

Incohesive		Random fabric			Foliated					
		Fault breccia visible fragments > 30% of rock mass								
		Fault gouge visible fragments < 30% of rock mass			Foliated gouge					
Cohesive	Nature of matrix	Glass, devitrified glass	Pseudotachylyte							
			Crush breccia fragments > 0.5 cm Fine crush breccia 0.1 < fragments < 0.5 cm Crush microbreccia fragments < 0.1 cm			0–10				
		Cataclasite series				Proto-cataclasite	Protomylonite		10–50	
							Cataclasite	Mylonite	50–90	
			Ultra-cataclasite	Ultra-mylonite	Phyllonite series				90–100	
		Pronounced grain growth				Blastomylonite				

Figure 1: Table of the terminology used for fault rocks. [Redrawn from table 3.1 in Scholz (1990)].

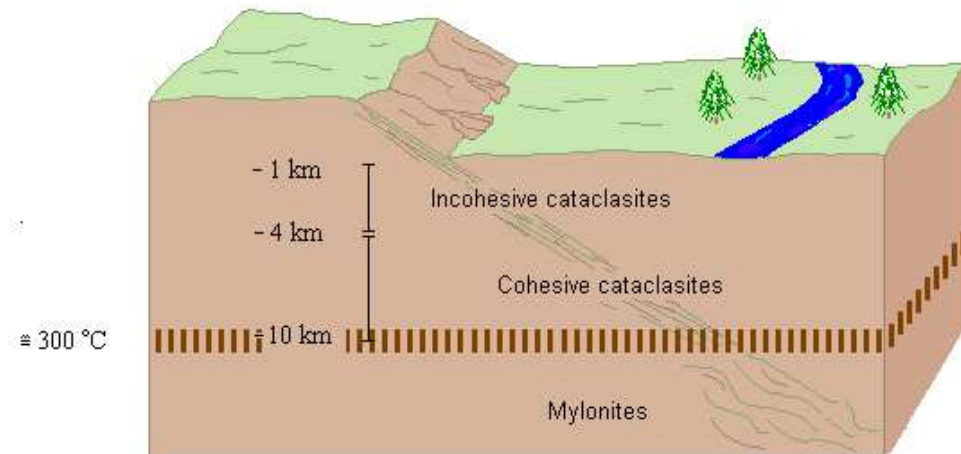


Figure 2: Schematic drawing of a fault scarp with associated landslides. The transition with depth of the fault rock types around the fault zone is also illustrated. Incoherent cataclasites are characteristic of the first 4 km of crust; coherent cataclasites can be found from about 4 km depth up to 10–15 km. At greater depths, when temperature reaches 250–350 °C, mylonites are present.

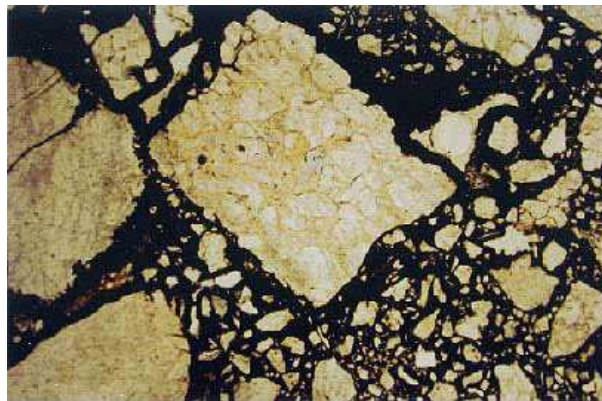


Figure 3: Photomicrograph of fault breccia in the Antietam Formation, Blue Ridge province (Virginia). Breccias form when rocks are extensively fractured in fault zones and are cemented together when minerals precipitate in the cracks and fractures. Note the angular fragments of quartz sandstone in a matrix of fine-grained iron oxide cement. Field of view 4×2.7 mm, Cross Polarized Light. Photo by Christopher M. Bailey, College of William & Mary.

