SLOW EARTHQUAKES: AN OVERVIEW

THEORETICAL SEISMOLOGY

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ABSTRACT

Slow earthquakes have been observed in California and Japan. This type of earthquakes is characterized by nearly exponential strain changes that last for hours to days. Studies show that the duration of these earthquakes ranges from a few seconds to a few hundred seconds. In some cases, these events are accompanied by non-volcanic tremor that suggests the presence of forced fluid flow. Slow earthquakes occurrence suggests that faults can sustain ruptures over different time scales, that they have a slow rupture propagation, a low slip rate or both. However, the loading of the seismogenic zone by slow earthquakes has not been yet fully understood. In this work, I present an overview of the causes that originate slow earthquakes and I also provide the Cascadia example of events that have been studied in the past.

1. INTRODUCTION

Earthquakes occur as a consequence of a gradual stress buildup in a region that eventually exceeds some threshold value or critical local strength, greater than that the rock can withstand, and it generates a rupture. The resulting motion is related to a drop in shear stress. The rupture propagation may be controlled by the failure criterion, the constitutive properties and the initial conditions on the fault, such as the frictional properties of the fault surface or the stress distribution around it (Mikumo et. al., 2003). This rupture has been observed to be on the order of 10^{-1} to 10^{5} m on micro- to large earthquakes and on the order of 10^{-3} to 1 m on laboratory-created rupture (Ohnaka, 2003).

In seismology, several studies have focused on the determination of a constitutive law related to the failure process that specifies the dependence between fault slip, stress, slip rate, and other physical properties. It has been found that such relation plays an important role in the dynamic component of the rupture process and on strong ground motions during the occurrence of large earthquakes (Mikumo et. al., 2003). To formulate the constitutive relation for the earthquake rupture two different approaches have been followed: The rate-and state-dependent formulation and the slip-dependent formulation.

The former was developed by Ruina (1983) and Dieterich (1978, 1979, 1981 and 1986). It states that the shear traction is a function of independent variables such as the slip velocity and, at least, one state variable. However, this law does not consider the variations in stability of the shear fracture process when the material is intact. In other words, it does not consider slip failure on a pre-existing fault and shear fracture of the rock simultaneously. In contrast, the slip-dependent constitutive law expresses the shear traction as a function of the slip displacement while the breakdown stress drop and displacement, as well as the peak shear strength, are functions of the slip velocity or time (Onhaka et al., 1996; Onhaka et al., 1997). The concept of slip-weakening behavior has been widely accepted since it was proposed by Ida (1972) and Palmer and Rice (1973), and was later simplified by Andrews (1976a,b).

One of the most notable features of rupture phenomena is that, according to Ohnaka (2003), some of the physical quantities related to the rupture, such as the size of the nucleation zone, the slip acceleration, and the shear rupture energy are scale-dependent. To account for these scale-dependent quantities, Onhaka derives a constitutive scaling law that enables the proper integration and quantification of both frictional slip failure and shear fracture of intact rock over a broad range of earthquake sources. This scaling constitutive law will be discussed later in this paper.

In addition to ordinary earthquakes, events with high-speed rupture propagation that produces a train of high-frequency body waves, but their overall duration is relatively long compared to ordinary episodes, are known as slow earthquakes. This type of earthquakes also displays an anomalously high level of low-frequency excitation (Beroza and Jordan, 1990). This classification has been applied to events with durations in the range of ~1s to 10^6s (Kanamori and Hauksson, 1992; Linde et al., 1996). This paper focuses on the processes that generate earthquakes and discusses the variations in such processes that originate slow earthquakes. An example of well documented slow earthquakes on Cascadia is also addressed.

2. SLIP WEAKENING AND STRENGTHENING

According to Muhuri et al (2003) when an earthquake occurs, the slip instability is controlled by a combination of mechanical and chemical processes along the fault. Chemical activity over long time may strengthen the fault gouge. For slip to occur, fault shear stresses should exceed strength gained by such chemical activity. High-velocity friction experiments performed by Ujiie et al. (2009) showed that slip-weakening is associated with the formation and shearing of low-viscosity melt patches, whereas the subsequent slip strengthening occurs despite the reduction in shear strain rate due to the growth or thickening of melt layer, suggesting that the viscosity increases with slip, primarily by dehydration of the melt layer.

