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Triggering of instabilities in materials and geosystems

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Abstract

Materials from the laboratory to geological scales respond to perturbations in a complex nonlinear fashion. In particular, the response to finite perturbations which cause failure (triggering) is an important, yet poorly understood, issue. Causes of failure may either be exogenous (precipitations, pore pressure, seismic waves, or in the materials context, mechanical perturbations) or endogenous (chemomechanical deterioration, creep deformation, microcracking and microplasticity).

1 Introduction

Materials from the laboratory to geological scales respond to perturbations in a nonlinear fashion, leading to a multi-scale problem of immense complexity. Relevant processes range from the scale of atoms where their arrangement and the ensuing defects such as dislocations and microcracks are of crucial importance for the deformation properties, up to the geological scale where deformation instabilities manifest themselves in the form of extreme, catastrophic events. This complexity makes it difficult to understand and forecast failure of materials and devices, and natural hazards such as rockfalls, landslides, snow avalanches and earthquakes. While the general problem of the occurrence of natural hazards is extremely wide and difficult to treat comprehensively, a very important question is to understand how instabilities in materials and geosystems are triggered by external perturbations and possibly to suggest new routes for prediction strategies. Triggering causes can be exogenous, such as precipitations, pore pressure, seismic waves in the geological context, or, for materials one can identify perturbations of mechanical, chemical and electromagnetic origin. In addition, instabilities can arise due to endogenous causes related to the internal relaxation of the system through creep deformation, internal fracture or plastic flow. An example of an internal perturbation acting as a triggering cause is present in wet snow avalanches where, due to surface warming and melt water production, percolating water is caught at capillary barriers leading to a loss of shear support. Due to the complexity of the percolation process acting in stratified snow cover, wet snow avalanches are notoriously difficult to predict. In general, one may expect that several triggering mechanisms act simultaneously, and it is thus necessary to address the interplay between different phenomena.

To understand the triggering of extreme events one has to deal with interlocked effects involving different scales in time and space. This makes it necessary to integrate methodologies from materials and earth sciences within the more general perspective of complexity.

2 Granular materials

2.1 Friction

Due to its importance in different fields, granular matter has been investigated for many years, but has recently attracted a renewed interest from the physics community. Here, we would like to summarize some frictional properties of granular matter. For a more detailed treatment, we refer the reader to some excellent review articles already published (Marone, 1998; de Gennes, 1999; Jaeger et al. 1996).

As first noticed by Janssen (Janssen, 1895), the pressure beneath a tall vertical granular column is not a linear function of the height, as is the case for a regular solid or liquid, but is instead a constant, independent of the height! This behavior is due to the complex contact forces between the grains, creating a macroscopic horizontal force component. Frictional forces at the container walls then transfer a fraction of the vertical force to the walls themselves, thus reducing the normal pressure beneath the column. Using simple assumptions for the normal and tangential stress components and basic friction laws, Janssen showed that the pressure has an exponential dependence on the height, saturating to a constant value (Liu et al. 1995; Janssen, 1895). Note that this important phenomenon is responsible for the catastrophic collapses occurring in grain silos when the pressure on the wall exceeds its strength.

Due to the disordered arrangement of the particles, contact forces in granular media follow intricate paths, giving rise to a network of "force chains" (Liu et al. 1995; Travers et al. 1987), often with fractal and unpredictable structures, with 'chaotic' dynamics. The structure of these networks has been revealed by photoelastic imagery with which it is possible to directly visualise stressed particles in the assembly. Spectacular images such as those taken by the group of Behringer in a two dimensional rotating cell have been published and the formation and breakup of force chains as the cell is sheared have also been observed (Miller et al. 1996; Howell et al. 1999; Veje et al. 1999). Transmission of forces through this irregular network is responsible for many peculiar features of granular mechanics (such as the independence on height of the normal pressure discussed above), but also for stress fluctuations and the tendency of the medium to dilate under shear (Reynolds, 1885; Bagnold, 1966).

Due to the relevance of this problem for engineers and geophysicists, the mechanical response of granular matter has been investigated for many years. The basic phenomenology is often described in terms of soil mechanics which has become a discipline in itself (Taylor, 1945;

Wood, 1990). An ideal shear test consists of a granular medium sheared by a deviatoric stress σ under a normal pressure p. The yield stress for granular matter can be expressed in analogy with frictional sliding as

$$\sigma = \mu p + C \qquad (1)$$

where m is the friction coefficient and C is a cohesion parameter. The response of the medium also depends on the density of the granular assembly: a dense medium will dilate under shear, while a loose one will compact. Experiments show that there is a critical density at which the granular medium flows steadily, leading to the notion of critical state. Though the critical state of soil mechanics is not necessarily related to critical phenomena in phase transitions, sheared granular media have in fact been observed to produce scaling laws. Taylor (1945) has proposed a simple argument to describe dilation of a granular medium. The idea is that, in order to be sheared, the medium must also expand to overcome the constraints of interlocked grains. Under a shear stress σ and pressure p the medium deforms by dx and dy in the horizontal and vertical directions, respectively. One can then equate the mechanical work with by the frictional work obtaining

$$\sigma A dx - pA dy = \mu p dx \qquad (2)$$

where A is the sample area. From Eq. (2), we obtain a relation between the yield stress σ_c, the dilatancy angle ψ, defined by $\tan\psi \equiv dy/dx$, and the friction angle θ, defined by $\tan\theta \equiv \mu$:

$$\sigma_c = p(\tan\theta + \tan\psi) \qquad (3)$$

This equation can also be expressed as a relation between friction coefficients and the dilatancy as (Marone, 1998)

$$\mu_c = \mu + \tan\psi \qquad (4)$$

where μ_c is the measured friction coefficient of the granular medium. The soil mechanics description of granular shear is usually more complicated than the one we have sketched here. Granular friction is a very complex phenomenon that is difficult to capture entirely with simple phenomenological laws and a vast literature has been devoted to this topic.

2.2 Triggering and Failure

The conditions under which a granular material can fail are poorly understood by the scientific community (Tordesillas et al. 2004). In the case of blocked production lines or silo outflows, failure of any spontaneously formed granular structure is desirable; but in earthquakes, failure of the interlocked granulate constitutes one of the most feared of all natural hazards. Even a small improvement in our understanding of granular materials could have profound socio-economic implications if earthquake prediction or mitigation were to become possible or if commercial granular materials could be handled more efficiently.

The problem with understanding granular materials originates at more than one level - the interactions between individual particles are not simple, often discontinuous, whereas at the macroscopic level, the force-chains mentioned render analysis intractable. Furthermore, external conditions can make a granular medium behave either as a solid, a liquid or a gas, or even any combination of the three simultaneously! Apparently simple properties such as particle surface roughness or adhesivity can greatly change the resulting structures and the discontinuous interactions imply that there is a yield point at which a seemingly solid granular structure will abruptly fail. Even worse, the failure is not ductile in the sense of a slowly increasing deformation with applied load, but more brittle in nature - Young's and shear moduli change discontinuously (and unpredictably) while the granular material reorganises into a new structure with which to bear the new applied load.

Naturally, one of the most fascinating applications of this research area is geology and earthquakes. In general an earthquake zone consists of a 2.x-dimensional fault zone between two tectonic plates. If, at any point, the local stress exceeds some threshold, a fracture will nucleate and propagate throughout the fault zone. The fracture process itself consists of a shear motion between two tectonic plates and the propagation of seismic waves throughout the surrounding crust; these seismic waves can actually remotely trigger the fracture of another fault. Johnson and Jia (2005) have suggested that the triggering mechanism at work is strain-related weakening of the gouge material and is applicable to both dry and wet fault zones.

Research has demonstrated that granular materials approaching failure tend to become "fragile" (Cates et al., 1998), that is, to be divergently susceptible to small changes in the load-bearing structure, or in the load borne thereby. In order to proceed to a greater understanding of failure and triggering in granular materials, then, it will be necessary to excite granular materials with diverse mechanisms and under different conditions, and in this brief section we will summarise some recent work in this area.

2.2.1 Weakening of contacts

the indroduction of an interstitial fluid has been found to weaken the granular contacts (Morgues, 2007). This arises because the fluid carries some of the stress and so reduces the normal force at contact points, leading to a reduction in the yield shear force of that contact.

2.2.2 Stress increase beyond threshold

The principal failure mechanism of earthquakes is the slow inexorable motion of tectonic plates. This global stress increase eventually leads to local overstressing causing failure.

Savage and Marone (2007) have also expermientally investigated the effect of 'seismic' waves on a slowly stressed granular material and found that the stress oscillations tend to trigger an event primarily in faults which are already stressed close to failure. However, different surface types tend to be susceptible to different vibration types: stressed rigid surfaces can be triggered to fail by low-frequency waves, while granular materials tend to fail on application of high-frequency waves, presumably due to a degree of liquefaction of the granulate.

2.2.3 Local reorganisation of structure [creep], causing weakening

External mechanical excitation (e.g. vibration) can cause some grains to 'knock' off surrounding grains. This could potentially cause a load-bearing grain to slightly shift position, leading to a change in the structure, which may or may not cause macroscopic yielding.

An externally applied perturbation has been shown to induce gravitational failure (landslide) - Staron et al. (2006) have shown that the approaching stability limit is heralded by an increase in precursor probability and a sharp change in the microstructure. Daerr and Douady (1999) also found that the threshold perturbation for triggering vanishes as the slope reaches a critical value.

2.2.4 Weakening of the percolating structure

Lade (2002) indicates that the strength of a granular material is primarily proportional to the external confining pressure. As the pressure is reduces, the entire structure becomes proportionally weaker and so failure may be triggered.

2.3 Shearing

Perhaps the simplest situation of granular shear is when the material is subject to a horizontal shear stress. As mentioned above there is a critical value above which the medium starts to flow, but due to the disordered and non self-averaging nature of the medium, this value is not strictly constant. Moreover, a dynamic instability is often present, so that the flow can momentarily stop and the shear takes the character of intermittent motion. This kind of response to the applied shear can be either periodic or highly intermittent and irregular. This latter behaviour has been observed in many phenomena at very different length-scales, such as plastic deformation and seismicity (Sethna et al. 2001).

A quantitative description of the stick-slip phase has perennially constituted a great challenge for researchers. The deterministic model proposed by Burridge and Knopoff (1967) (and many successive modifications (Carlson et al. 1994)) were the first attempt to reproduce the chaotic response of a system to an external shear, pointing out that friction and elasticity are the two essential ingredients, as well as the presence of many correlated degrees of freedom. Devised with the aim of mimicking the main features of earthquakes, the different models succeeded in reproducing, to various extents, scale-free features of this phenomenon, like the Gutenberg-Richter and Omori laws.

With the aim of investigating the stick-slip response of a dry granular medium in a quantitative way, laboratory experiments have been set up in recent years. Nasuno and co-workers investigated the shear response of a horizontal granular medium when driven by a top plate via a spring connected to a motor (Nasuno et al. 1997; Nasuno et al. 1998). It was observed that the response ceases to be (almost) periodic for stiff springs or low driving velocity. Interestingly, the same property was observed in a similar experiment for solid-on-solid friction (Johansen et al. 1993). Periodic response was also observed in a vertically pushed granular columns (Kolb et al. 1999).