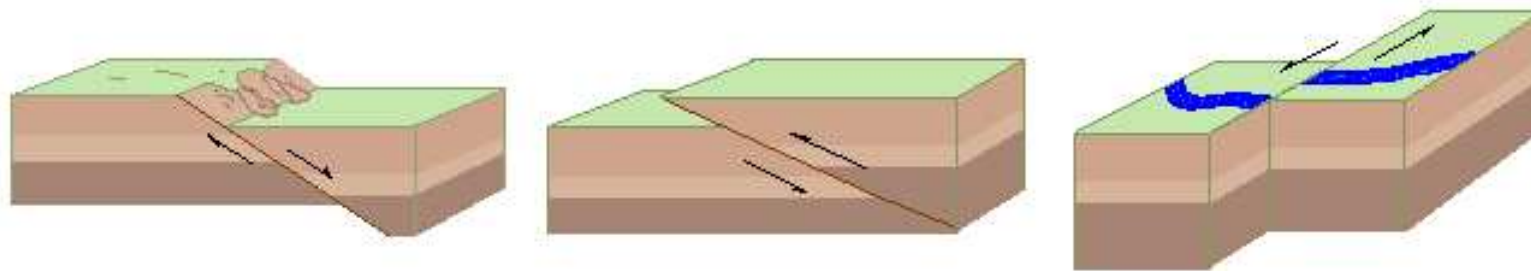


Figure 4: From left to right: A normal, reverse and strike-slip fault. In the normal fault the top block moves down relative to the bottom block; in a reverse fault the top block moves up relative to the bottom block; in a strike-slip fault the two blocks displace laterally. Natural faults are usually derived by a combination of these end-members.

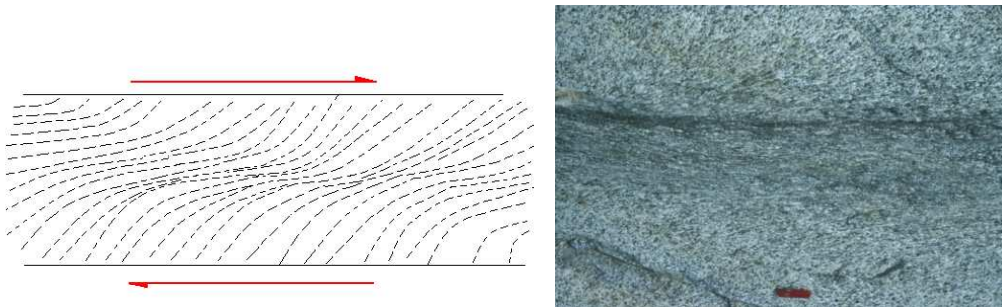


Figure 5: In a ductile shear zone the shear sense (arrows) can be inferred from the curvature of the foliation determined by the parallel alignment of the platy minerals. On the right, a mylonitic shear zone (amphibolitic facies) in a granitoid rock from the Adamello Massif (Italy) is shown. The sense of shear is dextral. Photo by G. Di Toro.

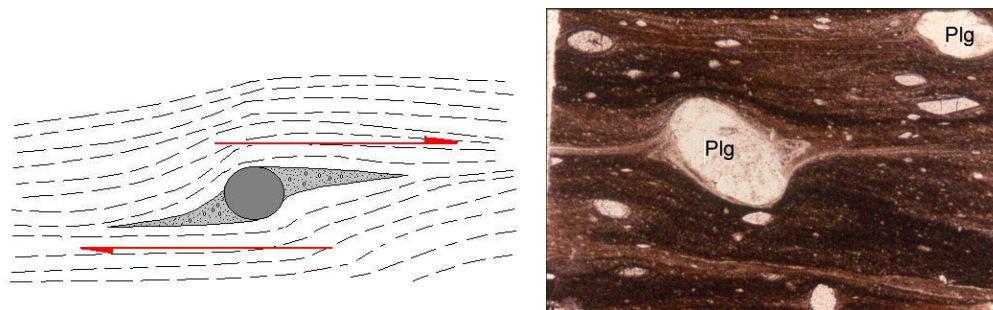


Figure 6: In a ductile shear zone, the sense of shearing can be inferred from the asymmetry of the tails of the recrystallized porphyroclastic material around the porphyroblast. The photo shows thin tails fed by a plagioclase porphyroblast. The orientation of the tails indicates a dextral sense of shear. The horizontal size of the picture is 1 mm. Photo by G. Di Toro.

