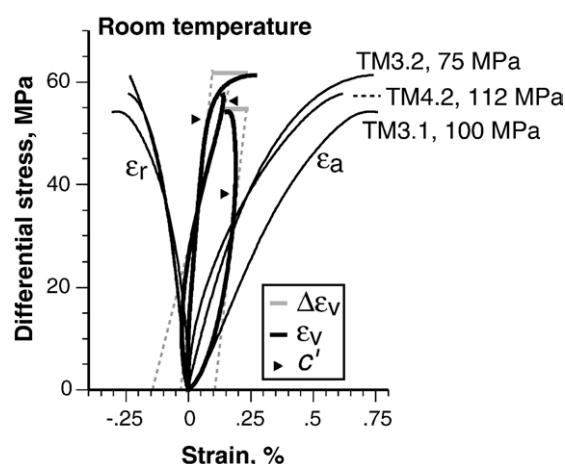


Fig. 5. Stress versus volumetric strain curves (bold lines, ε_v), calculated from the measured axial and radial strain (ε_a and ε_r). c' corresponds to the deviation from the linear trend of ε_v (dashed grey lines). The difference between observed ε_v and elastic strain at peak stress (gray solid lines, $\Delta\varepsilon_v$) corresponds to dilatancy ($\Delta\varepsilon_v < 0$) or to compaction ($\Delta\varepsilon_v > 0$).

friction is modest; at 50 MPa and 400 °C (TM07, Fig. 5b) $\Delta\mu_{\text{nom}} \approx 0.02$ for a strain-rate change between 3 and $9 \times 10^{-5} \text{ s}^{-1}$, and $\Delta\mu_{\text{nom}} < 0.01$ at 50 MPa and 600 °C (TM32, Fig. 5c). The modest changes in sliding stress are consistent with previous results on talc gouge deformed at room T and normal stresses ≤ 150 MPa (Scruggs, 1997). Samples that displayed distributed deformation showed a similar positive dependence on strain rate (TM06 and TM33, Fig. 5c, Table 1). We did not observe a systematic correlation between strain rate dependence and T or P . The jacket correction with strain-rate changes may not be adequate owing to the localization of deformation both in the sample and the jacket (Fig. 6).

Small stress drops were observed in both constant strain-rate and strain-rate stepping experiments at 600 °C (Fig. 2, Table 1), but not at 400 °C. While stick slip may be an indicator of velocity-weakening mechanisms, further calibration tests are required to determine the interplay between the dynamic machine behavior and the properties of talc. Hold tests conducted during some of the experiments (30 s for TM07, Fig. 5a, and 90 s for TM33, Fig. 5c) show an initial relaxation due to creep. The samples resume sliding with no apparent increase in strength (i.e., we did not resolve time dependent healing) after accounting for a linear interpolation of the strengthening trend observed prior to and after the hold. These results are consistent with the small increase in friction of $\Delta\mu < 0.003$ reported for talc gouge in slide-hold-slide tests (Scruggs, 1997).

3.4. Volumetric strain and permeability

At room temperature and low axial strain, the inelastic volumetric strains measured by strain gauges were modest. All experiments show non-linear axial strain during initial loading, which we attribute to seating of the pistons, sample, and jacket. After a subsequent interval of linear strain, three samples displayed either minor compaction or dilation at stresses of 70% to 95% of σ_{max} . Two samples provided no usable data due to

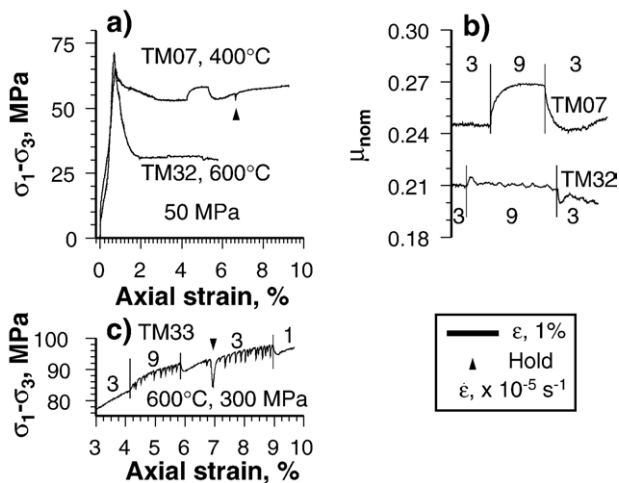


Fig. 6. a) Differential stress-axial strain curves for strain-rate stepping experiments at $P=50$ MPa, and $T=400$ °C and 600 °C. Rate changes were imposed when samples exhibited approximately constant strength after the initial strain localization. No jacket corrections are applied to the differential stress curves. b) Detail of the nominal coefficient of friction μ_{nom} vs. strain for the experiments shown in a). μ_{nom} is calculated from the differential stress after copper jacket correction using Eq. (2) (see Table 1 and [2]). c) Differential stress-axial strain curve for a strain-rate stepping experiment that displayed distributed deformation. Stick-slip behavior initiates after the first strain rate step and after the hold (arrow) in the second strain-rate step. The differential stress curve has no copper jacket correction.

strain-gauge failure or inhomogeneous strain distribution. The stresses at which the onset of dilation or compaction occurred, denoted as c' and c^* respectively (Fig. 7), were much nearer to the fracture strength in talc than those reported for granite or dunite (e.g., Brace et al., 1966; Shimada et al., 1983). In addition, the inelastic volumetric strain during the initial stages of failure, $|\Delta\epsilon_v|$, was less than that of the higher grade rocks ($|\Delta\epsilon_v| < 0.2\%$ for talc and $> 1\%$ for granite and dunite (Table 1 and Fig. 7). Small magnitudes of $|\Delta\epsilon_v|$, and the onset of inelastic volumetric strain at stresses near σ_{max} , have also been observed in serpentinites and weakly serpentinitized dunites (Escartín et al., 1997a, 2001).

A permeability test was carried out at $T=400$ °C and $P=50$ MPa using argon as a pore fluid (experiment TM09, Figs. 2a and 6b; Table 1). At these conditions, the permeability remains below the resolution of the apparatus ($\sim 10^{-20} \text{ m}^2$) after yielding and during localization of deformation. Any damage occurring during deformation at those conditions was therefore not efficiently connected.

4. Deformation textures

All samples deformed in the semibrittle regime, and displayed either localized or distributed deformation. Kinking accommodates plastic deformation, and its style varies with P and T , and strain. Below we first describe macroscopic and microstructural observations for experiments conducted at $T \leq 600$ °C, within the stability field of talc (Sections 4.1 and 4.2). Observations from samples deformed at $T \geq 750$ °C, which show evidence of dehydration, are described in Section 4.3.