In many other situations, however, the shear response was erratic: Dalton and Corcoran (2001, 2002) were the first to observe experimentally that there are scale-free distributions in this regime. They showed in particular that there exists a critical state in which duration, size and energy of each slip are all distributed according to power laws. On the other hand, when the system is sub-critical or supercritical, an upper cut-off to the distributions appears. The experimental set-up they employed consisted in an annular channel filled with grains to which a shear stress was applied by a plate placed on the top. This was kept at fixed height and put into rotation by a spring connected to a motor rotating at constant angular speed.

In more recent work, using a set-up very similar to the one described above but with the plate free to displace vertically to allow the dilation of the medium, Dalton et al. (2005) have then shown that the shear stress in the stick-slip phase is not Gaussian. And that this is true for any kind of stress one considers, i.e. yield, dynamic, or at any moment. This mark is an additional and very strong evidence that the stick slip phase is completely different in nature from the Bagnold phase, in which there is continuous sliding, where Gaussian fluctuations are found. They also observed that the average stress is not independent of the velocity, but slowly decreases down to an absolute minimum, before increasing again with some characteristic viscosity. This feature, at the basis of stick-slip motion, and common to other friction situations, e.g. solid on solid, had been predicted on the basis of the repose angle of granular piles (Jaeger et al. 1990) but had never been experimentally measured before.

The gap between simple characterization of the stick slip and an effective dynamic description has only been filled very recently. In fact, an equation of motion of the top shearing plate has been successfully derived by Baldassarri et al. (2006) starting from general considerations. The equation is stochastic, in force of the nature of the motion. This is taken into account by describing the reaction force as the sum of a deterministic frictional part (dependent only on the velocity), and a fluctuating part. The first part has been experimentally determined while the latter is assumed to perform a bounded random walk since the reaction force is determined by the internal rearrangement of the medium's force chains. The equation has proved to be effective in describing the stick slip motion quantitatively and reproducing the distributions of slip size and duration, as well as of velocity and stress.

3 Complex materials and time-dependent fracture

3.1 Complexity and fracture

The fracture mechanics of materials starts by considering the behavior of the average strength. The apparent strength as tested in the laboratory turns however out immediately to present surprises. The two first ones are that strength as a quantity exhibits a size-effect, bigger specimens are weaker, and that measuring a set of apparently similar samples leads to a non-trivial probability distribution for the values derived from the experiments. Of course, these two phenomena are coupled.

The classical Linear Elastic Fracture Mechanics (LEFM) way to approach these issues starts by considering a single flaw of size a in an otherwise homogeneous material. The same result can also be derived by the famous Griffith's scaling argument. These relate the apparent strength to "material parameters". It is instructive to consider an extended version which reads as

$$\sigma_{app} = K_c / \sqrt{\beta(a+\xi)} \,. \tag{5}$$

Here, the sample average strength is expressed via a stress-intensity factor (SIF)

$$K_c = \sqrt{G_c E} \tag{6}$$

and geometric quantities a and ξ, while β is a geometric factor. The SIF includes the elastic modulus E and the fracture toughness G_c. The important modification in (5) vis-à-vis usual LEFM is the "crack extension" or Fracture Process Zone size ξ. Its addition ensures that the scaling predicted by Eq. (5) extends also to cases where the size-effect is ruled by the "intrinsic strength" of the material, not a test flaw of size a. The presence of the FPZ is already in practice an indication of a "complex material", as it tells us that the internal processes in the sample lead to the corrective effect of the FPZ. In the Random Fuse Model, a scalar analogy of LEFM, one can show by simulations that the ξ is in fact related to the scale of an exponentially decaying damage cloud at the tips of flaw (Alava et al. 2007). The form of Eq. (5) is such that is equivalent to Dugdale-Irwin -style correction to LEFM for materials with FPZ plasticity. The SIF includes two quantities, of which again the fracture toughness is a highly non-trivial one. The Griffith's argument simply assumes it to be a parameter without paying attention to what kind of physics is important in determining the value of G_c. The most straightforward interpretation is that it is given by the surface energy needed to create new crack surface. This leaves open complications that may be of importance in certain materials (see e.g. Kierfeld and Vinokur, 2006). For instance, the crack surface can be rough, and arguments have been given that for self-affine surfaces with sample-size dependent anomalous scaling one should observe a "R-curve" or crack-length dependent fracture toughness (Morel et al. 2002).

The application of fracture mechanics to complex materials has lead in engineering applications to the re-invention of Equation (5) in several independent cases. Bazant has reviewed in a number of publications such applications to quasi-brittle materials, which can be translated to such where plastic deformation is not an essential ingredient in understanding fracture, and which have rich internal structure so as to induce a FPZ (Bazant and Planas 1997, Bazant 1999, Bazant 2004). A generic playground for these ideas is composites. Examples are concrete (Chiaia et al. 1998, Lilliu et al. 2003, Van Vliet and Van Mier 2000, Van Mier and Van Vliet 2003) and ordinary paper (Kettunen and Niskanen 2000). These consist of several phases (mortar and rock in concrete, fibers in paper, plus voids). Looking at the extended LEFM tells now that there are three important issues to understand, for a material scientist. Whether by theory, experiment, or both one needs to comprehend the value of the fracture toughness. In a fiber-matrix composite for instance, this could result from complicated microscopic processes involving the fibers, their interface, the matrix, or some combination thereof. Then, there is the FPZ size, and its origin. In quasi-brittle materials the right concept here would be Damage Mechanics, to account for the irreversible changes (microcracks) that accumulate as the sample is pushed close to its actual strength. Both the FPZ scaling and the fracture toughness are dependent on the presence and strength of disorder. Finally, in the limit that the flaw size a is let go to zero one has the "pure" strength of the material. This has however also a size-effect, a dependence on the sample size L, not revealed by Eq. (5). For the case of non-interacting flaws in independent sub-volumes, the possible outcomes for the statistical behavior are dictated by the so-called statistics of extremes (Gumbel 2004). Thus the Weibull or Gumbel distributions are indicated as limiting distributions depending on the coarse-graining of the microscopic strength (Duxbury 1986, Korteoja et al. 1998, Alava et al. 2006).

The discussion illustrates that even the simplest framework for fracture hides many deep issues about the physics of deformation. The FPZ scale in Eq. (5) and the fracture toughness summarize a history- or loading protocol dependence. In materials where damage occurs and is accumulated, its state will tell about the previous history the sample at hand has gone through. However, in the framework of extended LEFM including the size effect for specimens without macroscopic flaws there is nevertheless little predictability based on the current state of the system. In both limits the failure is abrupt, and is akin to first-order transitions in statistical physics. This also then implies that the triggering of failure is roughly understood since the essential physics is encoded in the average strength given L and the sample-to-sample variations around that: does the instantaneous stress exceed the actual critical one for the sample?

The unpredictability of failure is of course a boring claim, and moreover real materials present time- and history dependent response that complicates the issues at hand. The signatures of fracture processes ranging from geological scales to the laboratory show features of collective dynamics that is difficult to relate to classical fracture, where the physics is summarized via scaling formulas and the input quantities. An example is the “crackling noise” (Sethna et al. 2001) measured as acoustic emission, which indicates that the development during a loading history leading to final failure is not smooth and laminar, but “jerky” and with no typical scale. Therefore, the possibility of non-trivial triggering physics and understanding predictability reopens up. A distinct possibility for such is if the failure dynamics is governed by “mean-field” interactions. This means, that on mesoscopic scales the variations in local material response and the singularities in stress-fields close to microcracks may be neglected. What matters is the generic stress state, as induced for instance by the damage and the concomitant increase in the local stresses due to the decrease of the load-bearing parts. The most important statistical physics concept might be then the life-time t_c of the sample, and it is pertinent to recall that in the scenarios discussed above this is trivial to compute.

3.2 Time-dependent failure

The essential tool to comprehend various mechanisms that could play a role, or turn out to be interesting for the physics that ensues is the “fiber bundle model” (FBM). This is a simplified tool for including varying types of interactions – short-range, long-range etc. – and various dynamics for the failure. The stress per fiber, σ(x), is redistributed as failures of fibers take place and the rule for this is one of the crucial ingredients. The elastic-brittle FBM has a phase diagram with a mean-field –like, “Global Load Sharing” variant and a phase with local stress enhancements around cracks with “Local” rules. We discuss here the GLS version as a simplifed paradigm of complex fracture mechanisms.

The GLS fiber bundle is usually studied with random failure thresholds from a distribution $P(\sigma_c)$. At a given external force, the stress per fiber is $\rho = F(t)/N(t)$ where the denominator counts the intact fibers. The model indicates a behavior similar of a 2nd order phase transition, with avalanches at fixed force values due to redistributions of stress as fibers fail. The integrated avalanche size distribution is $P(N) \propto N^{-5/2}$ and the failure rate exhibits a divergence as $dN/dt \propto (F_c - F)^{-1/2}$ (with reasonable failure distributions P). The GLS model can be solved via various techniques eg. relating the increase of σ to a random walk with a drift. Another illustrative way of analysis is a graphical solution (da Silveira 1998) by looking at the right-most intersections of the curves $y = F(t)x/N_0$ and $y = 1 - P(1/x)$ where $x = N_i / F(t)$ and one considers a discrete system (in terms of the number of fibers).

The scenarios which one might now apply to such methodology (or models) are plentiful. The loading scenarios range from creep (constant load) to fatigue (cyclic loads) to tensile tests.

3.2.1 Viscoelstic elements

These elements exhibit some typical rheology as $\sigma = E\varepsilon + b\dot{\varepsilon}$ (Hidalgo et al. 2002, Kun et al. 2003a, Kun et al. 2003b, Nechad et al. 2005, Baxevanis and Katsaounis 2007). Other possibility is to say that $\dot{\varepsilon}$ follows an Eyring-type rheology. In this kind of models the idea is often most such that the stress is redistributed as in the elastic case. Since the dynamics of ε is the crucial quantity, usually one introduces a distribution for the failure strains. Studies of creep in this class of models reproduce e.g. an "Andrade type" primary creep with a $\dot{\varepsilon} \propto 1/t$ MF-like behavior, and it is very easy to construct scenarios where an infinite lifetime at small loads is at a critical threshold replaced with a finite t_c, with a 2nd order divergence. Qualitative comparisons have been made with creep data from composites and paper, with mixed results (Nechad et al. 2005).

3.2.2 Direct involvement of a temperature

FBMs with and without disorder have been studied with the idea that the failure probability of a fiber depends on the stress and T via an Arrhenius–like manner (Roux 2000, Peliti et al. 2002). The introduction of disorder, of varying failure thresholds, of course changes the energy landscape and enhances the effective temperature (Ciliberto 2001, Scorretti et al. 2001). It is of interest to note that one-dimensional local models describing the local growth of a crack lead to the opposite result, since the higher-than-average barriers of course are important (Santucci et al. 2003). The main interest here is that models that have an explicit T can be contrasted to t_c-results from creep-like experiments. Due to the mean-field character of the GLS model, it is not clear if there is any interesting collective behaviour. The opposite limit, of a crack tip moving in a landscape, is very different in that no collective dynamics is assumed in side an elementary excitation volume (Santucci et al. 2004, see also Kierfeld and Vinokur 2006). The problem here is that one does not know if the geometry is in fact zero-dimensional, as is highlighted in 2d experiments by crack branching. A further example is to use a local temperature – originally this was done for the random fuse model, where current enhancements lead to some quantitative variations (Sornette and Vanneste 1992) – which follows a differential equation including a source term from the local stress, and a linear dissipation. Failure is induced locally by exceeding the critical temperature.