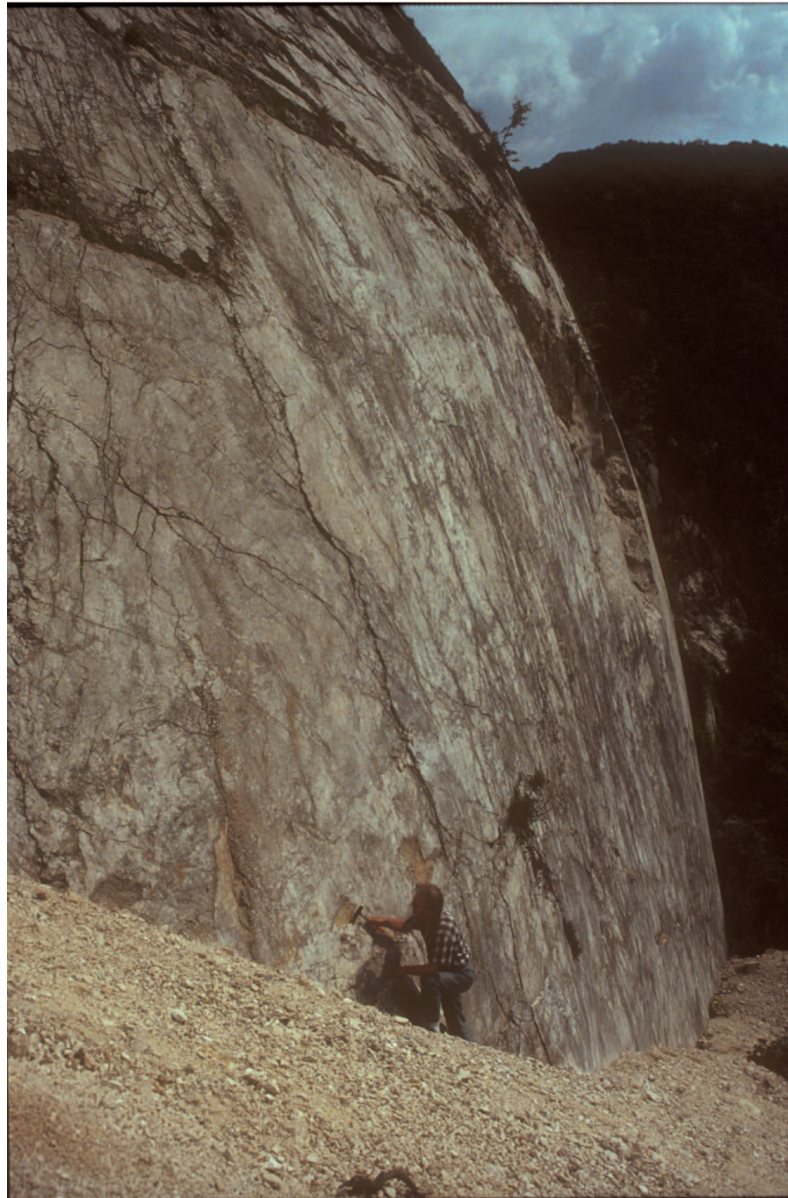


Figure 7: A very clear normal fault scarp in Carnia (North-Eastern Italy). Photo by G. B. Vai.



Figure 8: Example of pseudotachylyte fault vein and injection veins from the Adamello Massif (Southern Alps, Italy). The pseudotachylyte is the black vein with whitish clasts on the granodioritic host rock and it is associated to cataclasite (green bands). Photo by G. Di Toro.