4.1. Mode of deformation and shear zone structure

Examination of deformed samples suggests a strong correlation between macroscopic deformation mode and the shape of the stress–strain curves. Deformation localized in samples that strain weakened (e.g., TM3.2, TM09, TM28, Fig. 6), whereas deformation was distributed in samples that deformed with constant strength or that strain-hardened (e.g., TM33, TM10, and TM05, Fig. 6). Distributed deformation was observed at $P=300$ MPa and $T=400$ °C, and both localized and distributed deformation were observed at $P=300$ MPa and $T=600$ °C (Table 1). At room T Edmond and Paterson (1971b) report a transition at about $P=400$ MPa. In most cases strain localization occurred on a single fault (Fig. 6 and Table 1), however, 2 to 3 sub-parallel faults developed in some samples (e.g., TM04 and TM09, Fig. 6).

At $P \leq 100$ MPa and $T \leq 400$ °C, anastomosing shear bands developed with thicknesses varying from 100 to 500 μm (e.g., TM4.2 and TM07, Fig. 8a–c). Within the shear zone, a cataclastic foliation developed by grain-size reduction, rigid rotation and limited folding/kinking of individual grains (e.g., TM07, Fig. 8b and c). When deformation was stopped after the first evidence for yielding (e.g., TM20, Fig. 8e) samples exhibit highly localized shear zones. The shear zone geometry becomes more irregular in locations where the fault intersects talc grains that have (001) planes at high angles to the fault trace (Fig. 8e).

At $T \geq 400$ °C and $P=300$ MPa, distributed deformation is accommodated by a set of sub-parallel and closely spaced faults. Examination of the number of shear zones visible on the sample jackets, as well as deformation features seen on polished surfaces, indicates that the width of the deformation zone increases with strain (Fig. 6). Furthermore, in P -stepping experiments, a sample that faulted at $T=400$ °C and $P=50$ MPa and was subsequently deformed at $P=200$ and 300 MPa (TM08, Table 1, Figs. 2b and 6b) develop a broad band of distributed deformation (~ 5 mm, TM08 in Fig. 6b). By contrast, in a similar P -stepping experiment at $T=600$ °C (TM24, Table 1) deformation at $P=200$ and 300 MPa remained localized along a fault initially formed at $P=100$ MPa.

4.2. Deformation microstructures

Shear zone deformation was accommodated by both crystal–plastic and cataclastic mechanisms (Figs. 8–10). Within the shear zones, TEM observations show intense kinking and microcracking of talc grains (Fig. 9a, c and e). Less than 1 mm away from the optically visible shear zones (see Fig. 6), deformation intensity decreases significantly. Instead of the pervasive kinking found in the shear zones, TEM observations document minor bending of crystals and a decrease in the density of microcracks (Fig. 9b, d and f).

Distributed deformation (Fig. 10d) is accommodated by semibrittle mechanisms similar to those documented in the faulted samples. The intensity of microcracks, and the style of kink bands (Fig. 10a, d and e) are similar to those observed within and around fault zones (Fig. 9). While polished surfaces

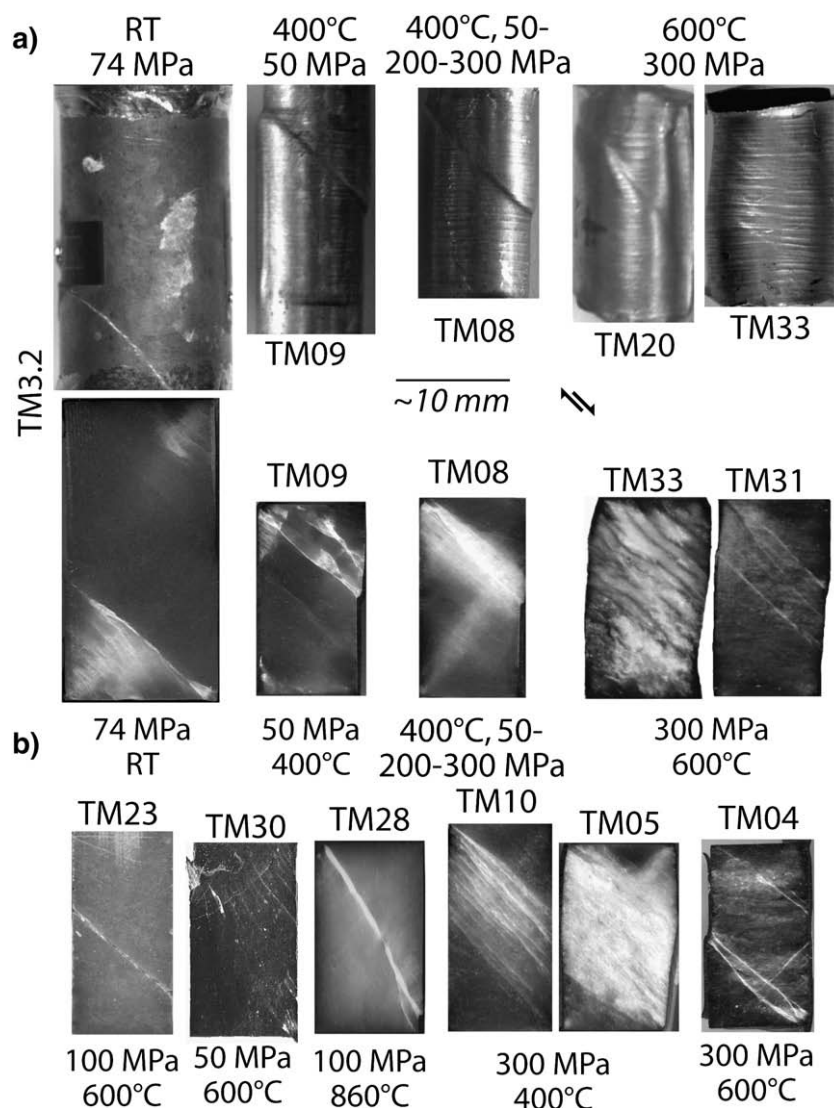


Fig. 7. Photographs of jacketed deformed samples (a) and of polished cut surfaces (b). Talc shows very efficient strain localization along a single fault, with evidence for distributed damage at low T (TM3.2) or P (TM09). Distributed deformation is observed at 300 MPa and 400 °C, while at 300 MPa and 600 °C both localized and distributed deformation occur (TM20, TM31 and TM33). At low ϵ (<2%) distributed deformation is characterized by numerous sub-parallel shear zones (TM10), that with increasing ϵ continue to develop while damage becomes widespread (TM05). At high T (TM28) deformation is localized along a shear zone with no evidence of damage away from it. P -stepping experiments (TM08) show a shear zone developed at low P , and distributed deformation around it developed at higher P . Fault angle is partially controlled by the weak foliation of the sample (TM30).

and jackets show that distributed deformation is at least partially accommodated by displacement on sets of sub-parallel faults (Fig. 6), at the TEM scale these faults are indistinguishable from the damaged matrix.