3.2.3 Accumulated damage

Clearly one can include more complicated history dependencies to mimic possibly more realistically damage dynamics in Nature. There are three ways at least of including these. First of all, one can use a local "hazard function" that tells the likelihood per fuse to fail per an instant of time, such that this depends on the stress and possibly a local random threshold (Moreno et al. 2001, Nanjo and Turcotte 2005, Shcherbakov et al. 2005, Turcotte and Shcherbakov 2006). This is in essence very similar to including a temperature, but the main point is that as in the GLS model, a critical point is included by design. The other choices are such that the damage variable can be taken to be an "internal one" or an external one, involving a decrease of the fiber elastic modulus and a redistribution of stress, in which case the model as such is rather closbe to the ordinary FBM (e.g. Kun et al. 2000). The growth rule of a damage variable is easily chosen to be governed by a differential equation with the instantaneous change a function of the current stress $\sigma(t)$ and in fact, possibly augmented by a finite memory, the resulting dynamics are similar to what one has in the local temperature model discussed above. Kun and co-workers have demonstrated that the combination of local random damage thresholds and such dynamics on one hand and local random stress thresholds reproduces the empirical Basquin law of fatigue fracture, stating that the number of stress cycles to failure is a power-law function of the maximum stress, $N_f \propto \sigma_{\max}{}^{-\alpha}$, where α is the Basquin law exponent (Kun et al. 2007, see Alava 2007).

Above, we have listed a number of ways of including nonlinearities and complexity into the FBM with the hope of repeating experimental fracture signatures. Next, we turn into the experimental situation.

3.3 Experimental state of the art and outlook

The statistical physics viewpoint about time-dependent fracture would stress the universal aspects of the phenomena. The two central questions that one would like to understand are the following. First, the long-range stress fields in fracture problems – whether due to stress enhancements or global mean-field load sharing – and the disorder present make the response of sample in a fracture experiment in general very similar to many other scenarios in statistical mechanics, to start with the Ising model. Understanding the limitations and strength of this analogy is of great importance. Second, the failure of a specimen in a test – or of a model system - might look like a "phase transition" as various kinds of models highlight. The existence of a critical point is often equated with a t_c, and the approach by tuning a control parameter (stress) is considered to be akin to approaching a 2nd order phase transition. However, as noted above models with stress enhancements often imply an abrupt failure more like what happens in the proximity of 1st order transitions (Zapperi et al. 1998). It is an open question whether there are scenarios where the latter case is combined with precursors such that the triggering and predictability of final fracture become interesting.

The experimental signatures that are of interest here can be divided into three separate classes. We have acoustic emission from many kinds of materials implying that the fracture development is jerky, and such that the AE event distributions are characterized by power-laws akin to the Gutenberg-Richter law, $P(E) \propto E^{-\mathrm{B}}$ (Petri et al. 1994, Guarino et al. 1998, Maes et al. 1998, Salminen et al. 2002, Deschanel et al. 2006, Davidsen et al. 2007). The same holds also for waiting times t_w, which can be characterized by Omori's law-like power-law distributions. The existence of such scalefree distributions is not sensitive to dimension, or loading mode or material. The interesting points are twofold: first, there is very little agreement with the measured power-law exponents and models (for "b-value analysis" see (Amitrano 2003, Amitrano and Helmstetter 2006). Second, it is very hard to come up with an experiment without such scalefree scaling. So far there are not many studies of the detailed correlations in AE time series. One problem is that fracture experiments are typically not stationary (creep, fatigue, tensile: all share this problem, but see Salminen et al. 2006, and Koivisto et al. 2007). The situation in laboratory materials is thus different from geophysical systems, where this is often assumed.

Second, the time-dependence of fracture is often measured by t_c or the rheology, of the type $\dot{\varepsilon} = f(\sigma)$, that leads to the breakdown. This dependence together with the divergence often results as a coarse-grained behavior from mean-field-like models with damage mechanics (Main 2000, Lyakhovsky et al. 2005, Nanjo 2005, Kawada and Nagahama 2006). In a tensile test one can interpret the integrated AE activity as a measure of the damage, and the concomitant release of elastic energy (Lockner et al. 1991, Colombo et al. 2003). A divergence of the release rate might then be interpreted as an approach to a critical point (Guarino et al. 1998). In practice, it is hard to come up with credible evidence over several magnitudes of energy. The creep life-time is an interesting quantity, in particular as a function of load (Guarino et al. 1999, Guarino et al. 2002, Nechad et al. 2005). In creep and fatigue (see above) tests, there are the usual customary empirical classifications of the material behavior. The creep response is often divided roughly into two different "universality classes". We have the classical division of a creep history into three parts: primary (power-law Andrade) creep, secondary (logarithmic) and tertiary with an acceleration of the strain-rate when final failure is approached. Here, it is very unclear as to which parts of the physics are coupled to fracture and which to viscoplastic response and the standard materials science characterization is the so-called master curve formalism (Penny and Marriott 1995) AE measurements show similarities to ordinary tensile failure, while the primary creep in particular may result from plastic deformations. An alternative scenario is rheological response of the Dorn equation type, $\dot{\varepsilon} = \sigma^n$. This kind of creep can be derived from FBMs where the damage mechanics has been tuned. The role of the ambient temperature is in these scenarios of primary importance, since activation-type dynamics is assumed in theory, but unfortunately has not been tested much in practical experiment. For instance, in the fracture of notched samples one can try to interpret the role of the temperature by assuming that it controls the growth of a crack upto a critical size, independent of T. Such assumptions (that the failure strain is a material-dependent constant) are often made in interpreting e.g. creep or fatigue fracture.

Finally, a much less used but in particular for the future promising approach is the visualization of the fracture process by various techniques. These range from X-ray tomography of 3D samples to AE event localization to Digital Image Correlation analysis (Kettunen and Niskanen 2000, Roux 2006, Wang et al. 2003). The main idea is to provide with a spatiotemporal picture of the material response, which already qualitatively should allow to exclude certain types of models. The information one may hope for concerns the causal relations among events – in the case they can be well separated as entities, like oftenmost in an AE timeseries – and their spatiotemporal characterization (localized vs. non-localized, affine vs. self-affine). Such data can be compared with the final failure to see if it can be noticed or predicted, where catastrophic crack propagation is nucleated, and likewise one can analyze the possible localization of damage and/or deformation (Lockner et al. 1991, Krysac and Maynard 1998). It is also an important prospect to combine such advanced analysis techniques with direct experiments on triggering.

4 Strain softening and material instabilities

To develop a generic conceptual framework for categorizing material instabilities in materials and geosystems, we refer to a the evolution of a material system along a given 'deformation path' which we parametrize by a generalized displacement coordinate s. This may be the sliding path in a friction experiment, the axial strain in a tensile test, or the interface shear strain in shear testing of a weak interface. In general the variables characterizing the deformation state of a material system will be vectorial or tensorial quantities; however, if the system is constrained to evolve along a given path, s may be considered the scalar projection of these quantities. For simplicity of notation, we will assume this to be true in the following.

The generalized force required for driving the evolution is denoted as f . In the examples mentioned above this may represent the tangential force in frictional sliding, the tensile stress, or the interfacial shear stress. Again these quantities may be vectorial or tensorial in nature, but F is envisaged as their scalar projection on a prescribed deformation path.

The mechanical properties of the material are, in the simplest case specified by a constitutive law connecting the required driving force f , the displacement s, and the rate of displacement $\dot{s}$ as well as external control parameters c^i :

$$f = f\left(s, \dot{s}, \{c^i\}\right). \tag{7}$$

Or, in differential form (see e.g. Estrin and Kubin, 1988; Estrin and Kubin, 1991):

$$df = \theta \mathrm{d}s + m \mathrm{d}\dot{s}, \tag{8}$$

where the generalized hardening coefficient (or tangent modulus) $\theta = \partial f / \partial s|_{\dot{s}}$ of the material and the generalized rate sensitivity $m = \partial f / \partial s|_{\dot{s}}$ are functions of s and $\dot{s}$.

We now have to specify what we mean with the term 'instability'. Assume the system is driven by a constant force $f_{ext} = f\left(s, \dot{s}, \left\{C^i\right\}\right)$. As material instability we then define any situation where a spontaneous acceleration of the deformation process may occur. For constant force $\mathrm{d}f = 0$, hence

$$\partial_t \mathrm{d}s = \frac{\theta}{m}\mathrm{d}s \quad . \tag{9}$$

Instability thus occurs, in this scenario, whenever the ratio θ/m becomes negative. From this criterion, two basic types of plastic instability may be distinguished, namely deformation softening instability occurring for $\theta < 0$, $m > 0$ and rate softening instabilities occurring for $\theta > 0$ and $m < 0$ (Estrin and Kubin, 1988; Estrin and Kubin, 1991; Zaiser and Hähner, 1997).

We note that, by requiring the driving force and deformation path to be fixed, we have excluded a third type of instability which is associated with changes in the deformation geometry or deformation path. Such geometrical instabilities may lead to an increase in driving force even though the external action on the system does not change. A classical example is the Considere instability in stretching of a long thin tensile specimen: Since the driving force f (the tensile stress) is related to the externally applied tensile force by $f = f_{ext}/A$ where A is the specimen cross section, cross section reduction may increase the driving force in spite of a constant applied force. For volume conserving deformation, the axial elongation s and cross section reduction are related by $ds/s = -dA/A$ and, hence $df = (f/s)ds$, leading for $m > 0$ to the instability criterion $\theta < f/s$: at this point, local increases in the degree of deformation lead to an increase in the local deformation rate despite a constant external driving force, and thus necking may occur. Finally we note that geometrical instabilities may be associated with spontaneous changes in the deformation path of an insufficiently constrained system, the classical example being the Euler buckling of a column undergoing axial compression.

Strain softening instabilities are associated with an irreversible decrease of the deformation resistance after the onset of deformation. On the microstructural level, this corresponds to the destruction of material bonds between the parts of the system, or the elimination of obstacles in course of the deformation process. Examples include:

- The irreversible displacement softening of a weak interface in a snow stratification which precedes the release of a snow slab avalanche McClung (1979). The displacement coordinate *s* corresponds here to the relative displacement of both sides of the interface, and *f* is the shear stress acting at the interface. Softening is associated with the irreversible breaking of bonds between snow grains as s increases.

- The activation of dislocation sources in plastically deforming crystals after yield. After the expanding dislocation loops have overcome a critical saddle-point configuration (the stress value required to overcome this configuration is called the Orowan stress, see e.. Hirth and Lothe (1982), they may expand at an ever decreasing stress, leading to strain softening and a flow stress decrease after the onset of plastic flow.
- The irreversible destruction of obstacles along the glide path of dislocations. Examples include the cutting of precipitates or point defect agglomerates, or the destruction of short-range order in the wake of a dislocation sweeping across its glide plane.
- The breaking of asperities at the onset of sliding in dry friction.