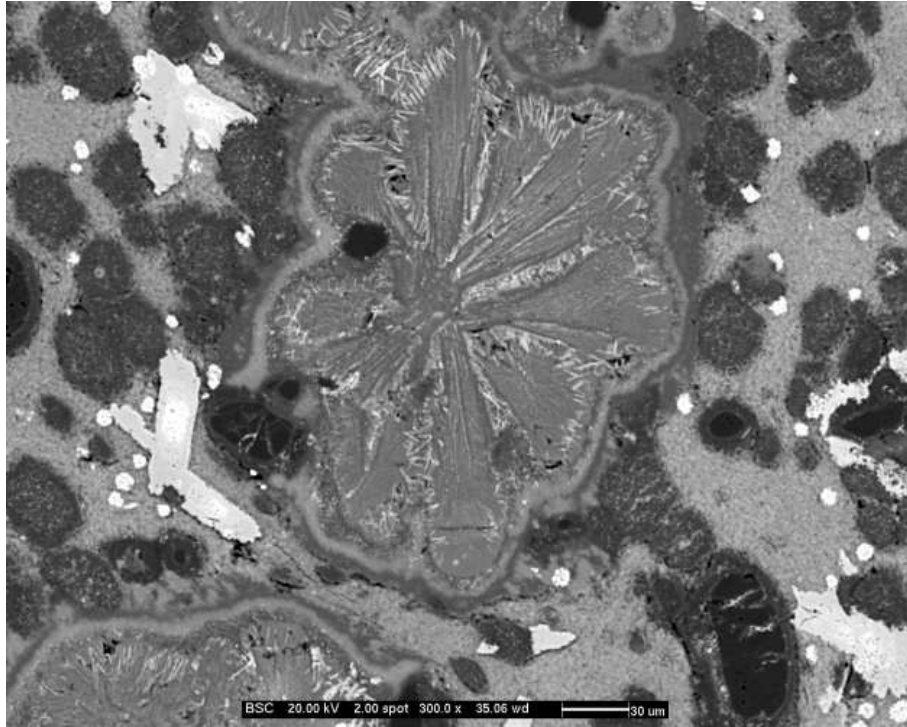


Figure 9: Cooling structures visible at the Scanning Electron Microscope (SEM) in a pseudotachylyte from the Adamello Massif (Southern Alps, Italy). Spherulitic structures consist of a plagioclase core rimmed by plagioclase microlites (middle gray) and very small acicular biotite (white). Spherulites are externally rimmed by devitrification haloes of K-feldspar (middle-bright gray) and silica (dark gray). Quartz grains are black and rounded, suggesting interaction with the melt and dissolution. Bright small dots (5 μm) are clusters of titanite and biotite, probably crystallized in the residual melt. Late overgrowth of epidote (large bright rods), suggests host rock temperature of 250–300 $^{\circ}\text{C}$. These microstructures are indicative of very rapid energy release with instantaneous melting and rapid cooling in the upper crust, providing additional evidence for their episodic (earthquake) origin. Photo by G. Di Toro.