Both P and T influenced the micromechanisms of deformation. At low pressures, kink formation was accommodated by slip along (001) planes and grain boundaries, resulting in delamination of grains into thin (<100 nm) slabs (Fig. 9a and c). With increasing P , the size of the arcuate voids and microcracks formed at kink-band boundaries decrease (Fig. 10a–b). TEM observations suggest that while increasing T may enhance kinking (Fig. 10c, d and b), the intensity of microcracking remains unchanged (compare Figs. 9c and 10d, and Fig. 10b and e).

4.3. Dehydration microstructures

Samples deformed at temperatures above the thermal stability limit of talc developed single shear zones dipping at $\beta \sim 45^\circ$ at 750 °C (TM27 and TRA, Table 1) and at $\beta \sim 30^\circ$ at $T \geq 860$ °C (TM28 and TRF, Table 1 and Fig. 6b). The shear zones that formed at these conditions are similar to those formed at ~ 600 °C (compare Fig. 8e and f). TEM observations show intense kinking and microcracking within the shear zone (TM28, Fig. 9e–f).

Evidence for dehydration reactions was found within the shear zones. TEM showed porous regions within the fault zone where talc grains disaggregated along basal planes and lost their crystallographic structure (Fig. 11a and b). These textures were

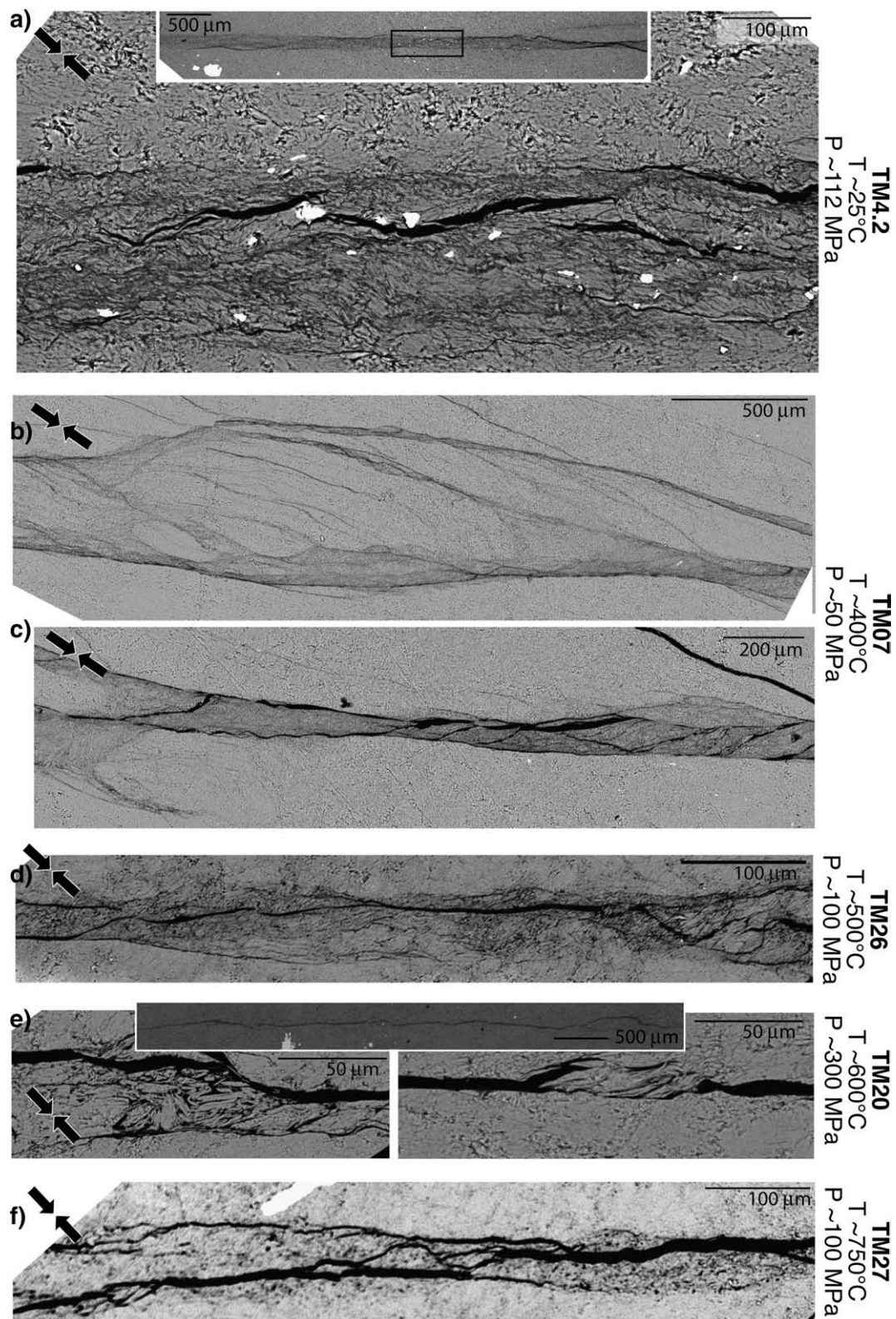


Fig. 8. SEM BSE images of fault zones at selected experimental conditions: room Temperature (a), 400 $^\circ\text{C}$ (b–c), 500 $^\circ\text{C}$ (d), 600 $^\circ\text{C}$ (e) and 750 $^\circ\text{C}$ (f). All image are oriented and arrows indicate the shortening direction.

associated with microcracking (Fig. 11a); however, the porosity within the disaggregated regions was greater than that associated with microcracking and kink-band formation at lower T (Fig. 9). In a sample deformed at 860 $^\circ\text{C}$, the porous regions also

contained dehydration reaction products, amorphous silica and distributed enstatite crystallites (Fig. 11c). Within approximately 1 mm of the fault zone, amorphous silica was also found in voids surrounded by unaltered talc with no enstatite (Fig. 11d).