2.1 The slip-weakening model

In 1976, Andrews proposed the simple slip-weakening model shown in Figure 1. In this figure, which exemplifies the relation between the shear stress and the slip displacement, τ_i is the initial stress at which the shear stress τ begins to increase up to the yield stress τ_p as the fault begins to rupture. D_a is defined as the critical displacement at which the shear stress reaches the yield stress τ_p. Then, it decreases with ongoing slip displacement D_c to the residual friction or dynamic friction level τ_r. The breakdown stress drop $\Delta\tau_b$ is defined as $\Delta\tau_b=\tau_p-\tau_r$, where the slip-weakening displacement D_{wc} is defined as the slip displacement required for the shear strength to degrade from τ_p to τ_r.

So we have that

$$D_{wc} = D_c - D_a \qquad (1)$$

The effective surface energy to accomplish this process is defined as fracture energy by Palmer and Rice (1973) and it equals the shaded area in the figure

$$G = \int_{D_0}^{D_c} [\tau(D) - \tau_r] dD \qquad (2)$$

We can see here that $\tau(D)$ shows a constitutive relation between the shear stress (τ) and the ongoing displacement (D), where D_0 is the slip displacement at which the curve of the slip displacement and the shear stress intersect at $\tau = \tau_r$. This constitutive relation governs the stability or instability of frictional slip failure present in the breakdown zone right behind the propagating rupture front. Ohnaka (2003) defines this breakdown zone as the zone behind the rupture front where progressive degradation of the shear strength occurs as the ongoing slip progresses. From Figure 1, it can be observed that the slip displacement D_c, the yield stress (τ_p) and the stress drop $\Delta\tau_b$ are crucial in the determination of the constitutive property for the shear rupture at any given environment (Ohnaka, 2003). Also, it is worth noting that slip-weakening is the prevailing constitutive behavior during dynamic rupture (Guatteri and Spudich, 2000).

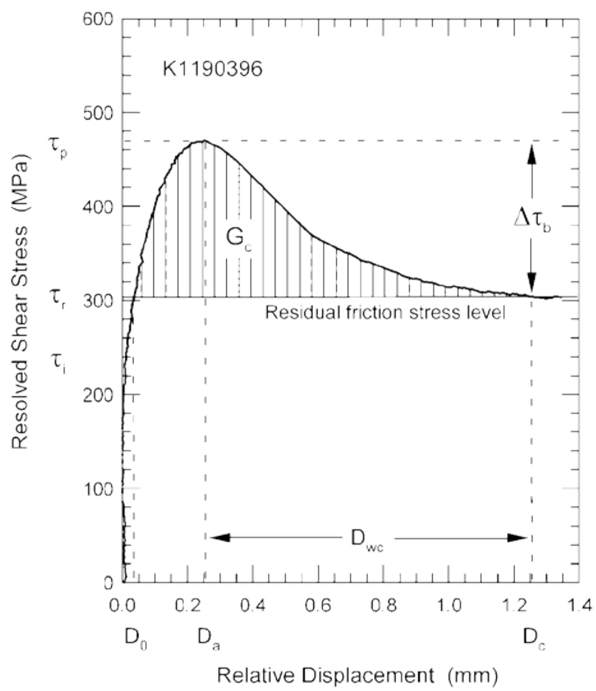

Figure 1. Slip-dependent constitutive relation observed for the shear fracture of intact rock (see text for more details). From Ohnaka (2003).

Guatteri and Spudich (2000) state that strength excess and stress drop are not material properties although slip-weakening distance may be. This concept derives from the assumption that the former are defined with respect to some initial stress level. Lab experiments performed by Okubo and Dieterich (1986) on homogeneous rock samples support this statement because both stresses vary from place to place on the studied samples.