Strain-rate softening, on the other hand, is usually governed by the interplay between a strain softening process and a second process that restores the strength. Examples include:

- Thermomechanical softening: As thermal fluctuations may help to overcome deformation barriers, the strength of most materials decreases with increasing temperature. At the same time, work done during deformation is converted into heat, thus leading to a temperature increase. This softening mechanism is limited by heat losses to the environment. This may lead to a strain-rate softening effect as the steady-state temperature increases with deformation rate (Zaiser and Hähner 1997, Zaiser 1997).
- Wet friction of ice: heat dissipated during frictional sliding of ice may lead to surface melting and the formation of a lubrication layer. This process, which decreases the

friction force, is counteracted by heat conduction away from the surface. At small sliding rates, heat conduction prevails and lubrication is poor, whereas at high sliding rates, lubrication may improve, which leads to a decreasing sliding force (Evans 1976).

- In dilute alloys, dislocations attract diffusing solute atoms. At low strain rates/dislocation velocities, the point defect concentration around the dislocations is high and as a consequence the dislocations are strongly pinned. At higher velocities, the dislocations outrun the diffusing solutes and as a consequence the point defect concentration around the dislocations decreases. This leads to a flow stress which in an intermediate strain rate regime decreases with deformation rate (Portevin-LeChatelier effect, Estrin and Kubin 1988, Zaiser and Hähner 1997, Zaiser and Hähner 1997).

In the simple theoretical model outlined above, strain and strain-rate softening appear as dual phenomena: One can have strain softening, or strain-rate softening, but not both at the same time –in which case the model predicts no instability to occur at all. The physical examples given above, on the other hand, suggest a more subtle relationship between the two classes of phenomena. In fact, if we complement our 'strain softening' mechanisms with a time-dependent mechanism that restores the strength, we end up with a strain-rate softening scenario and vice versa. For example, in the first example of a strain softening interface in a snow stratification, rapid sintering may lead to a partial restauration of bonds if the sliding rate $\dot{s}$ is low (Louchet 2001). In this case, a higher sliding rate implies less time for sintering, thus the density of bonds decreases with increasing sliding rate, and we are in a rate softening regime. Conversely, in alloys exhibiting the Portevin-le Chatelier effect,dislocations may be saturated by solutes after a heat treatment prior to deformation ('aging'). If we then deform rapidly, the initial flow stress will be high but, after the onset of deformation, dislocations will break free from their solute clouds who will never be able to catch up again. As a result, we will see an initial strain softening at yield and then stable plastic deformation (Zaiser at al., 1999).

4.1 Generic constitutive model with an internal variable

The above observations lead us to consider strain or strain rate softening in terms of the evolution of an additional internal variable φ in addition to stress and strain. This variable may characterize the density of bonds in case of the snow interface, the degree of lubrication in case of wet friction of ice, the characteristic concentration of solutes around dislocation cores in case of the Portevin-le Chatelier effect, or simply temperature in case of thermomechanical softening.

The counterpart of Eq. (8) now reads

$$\mathrm{d}f = \theta \mathrm{d}s + \mathrm{m}_0 \mathrm{d}\dot{s} + \frac{\partial f}{\partial \psi}\mathrm{d}\phi \qquad (10)$$

where m_0 is the instantaneous rate sensitivity which governs the strain-rate dependence of f at constant ϕ. m_0 represents the intrinsic viscosity of the system and is generally positive.

The evolution of the variable φ is governed by two competing processes, one of which is driven by straining:

$$\dot{\phi} = g(\phi)\dot{s} + \mathrm{h}(\phi) \qquad (11)$$

Under stationary conditions at deformation rate $\dot{s}_0$, the steady-state solution is given by a stable root ϕ_0 of the equation $g\dot{s}+\mathrm{h}=0$. We now study the response of the modified model to a small step-wise strain-rate change $\Delta\dot{s}$. This leads to an instantaneous strain rate change by $\Delta f_0 = m\delta\dot{s}$. In the following, ϕ that relaxes towards a new steady-state value, leading to a flow stress transient that is given by

$$\Delta f\left(t\right) = \dot{s}\left[m_0 + \left(m_\infty - m_0\right)\left[1 - \exp\left(-\frac{t}{t_\phi}\right)\right]\right] \qquad (12)$$

The relaxation time t_ϕ is given by $t_\phi = \left[g'(\phi_0)\cot s_0 + h'(\phi_0)\right]^{-1}$ where the prime denotes derivatives with respect to φ, and the asymptotic strain-rate sensitivity m_∞ is given by

$$m_\infty = m_0 - \frac{\partial f}{\partial \phi}\frac{g(0)}{t_\phi} \tag{13}$$

m_∞ governs the change in strain rate after the internal variable ϕ has reached its new steady state value. Strain-rate softening now corresponds to $m_\infty < 0$, i.e. it occurs if the response of the internal variable ϕ reduces the required driving force sufficiently to overcompensate the effect of the intrinsic viscosity of the system.

Next we investigate the evolution of a small perturbation of the steady flow at constant f. We get the system of equations

$$\partial_t \mathrm{d}s = \frac{\theta}{m}\mathrm{d}s + \frac{\partial f}{\partial \psi}\mathrm{d}\phi \tag{14}$$

$$\partial_t \mathrm{d}\phi = g(0)\partial_t \mathrm{d}s - \frac{1}{t_\phi}\mathrm{d}\phi \,. \tag{15}$$

This can be rewritten as

$$\frac{\partial}{\partial t}\begin{bmatrix}\Delta s \\ \Delta\phi\end{bmatrix} = \begin{bmatrix} -1/t_s & -C_{s\phi}/t_s \\ 1/\left(t_{s\phi}C_{s\phi}\right) & 1/t_{s\phi} - 1/t_\phi \end{bmatrix}\begin{bmatrix}\Delta s \\ \Delta\phi\end{bmatrix} \tag{16}$$

As can be seen, the stability of the system is governed by the interplay of three characteristic time scales: t_ϕ and

$$t_s = \frac{m_0}{\theta}, \quad t_{s\phi} = t_\phi \frac{m_0}{m_0 - m_\infty} \tag{17}$$

t_s is the characteristic relaxation time of the flow rate due to strain hardening, while $t_{s\phi}$ is the time required for the driving force to respond to changes of the variable ϕ.

The eigenvalues Λ of the system are given by

$$\Lambda_{1,2} = \frac{1}{2}\left[\frac{1}{t_{s\phi}} - \frac{1}{t_s} - \frac{1}{t_\phi}\right] \pm \left\{\frac{1}{4}\left[\frac{1}{t_{s\phi}} - \frac{1}{t_s} - \frac{1}{t_\phi}\right]^2 - \frac{1}{t_s t_{s\phi}}\right\}^{1/2} \tag{18}$$

The system becomes unstable as soon as at least one of these eigenvalues has a positive real part.

Let us first identify those situations that correspond to strain-rate softening instabilities as discussed previously. Then, both $t_{s\phi}$ and t_s are positive ($m_\infty > m_0$ and $\theta > 0$) and instability takes place when

$$\frac{1}{t_{s\phi}} > \frac{1}{t_s} + \frac{1}{t_\phi}. \tag{19}$$

At the instability margin, Eq. (18) yields two conjugate complex eigenvalues (Hopf bifurcation). This indicates the onset of an oscillatory mode of plastic deformation (Zaiser and Hähner, 1997). The system becomes unstable as soon as the rate of destabilization due to a change in ϕ exceeds the sum of the rates of the stabilizing processes, viz. the relaxation of ϕ and the increase of the required driving force due to strain hardening. The instability criterion can in this case be written as

$$m_\infty < -\theta t_\phi. \tag{20}$$

This criterion is more restrictive than the strain-rate softening criterion $m_\infty < 0$ which is equivalent to $t_{s\phi} < t_\phi$. The two criteria coincide if the characteristic time ts is large in comparison with the relaxation time t_ϕ, such that one can replace ϕ by its steady-state value. It follows that the simplified constitutive description we used in the beginning implies an adiabatic elimination of the variable that brings about the strain-rate softening, resulting in an upper bound for the extension of the unstable regime.

An important difference with respect to the simplified model is encountered in situations where strain softening and strain-rate softening interfere with each other. For such situations Eq. (9) predicts stable deformation behaviour. This appears spurious: Strain softening means that the required driving force decreases with strain - in other words, at constant f there is an excess driving force available. Simultaneous strain-rate softening could lead to a stabilization of deformation if this excess in driving force would decrease the strain rate. But the stress–strain-rate characteristics cannot be inverted in this manner: A negative slope (negative m) only means that the driving force required to sustain a given strain rate decreases if this rate is increased. This is not equivalent with an increase in stress decreasing the strain rate. We are not dealing with equations of state here, even if our mathematical way of writing things down may suggest the opposite. If, on the other hand, the eigenvalues given by Eq. (18) are considered in the case that $1/t_s$ or, equivalently, θ become, then unstable behaviour is predicted: Since t_ϕ is positive, Eq. (18) yields at least one positive real-valued eigenvalue irrespective of the sign of m_∞. Instability is then characterized by the steady growth of local perturbations rather than by an oscillatory behaviour. Accordingly, in strain-rate softening materials a change in sign of θ corresponds to a modification of the character of the instability, but not to a stabilization of deformation.

The generic formalism discussed here covers several models proposed in the literature, including the models of Dietrich and Ruina for rate-and-state-dependent friction (Dieterich, 1972; Ruina, 1983), Hähner's model of the Portevin-le Chatelier Phenomenon (Hähner et al., 2002), Louchet's model of slab avalanche release (Louchet, 2001), as well as several other models of time-dependent material behavior. An important advantage of all these models is that the internal-variable approach is able to account for history dependence of the deformation process. In the simplest case, as the dynamics of the internal variable is not only driven by the straining process, the history of the sample before loading may lead to different initial values of φ and to different deformation behaviours. For instance, heat treatment of solute-hardened alloys before deformation increases the solute concentration in the dislocation cores and thus the strength, at the expense of an initial strain softening after yield. In more general, space-dependent situations the evolution of the internal variable at different points in space may lead to complex spatial strength patterns which, in conjunction with spatial couplings due to the elasticity of the material, may give rise to complex spatio-temporal deformation patterns.

5 Rock samples: acoustic emissions

An acoustic emission (AE) is defined as a transient elastic wave generated by the rapid release of energy within a material. In geological sciences studies of AE and seismology show a significant overlap. Both approaches deal with the radiation of elastic waves, although at different scales and frequencies. Generally the acoustic emissions recorded in the laboratory are generated by flaws at the grain size scale with source dimensions between micron and millimeter scale and frequency ranges between 100kHz - 20MHz (Lockner, 1993).

The damage affecting a rock under load, in the brittle regime, involves the growth of microcracks from stress concentrators such as voids, inclusions and grain contacts, resulting in both inelastic strain and acoustic emissions. The acoustic signals that are spontaneously generated from the microcracking provide information about the size, location and deformation mechanisms of the events as well as properties of the medium through which the acoustic wave travel (e.g. velocity, attenuation and scattering).

Rock fracture and earthquake rupture are also processes obeying similar statistics for source dimensions over more than eight orders of magnitude (Hanks, 1992; Zang et al., 1998).