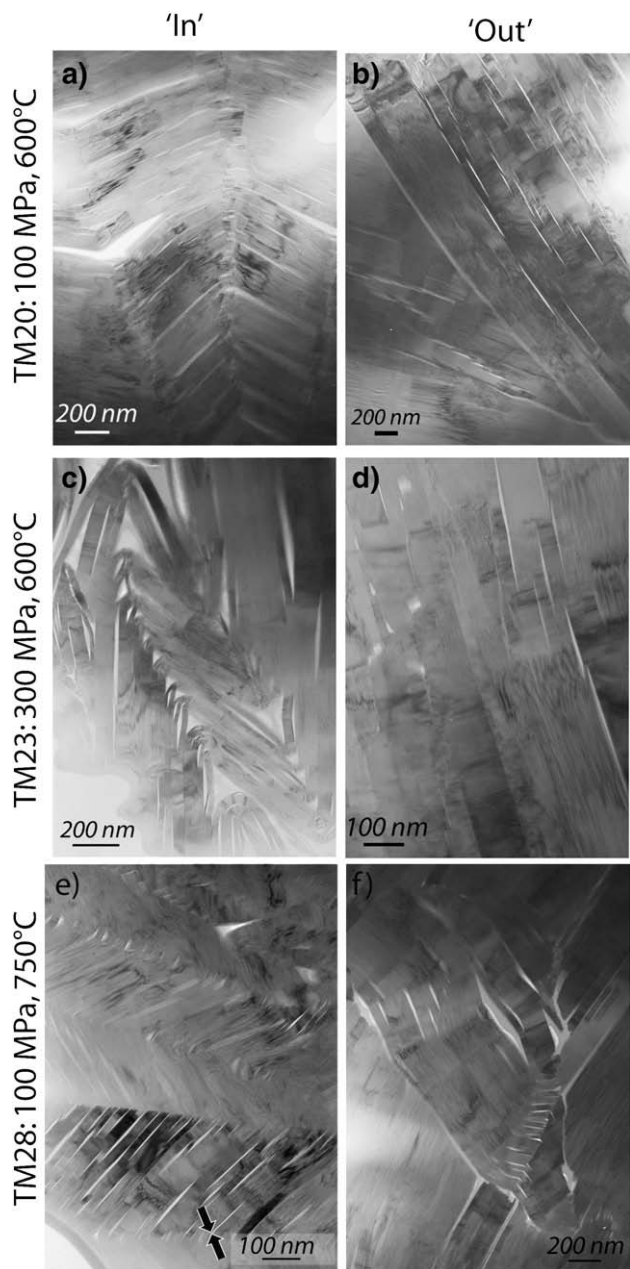


Fig. 9. Bright field TEM images of deformation microstructures within the shear zone ('In', left) and away from it ('Out', right) at different deformation conditions for samples TM20 (a–b), TM23 (c–d) and TM28 (e–f). Kinking develops in the shear zone, with associated microcracking. Away from the shear zone, talc shows mild bending and reduced kinking, with abundant microcracking along inter- and intra-granular (001) planes.

Apparently, silica was transported away from the reaction zone, along microcracks or voids associated with kinking. At $T=750\text{ }^{\circ}\text{C}$, some void walls were coated with an amorphous or microcrystalline phase, perhaps silica (Fig. 6a), but no associated enstatite crystals were identified. In all cases, evidence for dehydration was restricted to the shear zone.

5. Discussion

During conventional triaxial tests at laboratory strain rates, low temperatures, and low confining pressures, the fracture strength of

most low-porosity silicate rocks is larger than the confining pressure (Mogi, 2006; Paterson and Wong, 2005). Failure (i.e. a sudden drop in load-bearing capacity) is associated with the formation of a localized shear zone. As confining pressure is increased, μ_{int} decreases and becomes less than μ (for reviews see Paterson and Wong (2005), Mogi (2006), and Evans et al. (1990)). The magnitude of the permanent inelastic strain at failure, an indicator of ductility, often increases with pressure, particularly if conditions allow the material to deform in part by crystal plasticity (Mogi, 2006). Typically, as temperature and pressure are increased, the rock ceases to fail by localized rupture and deforms by distributed cataclastic strain, a state often called the brittle–ductile transition. At higher temperatures, when the rock deforms completely by crystal–plastic mechanisms, the pressure dependence of strength becomes very small (Heard, 1960; Evans et al., 1990). This transition is called the brittle–plastic transition, or the frictional-viscous transition behavior (Rutter, 1986; Mogi, 2006).

5.1. Macroscopic behavior of talc and the brittle–ductile transition

Our results, combined with those from previous investigations, illustrate several important differences between the mechanical behavior of talc and the generalized view outlined above. First, even at room temperature, intact cores of talc have very low cohesive strength and low coefficient of internal friction, $\mu_{\text{int}}=0.14$ (this study) or $0.1\text{--}0.15$ (Hickman et al., 1997). Second, for laboratory strain rates, the fracture strength of talc depends on pressure up to the temperature limit of its chemical stability, and deformation may be localized even at high temperatures. Third, the coefficient of sliding friction μ is small and nearly the same as μ_{int} : $\mu=0.08\text{--}0.10$ at $400\text{ }^{\circ}\text{C}$ and $600\text{ }^{\circ}\text{C}$ (this study); 0.15 at room T (Scruggs, 1997); or $0.24\text{--}0.36$ at room T (Morrow et al., 2000; Moore and Lockner, 2004). Both μ_{int} and μ decrease with increasing temperature. Fourth, the sliding strength on a newly formed fault increases slightly with increasing velocity, despite the fact that localization occurs at all temperatures. Fifth, the strength of unvented samples does not decrease by a large amount when the temperature exceeds that at which dehydration occurs.