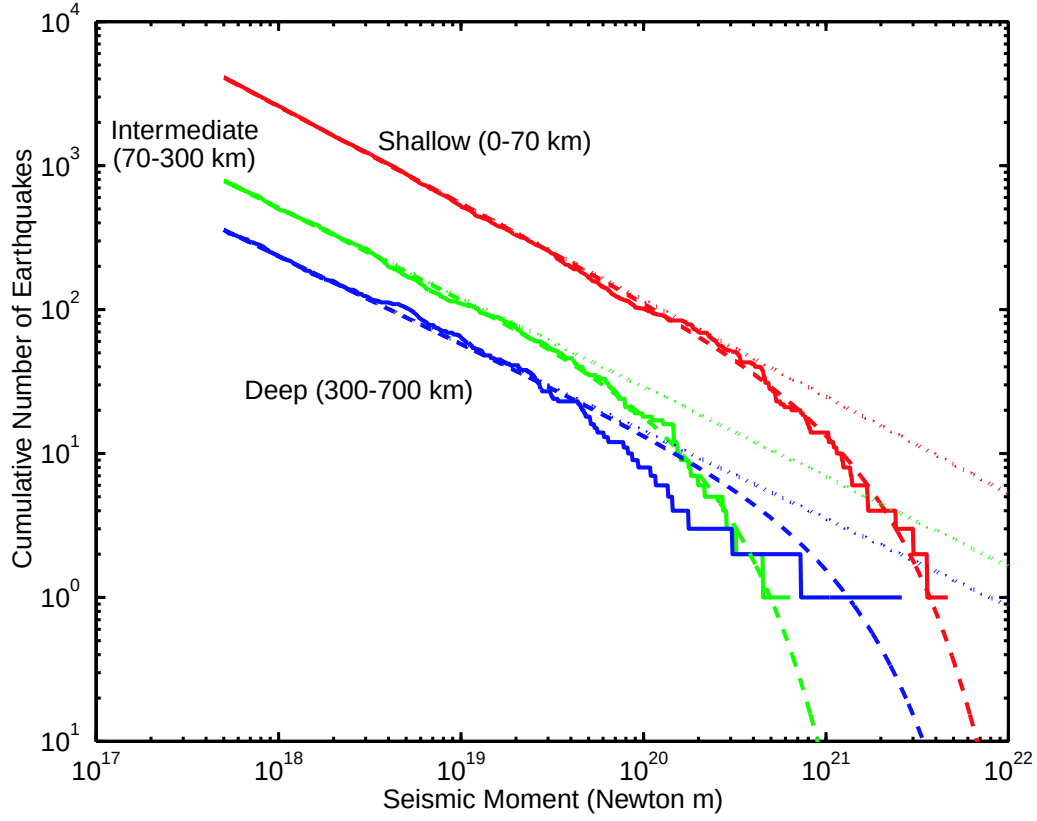


Figure 10: Cumulative number of earthquakes *versus* seismic moment for the global earthquake distribution in the January 1, 1977 – December 31, 2001 Harvard catalogue. The solid curves show the numbers of events with moment greater than or equal to M for deep, intermediate and shallow earthquakes. The dotted (straight) lines are the classic G-R distribution (equation 2.2) while the dashed curves which taper off exponentially are the tapered G-R distribution (equation 2.17). The latter fit the data whereas the former do not. The slopes of the linear part of the curves correspond to β equal to 0.672 ± 0.011 , 0.623 ± 0.024 , 0.608 ± 0.033 , and the corner moment $M_c = 1.6 \times 10^{21}$, 2.1×10^{20} , 1.2×10^{21} Nm, for shallow, intermediate, and deep earthquakes, respectively.

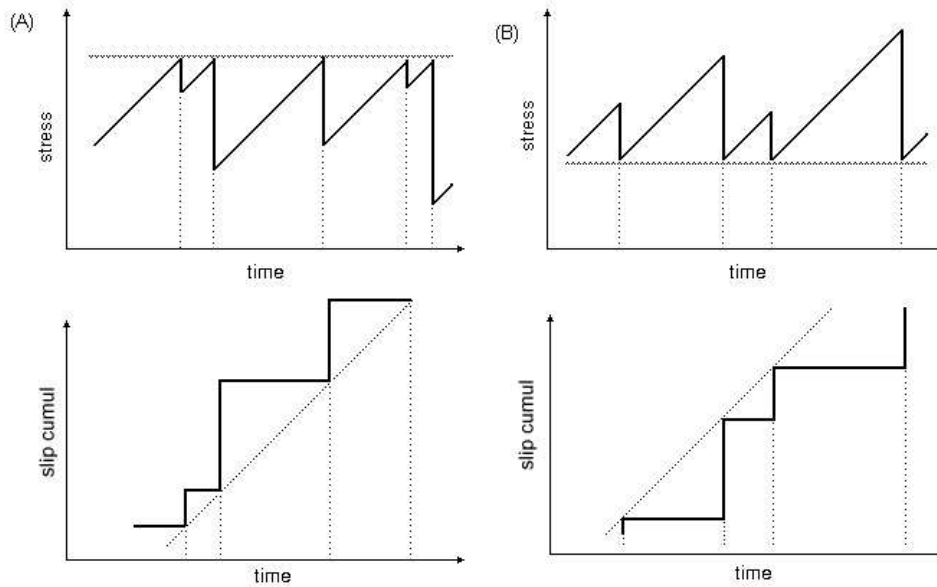


Figure 11: The slip-time behavior according to the (A) time- and (B) slip- predictable models (*Shimazaki and Nakata, 1980*). The two upper graphs show the non cumulative curves while the two lower ones show the cumulative curves.