The main goals of AE studies so far have been: the characterization of acoustic events, the localization of the source inside the sample, the statistical analysis of the event catalogue, and study of changes in presence of fluids.

5.1 Caracterization of acoustic event

Individual AE events are characterized in terms of their frequency content, amplitudes and durations so that they can be related to the micro-mechanisms that produce them.

Scholz (1968) observed a temporal correlation between the onset of AE and dilation in a sample under triaxial loading, confirming the interpretation of Brace et al. (1966) that dilation was caused by pervasive microcracking, primarily oriented parallel to the maximum compressive principal stress.

AE amplitudes and frequency have been observed to increase before failure (Zang et al., 1998 and references therein). AE focal mechanisms have also been extensively analysed Two methods have been used, taking in account the AE first-pulse amplitudes or the first- pulse sign. The first method is called moment tensor inversion, the second method is referred to as polarity study. Zang et al. (1998; 2000) used asymmetric uniaxial loading to propagate shear fracture in Aue granite. They examined fracture process zone characteristics using AE locations, and used AE first motion polarity to distinguish between dilatational (type-T), compressional (type-C) and shear (type-S) events, demonstrating that the predominant AE mechanism was shear-type cracking (about 70% of the events). AE focal mechanism studies during compressional tests on confined granites (Lockner, 1993 and references therein) revealed that in addition to pure tensile and double couple events, a significant number of mixed dilatational-compressional event types occur in the period leading to fault nucleation. Cracking at the tip of the fault was observed to be mainly tensile (type-T), while in the damage zone the shear (type-S) and mixed events were dominant (Lei et al., 2000).

5.2 Localization of the source

The localization of the source of AE events in 3D is carried on in order to image the failure processes. Insights have been provided into the nucleation phase of fracture, using AE source locations to map the temporal and spatial evolution of fracture. Lockner et al. (1991, 1992) slowed fracture to a quasi-static (stable) state using an AE feedback loading method. In the Lockner et al. study, AE source locations in granite were initially distributed uniformly throughout the sample, however close to peak stress, a clustering of AE occurred, forming a nucleus, from which fracture propagated as a process zone of intense activity. The clustering of AE during the early stages of deformation is strongly related to sample homogeneity. Pre-existing weakness or variations in material properties in the sample act as stress concentrators for damage accumulation. Thus the AE spatial clustering may allow to identify the fault nucleation zone even at low stress. By contrast, homogeneous low-porosity crystalline rocks (e.g. granites) showed much more uniform microcrack accumulation, until the peak stress is reached, or some times, even after peak stress (Lockner, 1993).

Lei et al. (2000) used a fast acquisition system to capture AE during triaxial tests on three hornblende schist samples under constant stress (creep) loading, during which a fault nucleated close to the top of the sample and propagated through to cause rupture.. In all tests, two different processes operated, a process zone in front of the fault tip and a damage zone following the process zone. The process zone was characterized as a region of intense tensile cracking and self-excitation, while the damaged fault zone was characterized by major shear events and weak self-excitation, indicating that linkage between cracks became the major mechanism of crack interaction and fault development.

Similar approaches revealed also that i) the process zone of fast propagating fault is wider then for slower ones; ii) the density of microcracks increases dramatically approaching the fault; iii) the fracture surface energy released was estimated to be ~2J (Moore and Lockner, 1995; Zang, 1998).

These experimental insights are particular important, since field observations on natural faults indicate that crack density of the fracture process zones decreases as the logarithm of the distance to the fault. Microfracture density was observed to reach a maximum value independent of fault length (Vermilye and Scholz, 1998).

5.3 Statistical analysis

The statistics of recorded events is analyzed to gain insights into the deformation processes and their rates. The Gutenberg-Richter relationship between frequency and magnitude of earthquakes also applies to experimental rock failure (Meredith et al., 1990; Sammonds et al., 1992; Ponomarev et al., 1996; Lei et al., 2003). Most of these studies were motivated by the fact that changes in *b*-value would occur prior to failure and therefore could be used for failure prediction. Laboratory studies of AE events consistently show a sharp decrease in *b*, after the initiation of faulting. Furthermore during the nucleation phases, *b*-value exhibits mutual short-term, large-amplitude fluctuations related to the heterogeneities of the fault plane. In presence of pore fluid held at constant pore fluid volume, an intermediate-term b-value minimum, after the initial b-value minimum is observed, because of the delayed formation of the fault, due to the presence of the fluid (Sammonds et al., 1992). *b*-values changes were also reported for AE events occurring prior to stick-slip events in sawcut samples (Lei et al., 2003). In this case, foreschocks which occurred during stress buildup had consistently lower *b*-values then background events, indicating a relative increase of large amplitude AE events prior to stick-slip. Moreover AE as well as earthquakes abide by power-laws in time (Ponomarev et al., 1997; Feng and Seto, 1999) and space (Ponomarev et al., 1997; Lei et al., 2004) distribution of events. Time distribution of AE displays single and multifractal structure. The heterogeneity of the rock, the stress applied and the microfracturing mechanisms control the fractal structure. AE decreasing decay rate, quantified through the Omori's law has also been related to the increasing stress level (Lockner and Byerlee, 1977). The spatial fractal dimension of AE hypocenters has been analysed through the two-point correlation integral. The decrease in D over time has been related to the change in the spatial distribution of AE events from clustering into a volume (pre-nucleation) to spreading along the fault plane (post-nucleation) (Lockner, 1993). Moreover the smaller fractal dimension of the hypocenters within individual clusters reflects local stress concentrations around the sites where intense microcracking originates (Lei et al., 2003).

5.4 Presence of fluid

Laboratory experiments were carried out both in drained and undrained conditions. The influence of fluid pressure on deformation has been investigated in drained conditions at constant pore pressure. A weakening effect of water in the brittle faulting and cataclastic flow regime has been found, because of reduction of both specific surface energy and friction coefficient. The influence of water weakening effect on the process of time-dependent deformation has been investigated during creep (constant stress) experiments, where the AE hits have been related to the slow cracking mechanisms preceding the final acceleration leading to the macroscopic failure. Recorded acoustic emissions show that the final stage of the deformation occurs for similar levels of cumulative events as well as cumulative energy (Baud and Meredith, 1997; Baud et al., 2000). Zang et al., 1996 analysed the fracture processes of dry and water saturated sandstones during uniaxial compression tests, based on elastic AE wave parameters, such as amplitude, duration time and pulse energy. Accumulated strain energy doubles for dry core. For wet specimens much lower AE activity was found, due to the increased wave attenuation. Increasing stress yields stronger AE with longer duration time independent of saturation. Recent studies focused on imaging deformation mechanisms, integrating AE during triaxial compression tests in drained conditions and post-mortem microstructural observations. This approach allowed to follow the throughgoing shear fault in Etna basalt (Benson et al., 2007) at its natural propagation (2-4mms^{-1}), without recourse to any slowing of the stress applied, as well as to monitor the nucleation and growth of compaction bands (Fortin et al., 2006) and highlight the micromechanisms associated (grain crushing and pore collapse). The strain rate and temperature dependence of Omori's law exponent has been recently showed during triaxial compression experiments in drained conditions (Ojala et al., 2004). The foreshock exponent *p'* decreases and the b-value increases with decreasing strain rate. On the contrary the aftershock exponent *p* increases with the test temperature. In undrained conditions, several studies have concentrated on dehydrating rocks at different stress conditions. New insights on the origin of intermediate and deep focus (50 to 200km) were obtained by measuring acoustic emission energy during antigorite dehydration at simulated lithostatic stress conditions, which show that brittle mechanisms can take place at depths, at which rocks should deform by plastic flow (Dobson et al., 2002). Schubnel et al. (2006) by carrying out triaxial compression experiments in creep conditions have recently shown that slow failure induced by rapid increases of pore pressure is characterised by bursts of AE, which accompany aseismic damage across the brittle-ductile transition.

5.5 New developments

New technological developments are allowing the setting of continuous ultrasonic waveform acquisition systems the Giga RAM Recorder, developed to record a 268 s segment of waveform data. This is a crucial attribute, since the rapid acceleration to failure often observed in the final phase of triaxial deformation of brittle rocks is commonly accompanied by a supraexponential increase in AE activity. Conventional recorders can infact miss important events during the 'mask-time' required to transfer data from volatile memory to permanent storage (Benson et al., 2007; Thompson et al., 2005;2007). This much higher sensitivity is allowing to image failure for a number of lithologies, allowing to track slow failure processes (Benson et al., 2007) and understand the relationships between fracture propagation and AE triggering.

To understand the physics of dynamic triggering, as well as the influence of dynamic stressing on earthquake recurrence, a pioneer study aiming to simulate periodic low loading due to Earth tides was carried out (Lockner and Beeler, 1999). Axial shortening rates between 10^{-4} and 10^{-6} mm/s were imposed to simulate tectonic loading. The degree of correlation of earthquakes was most sensitive to the amplitude of the periodic loading, rather then to the period and the average loading rate. However no AE were recorded during these laboratory experiments. More recently a laboratory study carried out with a biaxial testing system (Johnson et al., 2008) focussed on stick-slip in granular media with and without applied acoustic vibration. Small-magnitude failure events, corresponding to triggered aftershocks, occurred when applied sound-wave exceed several microstrain, inducing a strain memory in the granular material. New frontiers are given from a new generation of experimental deformation apparata, which will allow in the near future to modulate low frequency signals, that can simulate triggering mechanisms, while time to failure and AE time and spatial evolution is imaged.

Future directions will focus on the fundamental similarity of the physical processes involved in generating different frequency content seismic signals (e.g. in tectonic areas the seismic tremor under the subduction zones and the volcanic tremor and Long-period events in volcanic areas) by comparing the spectral character and frequency scaling of acoustic emissions obtained during dehydration of serpentinites and the reproduction of melt migration at high temperature while recording seismic signals (Burlini et al., 2007;2008)

Frequency scaling offers the strongest argument to assess the equivalence of the physical processes between laboratory experiments and natural volcanic seismic signals. Experimental low frequency events and tremor have frequency of about 5 MHz for intrusion lengths ranging between few tens to 200 mm. In natural earthquakes, dominant frequencies around 1-2 Hz are associated with fracture lengths of some hundreds meters to 1 km. Considering that dominant frequencies of earthquakes scale inversely with source dimension (Aki and Richards, 1980), one may write $d_1 \times f_1 = d_2 \times f_2$, where d_1, d_2 and f_1, f_2 are the dimension and frequency of laboratory $_{(1)}$ and nature $_{(2)}$, respectively. Comparing laboratory data with typical frequency (1-2 Hz) and size (1 km) of low frequency earthquakes, we obtain $d_1/d_2 = 5 \times 10^6$

and f_2/f_1= 2.5-5 x 10^6, which indicates excellent agreement between laboratory information and natural cases.

6 Triggering of earthquakes

The most common observation of earthquake triggering is the increase in seismic activity following large earthquakes, known as aftershocks. Typically, the seismicity rate just after and close to a large magnitude 7 earthquake can increase by a factor 104, and stay above the background level for several decades. Despite intense research on this problem, the mechanisms responsible for aftershock triggering are still unknown. There is a controversy in seismology about the relative importance of static stress changes (coseismic stress step induced by an earthquake), postseismic stress change (progressive reloading due postseismic deformation) and dynamic stress changes (associated with seismic waves) in triggering earthquakes.