At the sample or macroscopic-scale, the transition from localized deformation (associated with strain weakening) to distributed deformation (associated with strain hardening) occurred at a confining pressure of about 200 MPa at T to $400\text{ }^{\circ}\text{C}$ (Fig. 3a). This transition occurs at the intersection of the Mohr strength envelope and the friction law of talc, recalling the insight of Maurer (1965), Mogi (1966), and Byerlee (1968) that the brittle–ductile transition occurs when the cataclastic strength of a rock becomes less than the stress predicted for frictional sliding along an optimally oriented internal surface. A similar transition was observed in serpentinites deformed at room temperature (e.g., Escartín et al., 1997a). However, at higher temperatures, the transition was not as well defined; localized deformation was observed at $500\text{ }^{\circ}\text{C}$ (TM26, Table 1), and both localized and distributed deformation were observed at $600\text{ }^{\circ}\text{C}$ (Fig. 3a). Assuming Byerlee's (1968) assertion is correct, these observations suggest that the transition from localized to distributed deformation is poorly defined owing to

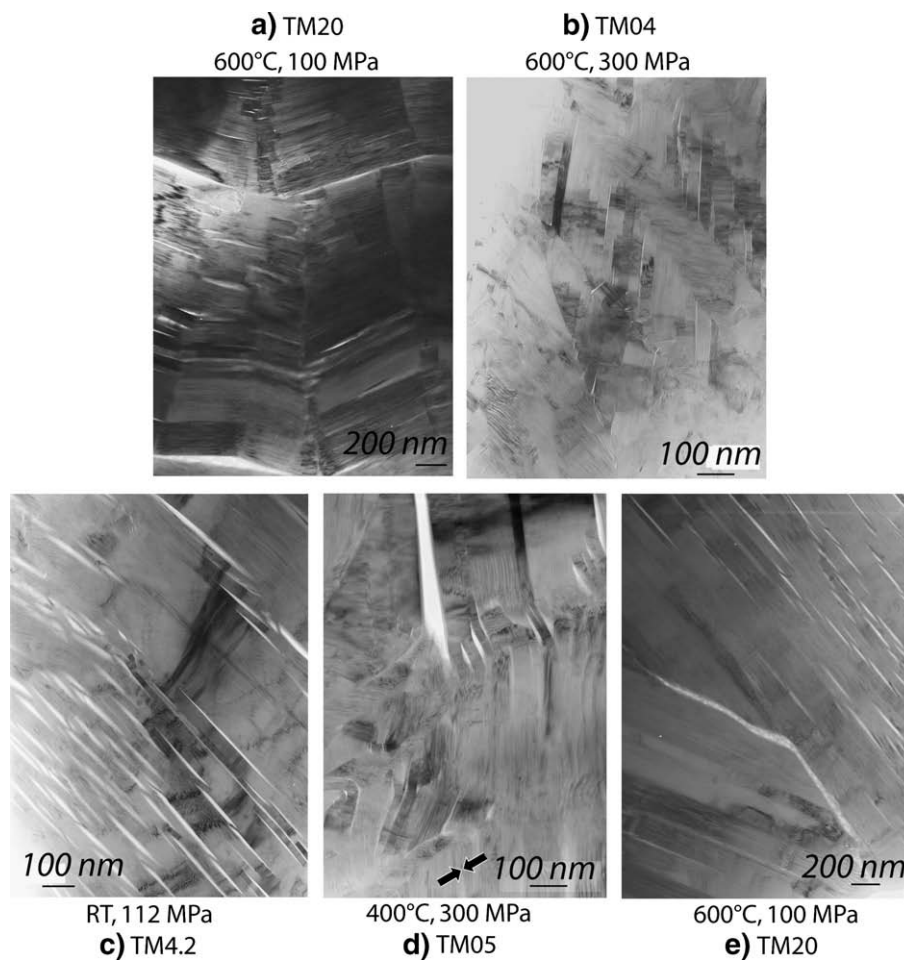


Fig. 10. Bright field TEM micrographs from experiments conducted at different P and T conditions. Porosity associated with kinking is reduced with increasing P (b and c). Microcracking along (001) planes is somewhat reduced with increasing P , but displays no T dependence (c–e).

small changes in μ_{int} and/or μ with increasing temperature (e.g., Fig. 12), pressure, or other parameters.

Deformation also seemed to be slightly more unstable at higher T . The small stress drops observed during stick–slip events became larger with increasing temperature at $T > 400$ °C (Fig. 5). However, we repeat our caution concerning the interplay of dynamic machine behavior with material behavior. Several of the unusual characteristics of talc could arise owing to dehydration. For example, dehydration might promote both localization and stick slip deformation. However, we did not observe any evidence for dehydration at 600 °C.

5.2. Micromechanics of deformation and the brittle–plastic transition

Microstructural observations indicate that deformation at all conditions was accommodated by a combination of crystal plasticity, frictional sliding, and cataclasis. Slip is likely restricted to the basal plane of talc, either by the motion of full or partial dislocations or by large-scale shear rupture. Plastic flow along non-basal planes may be difficult to achieve, as the stress necessary to activate such systems is very large. Qualitatively, this large anisotropy in slip strength can be attributed to weak bonding

across the basal plane (e.g., Morrow et al., 2000), and much stronger bonding within (001). At most, two independent strain components may be generated by slip on a single plane, thus, basal slip is insufficient to promote fully plastic creep (i.e., satisfy the von Mises criterion (Paterson, 1969)). The reduction of μ_{int} with increasing temperature is consistent with an enhancement of crystal plasticity (Fig. 9). The small voids associated with kinking and bending (Fig. 9), even near the dehydration temperature, indicate that some dilatancy is always necessary, and accounts for the modest pressure dependence of the fracture and sliding strength (Fig. 3).

The competition between crystal plasticity and dilatant cataclasis in talc may be responsible for the complicated and ambiguous range of deformation modes displayed. Plastic deformation (bending and kinking) is probably either modestly velocity strengthening or, at the extreme, rate-independent, as suggested by the strain rate-stepping experiments (Fig. 5). Stick–slip events, if attributable to material response, may arise from the local breakdown of cohesion within shear zones, as voids are produced. Grains or groups of grains might then rotate to orientations less advantageous for basal slip, causing hardening and broadening of the shear zone, as observed in P -stepping experiments (TM08, Fig. 6b). Such a scenario is consistent with