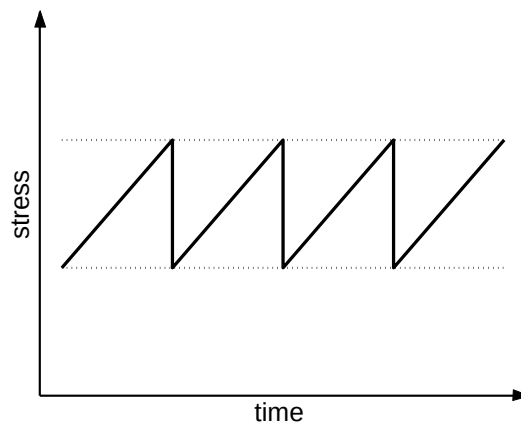


Figure 12: The stress vs. time cyclic and regular behavior according to the characteristic earthquake model.

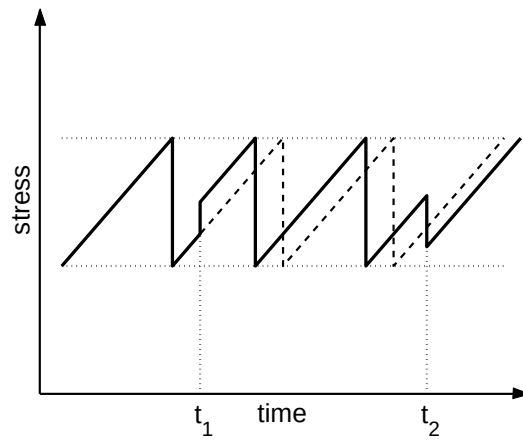


Figure 13: According to the CFS model, an earthquake occurring at time t_1 would take a hypothetical nearby fault closer, in time, to rupture. Another earthquake occurring at time t_2 would, instead, take the fault further in time from rupture. The characteristic model has also been reported to show time 'advance' and time 'delay'. Note that this is just speculative since it would imply a lower and upper threshold which have not been found to be compatible with the data (see text).