In addition to aftershock sequences, triggered by another earthquake, there are also observations of earthquakes induced by slow earthquakes (Segall et al., 2006), rainfall (Muco, 1999; Ogasawara et al., 2002), volcanic activity (Dieterich et al., 2000, Spicak and Horalek, 2001), and deep crust degassing (Chiodini et al., 2004; Miller et al., 2004). Earthquakes can also be induced by human activities, such as quarrying, mining activities, artificial lakes impoundments and their water level fluctuation, gas and oil production, geothermal production and fluid injection at depth (see Mcgarr et al. (2002) for a review).

In summary, any artificial or tectonic stress change may trigger seismic activity. There is no clear threshold in the amplitude of the perturbation (Ziv and Rubin, 2000); ocean tides produce significant changes in seismic activity even if the amplitude of the stress change is much smaller than the stress drop associated with an earthquake (Cochran et al, 2004). Seismicity can be triggered by fast perturbations such as coseismic stress change or seismic waves, or by much slower forcings (tides, gas and oil production, slow earthquakes, water level fluctuation, …).

We first summarize observations of seismicity and then discuss the different models that have been suggested to explain earthquake triggering.

6.1 Triggering of aftershocks

6.1.1 Aftershock rate as a function of time

Seismicity rate increases immediately at the time of the mainshock, and then decays approximately as the inverse of the time since the mainshock (Omori, 1895). The aftershock decay rate is well described by the modified Omori's law $R(t) = K/(t+c)^{p}$ (Utsu et al., 1995) where t is the time elapsed since the mainshock. The exponent p is generally close to 1 and the short-times cutoff c is not larger than a few minutes (after correcting for the increase in the detection threshold of the seismic network after a large mainshock) (Kagan, 2004; Peng et al., 2006). This decay is roughly independent of the mainshock magnitude (Helmstetter et al., 2005).

6.1.2 Aftershock rate as a function of mainshock magnitude

Aftershock sequences of small earthquakes are less obvious than for larger mainshocks, because the aftershock productivity is weaker, but they can be observed after stacking many sequences. The rate of triggered earthquakes then increases with the mainshock magnitude M as 10^{aM}, with an exponent a of the order of 1, i.e., the number of aftershocks is roughly proportional to the mainshock rupture area $S \sim 10^{M}$ (Felzer et al., 2004; Helmstetter et al., 2005). Small earthquakes thus have a significant contribution in earthquake triggering because they are much more numerous than larger ones (Helmstetter et al., 2005). As a consequence, many large earthquakes are triggered by previous smaller earthquakes (« foreshocks »)

6.1.3 Magnitude distribution of "triggered earthquakes"

Aftershocks are usually thought as small earthquakes induced by a preceding larger earthquake. The magnitude of the largest aftershock is on average 1.2 unit of magnitude below the mainshock (Bath, 965). But there are a few cases of large earthquakes triggered by a preceding smaller event. In this case, they are called "foreshock-mainshock" sequences rather than "mainshock- aftershocks" sequence, the largest event of a sequence being usually called "mainshock". However, the same mechanisms may explain both the triggering of aftershocks by a previous larger earthquake, and the triggering of a large earthquake by a previous smaller one (Helmstetter and Sornette, 2003). Indeed, the distribution of times between an earthquake (‘‘foreshock’’) and a subsequent larger earthquake follows the Omori law as for usual aftershocks.

If "aftershocks" are now defined as any earthquake triggered by a preceding event (selected within the "influence zone" of that earthquake), then the magnitude distribution of aftershocks is found to obey the Gutenberg Richter distribution, as for any other earthquake. The size of a triggered earthquake is not constrained by the mainshock size, only the number of triggered events is observed to grow with the mainshock magnitude. This suggests that the size of an earthquake is not predictable, and that any small event may grow into a large earthquake.

6.1.4 Aftershock rate as a function of distance from mainshock

Most aftershocks are located on or close to the rupture surface. But there are also observations of triggering at very large distance from the mainshock, up to about 10 times the mainshock rupture length (Hill et al., 1993; Felzer and Brodsky, 2006). Far field triggering is more frequent in geothermal or volcanic area (Brodsky et al., 2000). Seismic activity is found to increase almost everywhere close to the mainshock: there are no clear observations of "stress shadows", i.e., a decreased seismicity after the mainshock (Marsan, 2003; Felzer and Brodsky, 2005). Aftershock distribution in space is almost constant with time; there is only a very slow diffusion of aftershocks with time, if any (Marsan and Bean, 2003; Helmstetter et al., 2003).

6.2 Induced seismicity

Seismicity triggered by human activities was first recognized in South Africa in 1894, when quakes triggered by gold mining were felt. Since then, seismicity associated with petroleum production became apparent in the early 1920s, with reservoir impoundment in the late 1930s, with high pressure injection at depth in the mid-1960s, and with natural gas production in the late 1960s (for a review and references see Mcgarr et al. 2002). Induced seismicity can be triggered by changes in pore pressure, in normal stress or in shear stress (Mcgarr and Simpson, 1997). Surface quarries and deep mines induce seismicity primarily by changing the elastic stress field. Increased pore pressure is the dominant factor for earthquakes triggered by fluids injection at depth. Reservoir loading can trigger earthquakes either by changing the normal stress, the shear stress, or the pore pressure. Oil and gas field compaction primarily results in changes in the state of stress within the rock mass surrounding the reservoir.

It is always difficult to identify the forcing that trigger earthquakes, and even more to predict how the crust will respond to a given perturbation. There is no clear relation between the forcing (amplitude and loading rate) and the change in seismic activity (maximum size of the triggered earthquakes, time delay between loading rate and seismicity rate, interaction distance) (e.g. for a review Grasso and Sornettte 1998; Mcgarr et al. 2002). The case studies provide opportunity to quantify the distance to failure, which can be as low as a few 0.01MPa. This observation is consistent with the idea that the deviatoric stress in the crust is very close to the strength (e.g., Zoback and Harjes, 1997; Grasso and Sornette, 1998).

6.3 Stress changes responsible for earthquake triggering

6.3.1 Static stress changes

Aftershock triggering is commonly explained by the static stress change induced by the mainshock (e.g., Stein, 1999; Steacy et al., 2005). Static stress changes are permanent changes induced by the coseismic slip on the mainshock fault. Coulomb stress change calculations have been used to predict the locations, focal mechanisms and times of future earthquakes (see reviews by Stein (1999) and Steacy et al. (2005)). The success of this approach is significant but limited. Only about 60% of aftershocks are located where the stress increased after a mainshock (Parsons, 2002); stress shadows (proposed decrease of the seismicity rate where Coulomb stress change is negative) are seldom or never observed (Marsan, 2003; Felzer and Brodsky, 2005) and the correlation of Coulomb stress change with aftershock locations is rather sensitive to the assumed slip distribution (Steacy et al., 2005).

Other studies have focused on the temporal decay of aftershocks, and suggested several models to explain how a permanent stress change may explain Omori's law. The temporal decay between the stress step and aftershocks occurrence can be explained by the existence of a "nucleation phase" between the time of the stress change and slip instability (Dieterich, 1994).

6.3.2 Dynamic stress changes

Dynamic stress changes are associated with the seismic waves induced by the mainshock. Dynamic stress change can be much larger than static stress change close to the mainshock, and decay slower with distance (Aki and Richards, 1980). But they have a limited duration, of the order of minutes for "large" earthquakes. The observation of triggering at very large distances from an earthquake seems to suggest that triggering is caused by a dynamic stress change (Hill et al., 1993; Brodsky et al., 2000; Felzer and Brodsky, 2006). Whereas the observation of triggering at very long times after an earthquake, up to decades, is hard to explain by dynamic perturbations (Gomberg et al., 1998; Gomberg, 2001), but more likely result form permanent stress changes (Dieterich, 1994). Recently, Brodsky et al. (2003) suggested that far field triggering may be explained by dynamic stress perturbations induced by seismic waves, which can produce permanent water level changes (and thus stress and deformation change) by removing barriers in fractures within an aquifer, a mechanism that may also apply to landslides. The observation that aftershock density is correlated with the directivity of the rupture also favors dynamic triggering, because rupture directivity increases seismic wave amplitude but does not affect static stress change (Gomberg et al., 2003).

6.3.3 Postseismic deformation

Most shallow large earthquakes are followed by a significant postseismic deformation, and by an increase in seismic activity, which both start immediately after the mainshock and last for several years (Marone et al., 1991). Postseismic deformation is most often localized around the rupture zone, and is thus modeled as afterslip on the mainshock fault. The link between aseismic afterslip and aftershock activity is however not clear. The cumulative moment released by aftershocks is usually much lower than the one associated with afterslip, which implies that postseismic deformation is not due to aftershock activity. The similar time decay and duration of postseismic deformation and aftershocks rather suggests that aftershocks can be induced by afterslip (Dieterich, 1994; Wennerberg and Sharp, 1997; Schaff et al., 1998;

Hsu et al., 2006; Savage, 2007). Indeed, afterslip induces a progressive reloading of faults that are locked, which can trigger aftershocks.

Most studies who suggested that aftershocks are due to afterslip assumed that aftershock rate is simply proportional to stress or strain rate (Wennerberg and Sharp, 1997; Schaff et al., 1998; Hsu et al., 2006). However, the relation between seismicity rate and stress rate can be much more complex (Dieterich, 1994; Helmstetter and Shaw, 2007). Afterslip is thus a possible candidate to explain observations of aftershock rate decaying as a power-law of time with an Omori exponent that can be either smaller or larger than 1.

The mainshock will also induce a change in fluid pressure. Fluid flow will then progressively redistribute the fluid pressure. The postseismic reloading induced by fluid flow is thus a possible mechanism to explain aftershock activity (Nur and Booker, 1972; Miller et al., 2004).

6.3.4 Fluid flow

The main role of fluids pressure perturbations in the upper crust, in addition to the change in physical and chemical properties of rocks, is their impact on the stress field. The fluid pressure variations impact differently the stress field applied on the fractures or on the bulk material, depending on their relative permeability (or on their biot's coefficient when considering the hydro-mechanical behavior). This effect has been early modeled through the concept of effective stress established by Therzaghi (1925, 1943). The effective stress is defined as the standard stress minus the fluid pressure. This model applies to a restricted domain of the hydro-mechanical behaviour of porous rocks for which the fluid pressure is integrally subtracted from the standard stress (see the Biot's theory for a more general approach (Biot, 1941)). Despite this restriction, this concept gives an efficient framework to explain how an increase of fluid pressure can lead to failure by reducing the distance to the failure criterion. This is the first order explanation for the triggering process of seismicity induced by changes in fluid pressure (see the reviews of Grasso and Sornette (1998), McGarr et al. (2002) and Gupta (2002)).