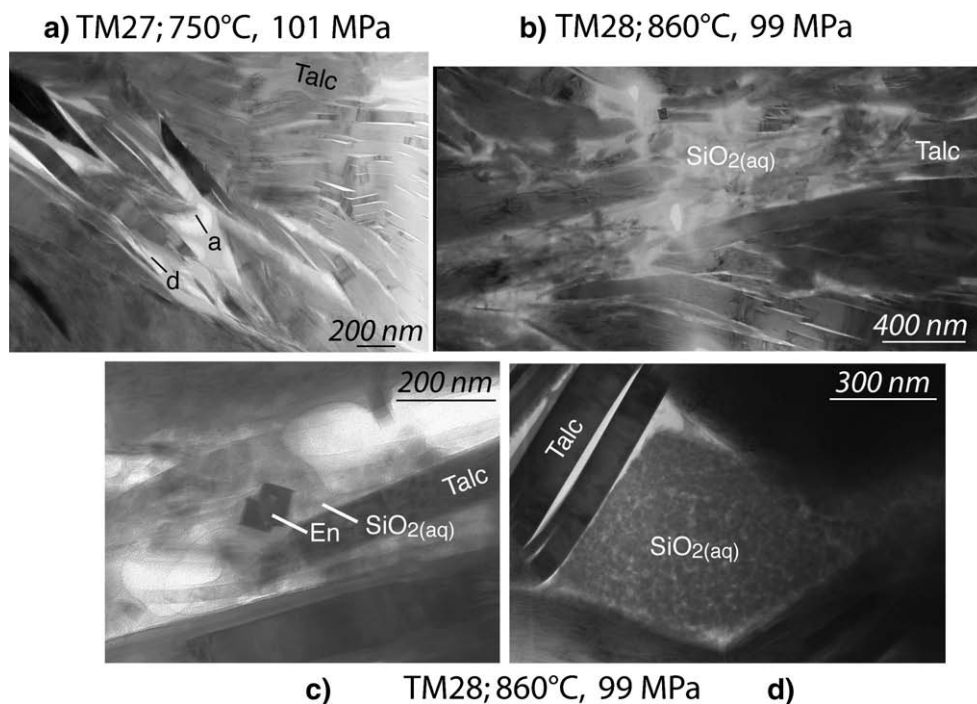


Fig. 11. Bright field TEM micrographs from experiments conducted at $P=100$ MPa and $T=750$ °C and 860 °C. a) At $T=750$ °C talc locally disaggregates in very thin lamellae near voids created during deformation (labeled d). An amorphous or microcrystalline phase (labeled a) is observed associated with this disaggregation, and along the grain boundaries surrounding voids. b) The sample deformed at $T=860$ °C shows localized zones that are highly porous within the fault zone. c) Talc dehydration results in the precipitation enstatite (En) and amorphous silica ($\text{SiO}_2(\text{aq})$) within these porous zones. d) Amorphous silica is also found within pores approximately 1 mm away from the fault zone, suggesting limited mobility of the dehydration products along the fault zone.

small volumetric strains (Fig. 7), because the zone in which dilation occurs is a very small volume fraction of the sample.

The extreme anisotropy of slip strength also has implications for the transition from cataclasis to full crystal plasticity that often occurs with increasing temperature, and for the transition from localized to ductile flow, that typically occurs with increasing pressure. Because slip on the basal plane of phyllosilicates is very easy, even at room temperature, rocks composed of talc and serpentine (Escartin et al., 1997a) have initial yield stresses that are much lower than those of other silicates, e.g., the yield stress for talc is <10% of that of dunite and ~20% of that of lizardite. In addition, the coefficients of friction in such rocks are much lower than 0.6, i.e., Byerlee's law. The latter property allows localized failure to occur at lower differential stresses than in most other silicates. The extreme anisotropy in yield stress results in a low coefficient of internal friction, which persists to thermal stability limit of talc. The net result is that, for laboratory test conditions, the aggregate strength never rises to the level of the confining pressure, nor does the material achieve full crystal plasticity at the conditions tested. Thus, two common rules of thumb for deformation mode applicable to other intact silicate rocks, Byerlee's law and the Goetze criterion (Evans et al., 1990; Kohlstedt et al., 1995), do not strictly apply to talc rocks.

5.3. Consequences of dehydration

Talc dehydration during subduction is important for the evolution of strength along faults within the slab. In our

experiments, dehydration products were observed only within the porous shear zones (e.g., TM28, Figs. 6b and 10), and not within the remainder of the sample, suggesting that the temperature overstep was not large enough to produce reaction in the bulk during the time of the experiment. Apparently, the reaction progress is enhanced by the creation of porosity associated with kinking and microcracking. If the pore space is effectively undrained, dilatancy could reduce the local pore pressure and, consequently, lower the chemical activity of the reaction products. Additionally, delamination during deformation would increase the surface area of the reactants. Either might be expected to increase the reaction rate.

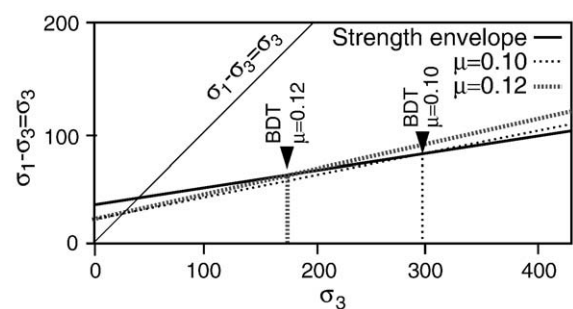


Fig. 12. Variation in the position of the brittle to ductile transition (BDT) induced by small changes in the coefficient of sliding friction. Small changes in μ can significantly modify the point of intersection between the strength envelope and the friction law, which corresponds to the BDT. Changes in T , P and other parameters can also modify the coefficient of internal friction μ_{int} , the cohesive strength τ_0 , or the frictional cohesion c_0 (see Eq. (1)) of talc, resulting in a poorly defined BDT.

Judging from the presence of amorphous silica away from reaction zones, we suppose that water and other reaction products were mobilized by dilatancy, even if only over very short distances. If sufficient water were liberated by dehydration, the effective confining pressure would be lowered, thus promoting weakening. The lack of dehydration-related weakening in our experiments (Fig. 3b) might be attributed the relative rates of reaction kinetics and dilatancy (e.g., Ko et al., 1997; Wong et al., 1997). If dehydration kinetics were accelerated relative to the rate of dilation, the effective confining pressure and therefore the aggregate strength might be lowered more efficiently. At natural conditions, active fault zones may thus serve as the locus of dehydration reactions prior to bulk dehydration of the host rock.

5.4. Implications for long-lived faulting

While talc is not a major component of the lithosphere, it is commonly recovered from both long-lived oceanic faults (e.g., Escartín et al., 2003; Schroeder and John, 2004; Boschi et al., 2006), and more recently from continental faults (e.g., ICDP, 2005; Moore and Rymer, 2007; Floyd et al., 2001; Taylor and Huchon, 2002) such as San Andreas. Talc can thus have a very important weakening effect, even if present in small proportions. Its mode of deformation can contribute to elevated pore fluid pressures, as proposed for serpentinites (Escartín et al., 1997a,b). Both talc and serpentine are stable to high pressures, but talc is stable to higher T (~700 °C) than antigorite (~600 °C) or lizardite (~400–500 °C). In addition to a more important weakening effect than serpentinites (Escartín et al., 1997b), talc can thus promote strain localization and the weakening of faults. This effect can thus affect almost the complete thickness of the lithosphere and faults rooting close to the brittle to plastic transition ($T \sim 750$ °C (Hirth et al., 1998)). Talc can also form in high silica activity conditions associated with the alteration of gabbros of the oceanic lithosphere, and possibly linked to hydrothermal activity near mid-ocean ridges (McCaig et al., 2007). The presence of talc in subducting oceanic lithosphere can also contribute to the weakening of the slab beyond the stability field of serpentinites and to P of at least 8 GPa (Pawley and Wood, 1995). Fluids from the dehydrating slab wedge can hydrate the overlying mantle wedge, and result in talc formation along olivine grain boundaries. Motivated by results on weakly serpentinitized peridotites (Escartín et al., 2001), we speculate that substantial weakening may occur even with small amounts of talc of 10% or less.