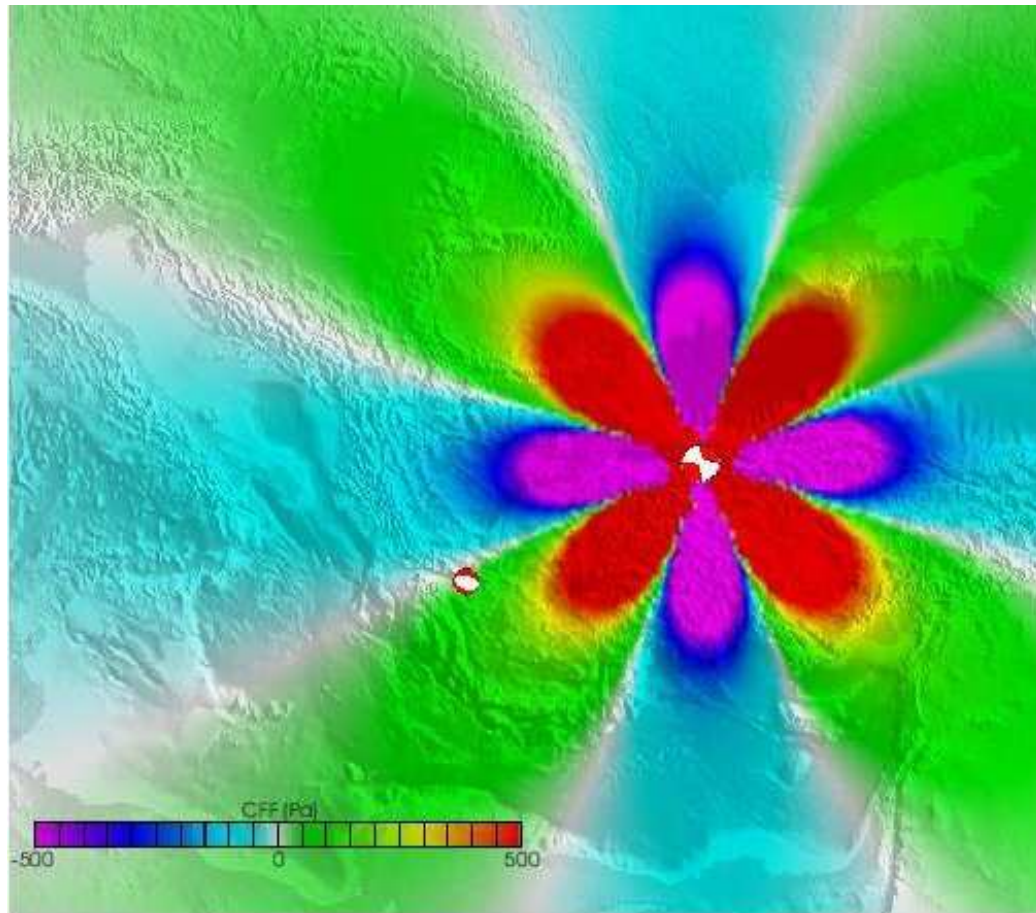


Figure 14: Map showing the calculated CFS change after the 1999 $m_w = 7.4$ Izmit (Turkey) earthquake. Figure after <http://www.ingv.it/>.

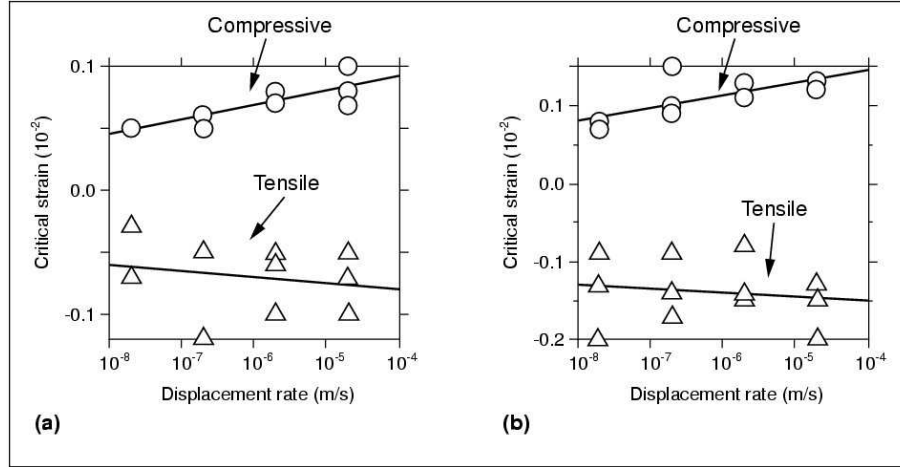


Figure 15: Critical strain in the axial (compressive) and radial (tensile) directions at the moment of dynamic failure as a function of strain rate, for (a) Inada Granite and (b) Noboribetsu welded tuff, after *Fujii et al. (1998)*. Dilatancy in the radial direction (the tensile component marked) decreases systematically with respect to strain rate.

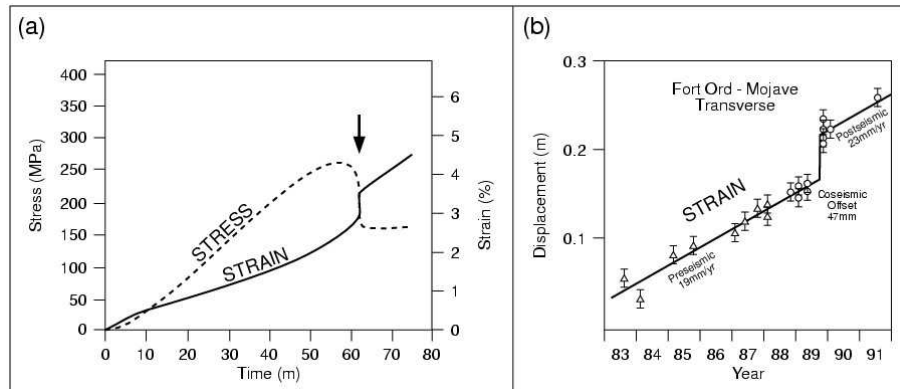


Figure 16: Comparison of (a) laboratory and (b) field measurements of precursory strain (solid lines). The accelerating strain seen in the laboratory sample (after *Main et al., 1993*) is not observable within the resolution of the geodetic method used to measure crustal strain around the time of the earthquake Loma Prieta, California, in 1989 by *Argus and Lyzenga (1994)*. It is only in the laboratory that we can directly monitor the stress, which may decrease as well as increase before dynamic failure (dashed line).