6.4 Modeling the response of the Earth's crust to stress changes

Many mechanisms have been proposed to explain earthquake triggering by stress changes, such as stress corrosion (Das and Scholz, 1981; Yamashita and Knopoff, 1987; Gomberg, 2001; Shaw, 1993), rate-and-state friction (Dieterich, 1994), static fatigue (Scholz, 1968) and damage rheology (Ben-Zion and Lyakhovsky, 2006). Other studies simply assumed that seismicity rate is proportional to stress rate (Schaff et al, 1998; Hsu et al., 2006; Felzer and Brodsky, 2006). This would be the case if a fault breaks instantaneously when it reaches the rupture threshold, i.e., without nucleation phase.

The rate-and-state model (Dieterich, 1994) is the more widely applied method to model the response of faults to stress changes. This model assumes a population of faults obeying the laboratory-derived rate-and-state friction law (Dieterich, 1978). If the stress change is increased uniformly due to a mainshock, the change in seismicity can be approximated by Omori’s law with p=1 for intermediate times after the main shock. Introducing spatial heterogeneity of the stress change modifies the temporal decay of aftershocks. It produces triggering at short times with an apparent exponent p smaller than one, followed by a "stress shadow" for times of the order of the nucleation time (seismicity is lower than the background rate) (Marsan, 2006; Helmstetter and Shaw, 2006).

The rate-and-state friction law can also be used to model seismicity triggered by dynamic stress change (Gomberg et al., 1998; Gomberg, 2001), by a slow slip event (Segall et al., 2006), or by postseismic reloading (Marone et al., 1991; Dieterich, 1994; Helmstetter and Shaw, 2007). This model is unable to reproduce Omori’s law in the case of dynamic triggering, because seismicity rate returns to the background rate when the stress perturbation is over (Gomberg, 2001). In the case of power-law decaying afterslip, this model produces Omori’s law decay of aftershocks, with an exponent p smaller or larger than 1, as observed for aftershock sequences (Dieterich, 1994; Helmstetter and Shaw, 2007).

More generally, Dietrich’s model predicts that seismicity rate scales with stress rate for slowly varying stress changes (with period of the order of several years), but scales with the exponential of the stress change in the case of fast perturbations (at time scales much smaller that earthquake nucleation, e.g., for tides) (Dieterich, 1994; Cochran et al., 2004; Helmstetter and Shaw, 2007). In general, the process of earthquake nucleation introduces a time delay between stress change and seismicity rate. The distribution of aftershocks in time and space is thus generally different from that of the stress perturbation. The complexity of fault mechanics thus makes it difficult to infer the mechanisms responsible for earthquake triggering based on observations of stress changes. The modeling of fault slip and seismicity, and even more the characterization of the fault rheology based on seismicity or geodesy data, is thus a difficult challenge, in terms of finding which mechanism may be causing an observed time dependence.

7 Triggering of landslides

Catastrophic ruptures of landslides are observed to increase following large nearby earthquakes, heavy rain, and/or freeze–thaw cycles (e.g., Caine, 1981; Keefer, 1984; Sandersen et al., 1996; Matsuoka and Sakai, 1999; Frayssines and Hantz, 2006). The displacement rate of many slow moving slopes is also observed to be modulated by changes in external loading, such as rainfalls, snow melt episodes, moderate and small regional earthquakes, deglaciation. The time-scales of these forcings range in time from a few Hz for seismic pulses to 100000 years for deglaciation cycles. From soil and rock mechanics point of view, four processes are identified to trigger landslides. The combination of these drivings is more likely to trigger landslides than a single one but for clarity we review each process individually

7.1 Triggering of landslides by increasing the slope

This geometrical effect can be driven by anthropogenic processes, such as road cutting within hillslope or other urbanization structures as reported for landsliding in urban area of Rio de Janeiro, Brazil (Smyth and Royle, 2000). As another example, Starck et al. (2005) analyze the influence of excavation of the landslide toe. Natural processes such as erosion of the hillslope toe by fluvial incision or glacial retreat are also proposed to change the slope inclination. Holm et al. (2004) demonstrated the influence of glacier retreat on landslide triggering in British Columbia. Note that a change of slope can also result from an earthquake, but it is difficult with the available data to test whether the landslides are triggered by geometrical effects or by the shaking due to seismic waves.

7.2 Triggering of landslides by increasing the load applied on the slope

An increase of the weight of the landslide may be due to erosion processes or rainfall. For instance, Chigira and Yokoyama (2005) showed the role of weight increase for landslides triggered by rain in Kyushu, Japan. Infiltrating water from rain increases the weight of weathered material and decreases the suction within the material, which is the final trigger of a shallow landslide.

7.3 Influence of weathering on landslide triggering

Ground water level and pore pressure changes result from the infiltration from surface, or exfiltration from bedrock, preferential flow and convergent flow leading to water accumulation. A few simple empirical relationships have been suggested between the amount of rainfall and landsliding. The first relationship is for wet soils that have endured more than 20mm of rain during two days before failure, and gives the threshold intensity of the rainstorm I for triggering landslides: $I = 12.64 D_S^{-0.39}$ with D_S is the storm duration.

The other relationship for dry soils (<20mm of rain during the 2 days before) $I = 19.99 D_S^{-0.58}$ was first suggested by Caine (1980) and later modified by Sidle and Ochiai (2006). These relationships were obtained using 67 worldwide events. Crozier (1999) proposed another model taking into account soil moisture for events in Wellington City, New Zealand.

There are other factors that influence the triggering of landslides by rain. Dhakal and Sidle, (2004) analyzed 82 major rainstorms from 1972 to 1990 in Carnation Creek, British Columbia, and found that the combined influence of mean and maximum hourly intensity, duration and total amount of rainstorms were important in generating landslides. In addition, other characteristics of the landslide such as slope angle, vegetation cover, soil depth, and drainage pattern should be taken into account for understanding and forecasting rainfall induced landslides (National hazardscape report, 2007)

In the French Alps, a positive correlation has been obtained with freeze–thaw cycles and a slight correlation with rainfall (Frayssines and Hantz, 2006). This suggests that ice jacking could the main physical process leading to failure by causing microcrack propagation. In Norway, the distribution of rock falls along the year shows two maxima, in early spring and late autumn, which coincide with the periods of frequent variations of temperature around the freezing point (Sandersen et al., 1996). The first maximum also coincides with the time of highest rate of snowmelt, the other with the months of highest precipitation. An analysis of the rock fall activity in the Hosozawa Cirque, Japan (Matsuoka and Sakai, 1999), concluded that the intense activity does not reflect precipitation events, but that the primary factor controlling rock falls is seasonal frost weathering.

7.4 Triggering of landslides by earthquakes

This triggering process is very efficient but not quantitatively understood. Keefer (1984, 2002) observed that, for a worldwide catalogue of landslides and earthquakes, the minimum magnitude of a triggering earthquake is 4, and that the area were landslides occurred increases with magnitude M from A=0 for M=4 to A=250 km² for M=5.4 and A=500000 km² for M=9.2. The relationship between the potential area A affected by landslides (in km²) and M is given by Keefer and Wilson (1989) : $\log_{10} A = M - c$ with $c = -3.46 \pm 0.47$.

Other studies were carried on by Hancox et al. (2002), who found that the minimum magnitude for landslide triggering in New Zealand was M=5. Hancox et al. (2002) suggested a relationship based on 22 events in New Zealand which predicts that the area affected by landslides and the maximum epicentral distances for landslide triggering is generally smaller than estimated from the worldwide data of Keefer (1984). Papadopoulos and Plessa (2000) analyzed 47 landslides in Greece and the data gave results similar to Keefer’s.

Many factors related either to the earthquake source or to the geological setting can influence the number, sizes and types of landslides. Factors related to the intrinsic properties of earthquakes include earthquake magnitude, focal depth (Keefer, 1984, 2002; Meunier et al., 2007). Factors related to the environment include inherent stability of the potential failure sites, existence of old or dormant landslides, vegetation and land use, regolith wetness, and slope gradient and other topographic factors (Sidle and Ochiai, 2006). In addition, two phenomena are known to enhance failure during earthquakes: the first one is site effects (amplification of seismic waves induced by local resonance), as suggested by Sepúlveda et al. (2005) for the 1994 M= 6.7 Northridge earthquake. The second one is liquefaction (soil behaving as a fluid under shaking) (Zerkal, 1996).

7.5 Modeling the triggering of natural hazards

The difficulty in modeling the triggering of natural hazards, and identifying the mechanisms responsible for failure, results both from the lack of data about natural phenomena and about material properties (stress, strength, friction coefficient…), and the complex response of natural objects to external perturbations.

A consequence of this non-linear relation between stress changes and seismic activity is that the behavior of the system not only depends on the average values of the stress changes, and of the material properties, but rather on their distribution (extreme values) (Marsan, 2006; Helmstetter and Shaw, 2006). In order to model earthquakes, or landslide triggering, we thus need to know the control parameters of the system (stress, strength, and other mechanical properties) at all scales, which is not possible for natural objects.

Because physical modeling of earthquakes and landslides is so difficult, it is seldom used to forecast natural hazards. Rather, empirical approaches are applied. For earthquakes, epidemic-type models are used to forecast the distribution of seismicity in space, time, and magnitude (Kagan and Knopoff, 1987; Helmstetter et al., 2006). These models assume that any earthquake is able to trigger other earthquakes, with a rate that decays in time according to Omori's law, and increases with the mainshock magnitude. The spatial distribution of future aftershocks is also better predicted by previous aftershocks than by coulomb stress change calculations (Felzer et al., 2003). For landslides or rockfalls, the distribution of the rock mass involved in past events can be used to infer the probability of having a future event of any size (Dussauge et al., 2003).

8 Triggering of snow avalanches

Snow avalanches represent a major natural hazard in snow covered mountain areas throughout the world. Their occurrences affect ski resorts, roads, railways, power lines, communication lines, forests, backcountry recreationists, residential areas, and industrial facilities (e.g. mining). The number of fatalities per year due to snow avalanches is estimated to be at least about 250 worldwide. Within the last ten years (1996-97 to 2005-06) about 1020 people were killed in the European Alps. In Switzerland, for example, the direct and indirect costs of the avalanche disaster in February 1999 amounted to about EUR 500 million. Due to avalanche protection work established in the past decades the damage to infrastructure and residential areas has been reduced, so that today most of the fatalities involve personal recreation on public land (Jamieson et al., 2002; Schweizer, 2004).

Whereas the population and recreation pressure in many regions of the European mountains is still increasing, the financial means for avalanche protection work are rather declining. This increases the relevance of temporary protection measures such as road closures. However, temporary protection measures heavily rely on reliable avalanche forecasting. It is therefore essential to know where and when avalanches are to be expected which requires an in-depth understanding of the triggering mechanisms.

8.1 Avalanche phenomenon

Snow avalanches are a type of fast-moving mass movement. They can also contain rocks, soil, vegetation or ice. There are two types of release: loose snow avalanches and slab avalanches. Loose snow avalanches start from a point, in a relatively cohesionless surface layer of either dry or wet snow. Initial failure is analogous to the rotational slip of cohesionless sands or soil, but occurs within a small volume ($<1m^3$) in comparison to much larger initiation volumes in soil slides.

Snow slab avalanches involve the release of a cohesive slab over an extended plane of weakness, analogous to the planar failure of rock slopes rather than to the rotational failure of soil slopes. Slab avalanches have similar physical appearance regardless of size (McClung, 2003). Avalanche width increases with avalanche thickness. The observed ratio between width and thickness of the slab varies between 10 and 10^3, and is typically about 10^2. Slab thickness is usually less than 1 m, typically about 0.5 m, but can reach several meters in the case of large disastrous avalanches (Schweizer et al., 2003). The slab thickness is the fundamental size parameter and follows a lognormal probability density function.