6. Conclusions

Deformation experiments conducted on intact cores of talc at T up to 860 °C and P up to 300 MPa demonstrate that while talc's strength is well below the confining pressure, deformation localized even at high temperatures. There is also a modest pressure dependence of strength corresponding to $\mu_{\text{int}} \sim 0.14$ at room temperature decreasing to $\mu_{\text{int}} \leq 0.1$ at $T \geq 400$ °C; the coefficient of sliding friction μ shows similar values and T dependence. At $P = 300$ MPa only distributed deformation is observed at 400 °C, while both localized and distributed deformation are observed at

600 °C. The poorly defined transition from localized to distributed deformation may arise owing to small variations in μ_{int} and μ with changes in temperature and pressure. Semibrittle deformation is observed at all conditions, but crystal plasticity (bending and kinking) is enhanced with temperature, consistent with the modest decrease in both μ_{int} and μ . Deformation microstructures within faults and distributed deformation zones result from a combination of kinking and microcracking, both of which are strongly controlled by deformation on the weak (001) plane. Samples are nominally non-dilatant at room temperature, suggesting that only a minor volume of the sample shows significant deformation. Evidence for incipient talc dehydration is observed at 750 °C, and dehydration products (enstatite crystallites and amorphous silica) are observed at ~860 °C. Dehydration microstructures are restricted to the shear zones, and result in highly porous zones. The creation of void space and the increase of surface area along faults apparently promotes dehydration reactions, and can facilitate fluid transport along them.

Talc's strength, which is <10% of that of peridotite and ~20% of that of lizardite, and its mode of deformation can have important rheological consequences for many oceanic and continental faults where talc is common. In subduction zones talc can weaken both the slab, and the overlying mantle wedge if it is hydrated. In these tectonic environments active brittle deformation on faults can also promote dehydration reactions that in turn affect the rheology and pore fluid pressure along shear zones.

Acknowledgments

This work was supported by NSF grant EAR0405709 to GH, OCE 045613 to BE and by CNRS/INSU Diety to JE and MA. We are grateful to J. Hirst for sample preparation and to X. Xiao, B. de Martin, N. Austin and U. Mok for discussions and their assistance during laboratory experiments at the Massachusetts Institute of Technology. S. Borensztajn (LISE, CNRS/UPMC) is acknowledged for technical assistance and skills during SEM observations. TEM observations were carried out at the French TEM National Facility at CRCM/Marseille. Mondo Minerals and E. Rutter at U. of Manchester kindly provided talc material used in this study. This is IPGP contribution #2339.

References

- Andreani, M., Escartín, J., Hirth, G. & Evans, B. (2006). Deformation mode of talc from subsurface to dehydration conditions: implications for the subducting oceanic lithosphere, *Eos Trans. AGU*, 87(52), Fall Meet. Suppl., Abstract V31C-0599.
- Boschi, C., Fruh-Green, G., Escartín, J., 2006. Occurrence and significance of serpentinite-hosted, talc-rich fault rocks in modern oceanic settings and ophiolite complexes. *Ophioliti* 31, 141–150.
- Brace, W.F., Paulding, B.W., Scholz, C., 1966. Dilatancy in the fracture of crystalline rocks. *J. Geophys. Res.* 71, 3939–3953.
- Byerlee, J.D., 1968. Brittle–ductile transition in rocks. *J. Geophys. Res.* 73, 4741–4750.
- Deer, W.A., Howie, R.A., Zussman, J., 1962. *Rock-forming Minerals. Sheet silicates*, vol. 3. Longman, London.
- D'Orazio, M., Boschi, C., Brunelli, D., 2004. Talc-rich hydrothermal rocks from the St. Paul and Conrad fracture zones in the Atlantic Ocean. *Eur. J. Miner.* 16, 73–83.