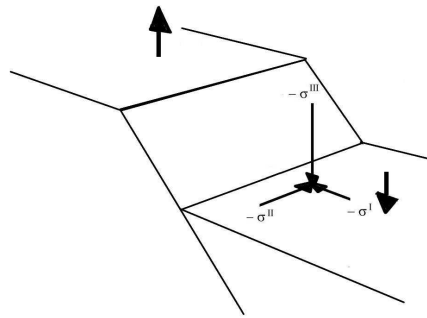


Figure 17: The principal stress values and axes for a normal fault.

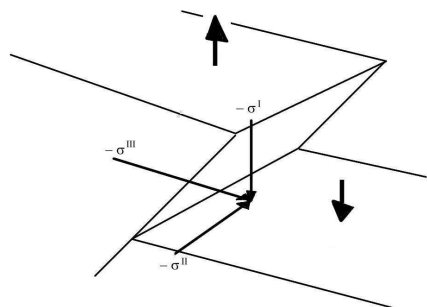


Figure 18: The principal stress values and axes for an inverse fault.

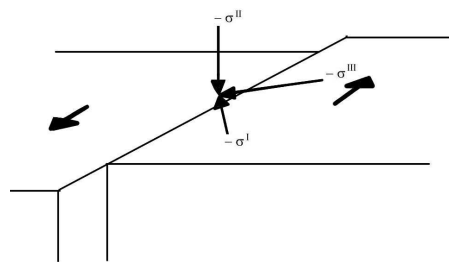


Figure 19: The principal stress values and axes for a strike-slip fault.

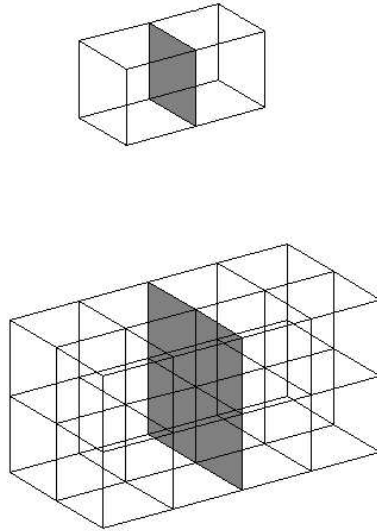


Figure 20: The self-similarity relation, with slip occurring at the contact face. Strain remains constant. The linear dimension of each cell is ~ 10 m.

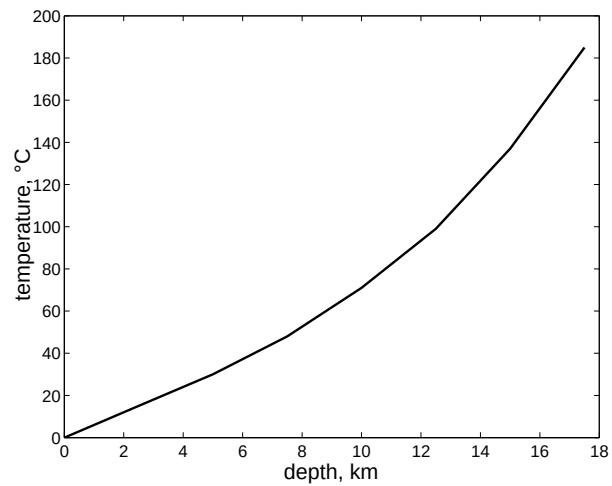


Figure 21: The temperature required to increase water pressure from hydrostatic to lithostatic (data are from *Burnham et al.*, 1969).