Avalanche size is classified according to its destructive power. A medium sized slab avalanche may already involve 10,000 m^3 of snow, equivalent to a mass of about 2000 tons (snow density 200 kg/m^3). Avalanche speeds vary between 50 and 200 km/h for large dry snow slides, whereas wet slides are denser and slower (20-100 km/h). If the avalanche path is steep, dry-snow avalanches generate a powder cloud.

Most avalanches release from terrain steeper than about 30°. Only a low percentage of dry slabs start on terrain under 30°, but wet slides can occur on slopes under 25°. The slope angle is the most important factor influencing avalanche formation.

Most snow avalanches start naturally during or soon after snow storms. Failure is due to overloading an existing weakness in the snowpack. High precipitation rates favor snowpack instability. In general, about 50 cm of new snow within 24 hours (equivalent to about 50 mm of precipitation) is critical for avalanche initiation. Large disastrous avalanches usual follow storms that deposit more than 1 m of snow.

Snow slab avalanches can also be triggered artificially – unlike most other rapid mass movements – by localized, rapid, near-surface loading by, for example, people (usually unintentionally) or intentionally by explosives used as part of avalanche control programs. In general, naturally released avalanches mainly threaten residents and infrastructure, whereas human-triggered avalanches are the main threat to recreationists.

Avalanche formation is usually, e.g. in avalanche control programs, assessed heuristically by weighing the so-called contributing factors: terrain, precipitation, wind, temperature and snow stratification, i.e. the complex interaction between terrain, snowpack and meteorological conditions are explored (Schweizer et al., 2003). In the following, we will describe the mechanical processes of avalanche formation. Snow avalanche formation depends on a combination of critical material properties at different scales. We will therefore briefly introduce snow as a material.

8.2 Snow as a material

Snow is an rather unusual material especially with respect to its high homologous temperature (T/T_{melt}). Under terrestrial conditions, the homologous temperature of snow is higher than 0.9, which can be considered as a material at very high temperature state. Snow is a sintered material of usually monocrystalline ice particles. Mechanical properties depend strongly on temperature and strain rate. The large changes in material properties are also caused by the large range of porosities and of grain sizes: the porosity can vary from over 90% to about 30%, where the pore close-off occurs. The grain size varies from about 10 micrometers to several millimetres (Schneebeli et al., 2006).

As snow can be seen as a foam of ice, it should be possible to derive the snow properties from the ice properties by considering the complex microstructural geometry. However, presently, a comprehensive set of geometrical parameters that describes the wide range of naturally occurring snow microstructure is lacking. Nevertheless, some geometrical parameters were found and related to selected physical properties. For example, the specific surface area of the snow microstructure was shown to be related to the optical properties of snow (Matzl and Schneebeli, 2006).

Recent advances in microstructural imaging and simulation (Schneebeli, 2004b) have improved our understanding of snow failure. Mechanical simulations at the microstructural level also show stress concentrations. The distance and extent of such stress concentrations is very dependent on the microstructure. However, no direct mechanical observations are yet available of theses processes. The upscaling from the microstructural behavior to a macroscale (continuum) property is in a very initial stage.

The specific strength of snow is low ($\sigma_t/\rho = 2 \ldots 20$ m^2 s^{-2}) and so is the fracture toughness (0.2 … 2 kPa m$^{-1/2}$) which makes snow one of the most brittle natural geo-materials. The bulk mechanical behavior of snow suggests that snow can be described as a quasi-brittle, pressure sensitive, strain softening material. During shear deformation dilatation is observed. The strong temperature and strain rate dependence are related so that snow is also described as a rheologically simple material. Snow deformation is characterized by irreversible microstructural changes, i.e. the elastic limit strain is very small (10^{-4}) (Camponovo and

Schweizer, 2001). Beyond this limit, substantial deformation implies re-arrangement of grains. These microstructural changes suggest that snow deformation and, in particular, the strain-rate and temperature dependence, can be understood in terms of the two fundamental (and competing) processes of bond breaking and bond formation (re-bonding or sintering) (Schweizer, 1999). This view has recently been confirmed by numerical simulation (Reiweger et al., 2007). Snow failure ultimately occurs when the bond breaking rate exceeds the bond formation rate. As snow is a highly porous material, the failure of snow, i.e. the breaking of bonds over a considerable area, will inherently cause structural breakdown (collapse). This vertical displacement has been observed and is on the order of the grain size of the failure layer (van Herwijnen and Jamieson, 2005).

8.3 Snow cover as a layered material

Snow avalanches occur when parts or the whole seasonal snow cover on a slope fail. The snow cover is a layered structure with varying properties in time and space (horizontal and vertical) (Colbeck, 1991). The layers originate from a deposition process either by precipitation or sublimation. The high homologous temperature and the high porosity cause that the layers, i.e. the snow microstructure and hence their properties, are constantly changing – even without additional external load – due to vapour pressure differences. Kinetic crystal growth from the vapour phase is the primary source for the formation of weakly bonded snowpack layers which represent a prerequisite for slab avalanche formation.

Spatial variations in snow cover properties play a crucial role in avalanche formation. Disorder exists at various scales starting with the disordered arrangement of the ice particles, the snow grains, (leading to unpredictable load lines), up to the disordered deposition of snow on an avalanche slope due to the interaction of wind with terrain.

8.4 Avalanche release

The release of a dry-snow slab avalanche is essentially a fracture process. Below a cohesive slab along an extended plane of weakness a fracture spreads from a local failure with the result that high tensile stress and tensile fracture develops upslope – and eventually the whole slab releases. Slab release includes (1) snow failure (damage) leading to failure localization (initial failure) and (2) fracture propagation. Both processes depend on the slab and weak layer properties as well as on their interaction. According to best estimates the critical size for self-propagating fractures is on the order of the slab thickness, i.e. approximately ≤ 1 m (e.g. Bazant et al., 2003).

One of the first snow slab avalanche release models (McClung, 1979) adapted a model that described the failure of an over-consolidated clay (Palmer and Rice, 1973). Recently, slab release has been revisited applying beam theory and taking into account bending of the slab due to the localized initial failure (Heierli and Zaiser, 2006; Heierli and Zaiser, 2007). They have shown that the energy barrier for crack propagation can be considerably lowered by an additional slope normal collapse of the weak layer.

To trigger a slab avalanche an initial failure (finite area below the slab that failed) is required (as mentioned on the order of ≤ 1 m). Apart from the steepness of terrain and the slab weak layer structure (which has to have sufficient spatial extent), the detailed conditions under which the initial failure forms are generally poorly understood. As avalanches are relatively rare events, it seems clear that only in a metastable condition, close to a critical state, external and/or internal perturbations might trigger an avalanche. The following triggering scenarios are the most common ones.

8.4.1 Stress increase beyond threshold

This is the most common failure mechanism. The initial failure is triggered by either gradual uniform external loading due to, for example, precipitation, or rapid localized external loading by, for example, people or explosives. For natural releases during or shortly after storms, the precipitation rate can strongly influence the critical balance between stress and strength. There is a competition between the rate of loading and the rate of strengthening of buried weaknesses (Schweizer et al., 2003). If the loading is rapid, the weak layer might not gain strength sufficiently quickly. It is believed that a slow damage process at the bond scale creates a failure along the weak layer (McClung, 1979) due to a combination of weakest point and strongest stress. If this failure reaches a critical size, it will rapidly propagate along the weak layer, thereby releasing a slab avalanche. For rapid localized near surface loading however, the external load directly imparts deformations in the weak layer which are large enough to create a propagating brittle fracture in the weak layer. Recent field observations have shown that fractures easily initiate in weak layer below skiers, without necessarily causing the release of a slab avalanche (van Herwijnen and Jamieson, 2005). This indicates that fractures must not only initiate, but also propagate in order to release an avalanche. Several studies have examined the role of snowpack parameters in skier-triggered avalanches (e.g. Schweizer and Jamieson, 2003) showing that there are significant variables associated with instability, such as weak layer grain size and hardness and difference in hardness and grain size across the failure interface. While these variables are likely to affect slab avalanche release in terms of fracture initiation (i.e. stress transmission, stress concentration in the weak layer and strength of the weak layer), the relation between these snowpack variables and fracture propagation is less clear.

8.4.2 Constant stress

Local reorganization of structure causing weakening (strain softening) without external loading. Temperature and radiation are decisive factors contributing to avalanche formation in situations without loading. A positive energy balance at the snow surface primarily affects the surface layers, whereas the weak layer is relatively unaffected due to the low thermal conductivity of snow. Even though snow strength decreases with increasing temperature, instability after rapid warming does not result from a weakening of the weak layer below the slab but rather from increased deformation of the surface layers of the slab, leading to increased strain and strain rates at the interface between the slab and the weak layer (Schweizer et al., 2003).

8.4.3 Weakening due to water percolation and storage at capillary barriers (or impermeable layers)

Liquid water dissolves the bonds between individual snow grains thereby decreasing the strength of the weak layer until it fails releasing a wet snow slab avalanche. This weakening preferably occurs when the water above a layer interface increases due to a significant difference in grain size (capillary barrier) or due to an impermeable layer (or the ground) (Schneebeli, 2004a). Tough air temperature is believed to be critical for forecasting these avalanches, recently it has been shown that – as in the case of dry snow avalanches – the snowpack properties are as essential as the external forcing (Baggi and Schweizer, 2007). When it rains, avalanches can release within minutes of the onset of rain, but delays of up to half a day or more are possible (McClung and Schaerer, 2006). Wet snow avalanches are notoriously difficult to forecast because release depends on the complex interaction of water percolation (which is highly non-linear and variable in space and time), topography and snowpack properties. The poor predictability follows from the assumption that the critical state of instability is achieved under only very special conditions so that timing and small differences in forcing become crucial – in other words the sensitivity to small perturbations (triggers) is highly non-linear.

8.5 Scale and variations

Regardless of the triggering mechanism the question of scale is essential for the understanding of avalanche formation. Spatial variations of snowpack properties at various scales are directly linked to the failure process. Variations are crucial because they provide the nucleus of fracture and under some conditions a necessary stabilizing mechanism to limit damage and inhibit fracture localization (Schweizer et al., 2003). Spatial heterogeneity of layer properties is due to various external and internal processes interacting with topography during and after the deposition process. If the autocorrelation length of spatial variation is less than the critical length for self-propagating fractures an initial failure might not propagate (Schweizer et al., 2007). Despite numerous studies, the typical variation is unknown but varies presumably according to the meteorological conditions at the time of deposition.

8.6 Modeling

Up to now, the variety of triggering mechanism, the wide range of scales and the variety of disorder at these scales, has impeded a comprehensive model of slab avalanche triggering, though some promising attempts exist (e.g. Fyffe and Zaiser, 2007). Modelling across scales taking into account the disorder (which has partly opposing effects on avalanche formation) is considered the key to a better understanding of snow slab avalanche release which is ultimately required to improve avalanche forecasting.

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Section 2

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Section 8

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