- Edmond, J.M., Paterson, M.S., 1971a. Strength of solid pressure media and implications for high pressure apparatus. *Contrib. Miner. Petrol.* 30, 141–160.
- Edmond, J.M., Paterson, M.S., 1971b. Volume changes during the deformation of rocks at high pressures. *Int. J. Rock Mech. Min. Sci.* 9, 161–182.
- Escartín, J., Hirth, G., Evans, B., 1997a. Non-dilatant brittle deformation of serpentinites: implications for Mohr–Coulomb theory and the strength of faults. *J. Geophys. Res.* 102, 2897–2913.
- Escartín, J., Hirth, G., Evans, B., 1997b. Effects of serpentinization on the lithospheric strength and the style of normal faulting at slow-spreading ridges. *Earth Planet Sci. Lett.* 151, 181–190.
- Escartín, J., Hirth, G., Evans, B., 2001. Strength of slightly serpentinized peridotites: implications for the tectonics of oceanic lithosphere. *Geology* 29 (11), 1023–1026.
- Escartín, J., Mével, C., MacLeod, C.J., McCaig, A.M., 2003. Constraints on deformation conditions and the origin of oceanic detachments: the Mid-Atlantic Ridge core complex at 15°45'N. *Geochem. Geophys., Geosys.* 4, 1067. doi:10.1029/2001GC000278.
- Evans, B., Fredrich, J.T., Wong, T.F., 1990. The brittle–ductile transition in rocks: recent experimental and theoretical progress. In: Duba, A.G., et al. (Ed.), *The brittle–ductile transition in rocks*. AGU Geophys. Monog., vol. 56, pp. 1–20.
- Floyd, J.S., Mutter, J.C., Goodliffe, A.M., Taylor, B., 2001. Evidence for fault weakness and fluid flow within an active low-angle normal fault. *Nature* 411, 779–783.
- Frost, H.J., Ashby, M.F., 1982. *Deformation Mechanism Maps — The Plasticity and Creep of Metals and Ceramics*. Pergamon Press, Oxford. 184 pp.
- Heard, H.C., 1960. Transition from brittle fracture to ductile flow in Solenhofen limestone as a function of temperature, confining pressure, and interstitial fluid pressure, Chapter 7, http://www-mi8.csa.com.ezp1.harvard.edu/ids70/view_record.php?id=2&recnum=5&log=from_res&SID=860a4b52384d67129e71e97670dae177, Memoir - Geological Society of America, pp. 193–223.
- Hickman, J.B., Zhurina, E.N., Kronenberg, A.K., 1997. Deformation of talc and pyrophyllite: disruption of van der Waals bonds and comparison with calculated forces. *Eos Trans. AGU.* 78, F724.
- Hirth, G., Escartín, J., Lin, J., 1998. The rheology of the lower oceanic crust: implications for lithospheric deformation at Mid-Ocean ridges. In: Buck, W.R., et al. (Ed.), *Faulting and magmatism at mid-ocean ridges*. AGU, Washington D. C., pp. 291–303.
- ICDP, 2005. SAFOD Mud Log. . published on-line, http://www.icdp-online.de/contenido/icdp/upload/pdf/safod/Phase2_Mud_Log_PDF.pdf.
- Jaeger, J.C., Cook, N.G., 1971. *Fundamentals of Rock Mechanics*. Chapman and Hall. 593 pp.
- Ko, S.C., Olgaard, D.L., Wong, T., 1997. Generation and maintenance of pore pressure excess in a dehydrating system, 1, experimental and microstructural observations. *J. Geophys. Res.* 102, 825–840.
- Kohlstedt, D.L., Evans, B., Mackwell, S.J., 1995. Strength of the lithosphere — constraints imposed by laboratory experiments. *J. Geophys. Res.* 100, 17587–17602.
- MacLeod, C.J., Escartín, J., Banerji, D., Banks, G.J., Gleeson, M., Irving, D.H.B., Lilly, R.M., McCaig, A., Niu, Y.L., Allerton, S., Smith, D.K., 2002. Direct geological evidence for oceanic detachment faulting: the Mid-Atlantic Ridge, 15°45'N. *Geology* 30, 279–282.
- Maurer, W.C., 1965. Shear failure of rock under compression. *Soc. Petrol. Eng. J.* 5, 161–176.
- McCaig, A., Cliff, B., Escartín, J., Fallick, A.E., MacLeod, C., 2007. Pressure dependence of rock strength and transition from brittle fracture to ductile flow. *Bull. Earthquake Res. Inst. Tokyo Univ.* 35, 935–938. doi:10.1130/G23657A.1.
- Mogi, K., 1966. Pressure dependence of rock strength and transition from brittle fracture to ductile flow. *Bull. Earthquake Res. Inst. Tokyo Univ.* 42, 215–232.
- Mogi, K., 2006. *Experimental Rock Mechanics*. Taylor & Francis. 361 pp.
- Moore, D.E., Lockner, D.A., 2004. Crystallographic controls on the frictional behavior of dry and water-saturated sheet structure minerals. *J. Geophys. Res.* 109, B03401. doi:10.1029/2003JB002582.
- Moore, D., Rymer, M.J., 2007. Talc-bearing serpentinite and the creeping section of San Andreas Fault. *Nature* 448, 795–797.
- Morrow, C.A., Moore, D.E., Lockner, D.A., 2000. The effect of mineral bond strength and absorbed water on fault gouge frictional strength. *Geophys. Res. Lett.* 27, 815–818.
- Paterson, M.S., 1969. The ductility of rocks. In: Argon, A.S. (Ed.), *Physics of Strength and Plasticity*. MIT Press, Cambridge MA, pp. 377–392.
- Paterson, M.S., 1978. *Experimental Rock Deformation — The Brittle Field*. Springer-Verlag, Heidelberg-Berlin-New York, N.Y.. 254 pp.
- Paterson, M.S., 1990. Rock deformation experimentation. In: Duba, A.G., et al. (Ed.), *The Brittle–Ductile Transitions in Rocks: The Heard Volume*. Geophys. Monograph. Ser., vol. 56. AGU, Washington D. C., pp. 187–194.
- Paterson, M.S., Wong, T.F., 2005. *Experimental Rock Deformation — The Brittle Field*. Springer, Berlin. 358 pp.
- Pawley, A.R., 1998. The reaction talc + forsterite = enstatite + H₂O: new experimental results and petrological implications. *Am. Mineral.* 83, 51–57.
- Pawley, A.R., Wood, B.J., 1995. The high pressure stability of talc and 10 Å phase: potential storage sites for H₂O in subduction zones. *Am. Mineral.* 80, 998–1003.
- Peacock, S.M., Hyndman, R.D., 1999. Hydrous minerals in the mantle and the maximum depth of subduction thrust earthquakes. *Geophys. Res. Lett.* 26, 2517–2520.
- Rutter, E.H., 1986. On the nomenclature of mode of failure transitions in rocks. *Tectonophysics* 122, 381–387.
- Schroeder, T., John, B.E., 2004. Strain localization on an oceanic detachment fault system, Atlantis Massif, 30°N, Mid-Atlantic Ridge. *Geochem. Geophys., Geosys.* 5, Q11007. doi:10.1029/12004GC00728.
- Scruggs, V.J., (1997), *Frictional properties and related microstructures of albite, muscovite, biotite and talc*, 322 pp, PhD Thesis, Brown University, Providence, RI.
- Shimada, M., Cho, A., Yukutake, H., 1983. Fracture strength of dry silicate rocks at high confining pressures and activity of acoustic emission. *Tectonophysics* 96, 159–172.
- Summers, R., Byerlee, J., 1977. Summary of results of frictional sliding studies, at confining pressures up to 6.98 kb, in selected rock materials. U.S. Geological Survey, Open-file Report 77–142. 139 pp.
- Taylor, B., Huchon, P., 2002. Active continental extension in the Western Woodlark Basin: a synthesis of Leg 180 results. In: Huchon, P., et al. (Ed.), *Active continental extension in the western Woodlark Basin: a synthesis of Leg 180 results*, pp. 1–36.
- Wong, T., Ko, S., Olgaard, D.L., 1997. Generation and maintenance of pore pressure excess in a dehydrating system, 2, theoretical analysis. *J. Geophys. Res.* 102, 841–852.
- Xiao, X., Evans, B., 2003. Shear-enhanced compaction during non-linear viscous creep of calcite–quartz aggregates. *Earth Planet Sci. Lett.* 216, 725–740.