Another concept worth noting is that fracture surfaces, especially those of intact granite, exhibit self-similarity at a finite scale range due to the smoothing the slipping process performs on geometric irregularities of the rupturing surfaces (Ohnaka 2003) (Figure 2). Since these fracture surfaces are scale-invariant, we can define the characteristic length λ_c for a shear fracture surface as the critical wavelength beyond which geometric irregularity of the surface loses the self-similarity. λ_c then, represents a predominant wavelength component of geometric irregularity of the rupture surface (Ohnaka 2003). Onhaka and Shen (1999) performed a series of high-resolution laboratory experiments. They found that the duration and the size of the nucleation zone of a frictional slip failure event were much longer when such event occurred on a fault with large λ_c (rough surfaces) than those of the event on a fault with smaller λ_c (smoother surfaces). Their findings indicate that both time-scales and size of the nucleation show a strong dependence on λ_c. Since the slip-weakening process is severely affected by geometric irregularities of the fault surfaces, we can expect that the slip displacement (D_c) scales with λ_c.

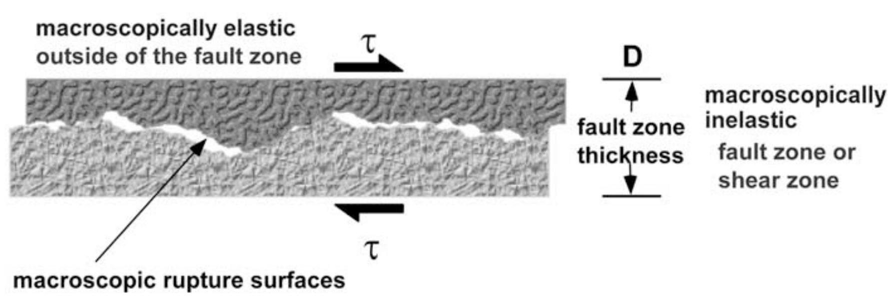

Figure 2. Geometric irregularities in the rupturing surfaces (Ohnaka, 2003)

Therefore, the relation between the stress and the slip displacement, taking into consideration geometric irregularities, can be represented by a law of the form (Ohnaka, 2003)

$$\frac{\Delta\tau_b}{\tau_p} = \beta\left(\frac{D_c}{\lambda_c}\right)^M \qquad (3)$$

Where M and β are numerical constants with values of 2.26 and 1.64 respectively. From equations (1) and (3) we can see that the critical weakening displacement D_{wc} is predictable from the critical hardening displacement D_a, which implies that the preceding process of displacement-hardening can prescribe the displacement-weakening process, as mentioned earlier in this paper. Equation (3) also shows how the parameters $\Delta\tau_b$, τ_p and D_c are constrained by the geometric irregularities λ_c. Taking this constraint into account, Onhaka (2003) established a slip-dependent constitutive law for the shear rupture relating the following five parameters: τ_i, τ_p, $\Delta\tau_p$, D_a, and D_c (or D_{wc}). Of these parameters, D_a, and D_c are scale-dependent. Rewriting equation (3) we have

$$D_c = m\left(\frac{\Delta\tau_b}{\tau_p}\right)\lambda_c \qquad (4)$$

In this equation, $m\left(\frac{\Delta\tau_b}{\tau_p}\right)$ is a dimensionless parameter and it is a function of $\frac{\Delta\tau_b}{\tau_p}$, m is written as

$$m\left(\frac{\Delta\tau_b}{\tau_p}\right) = \left(\frac{1}{\beta}\right)^{1/M}\left(\frac{\Delta\tau_b}{\tau_p}\right)^{1/M} = 0.662\left(\frac{\Delta\tau_b}{\tau_p}\right)^{0.833}$$

And D_c and D_a are written as

$$D_{wc} = 0.766 \times m\left(\frac{\Delta\tau_b}{\tau_p}\right)\lambda_c$$

$$D_a = 0.234 \times m \left(\frac{\Delta\tau_b}{\tau_p}\right)\lambda_c$$

Showing that the constitutive scaling law plays an important role in appropriately scaling physical quantities involved in the rupture process that are scale dependent (Ohnaka, 2003).

3. SLOW EARTHQUAKES

Earthquakes relax accumulated elastic tectonic stresses and strain in a fashion which strain energy stored for a long period of time, in a locked fault, is released within seconds or minutes as the fault slips fast enough to radiate seismic waves with broad spectral content. In contrast, it is known that some faults, or portions of faults, relieve stored stresses at durations ranging from hours to days (Kanamori and Hauksson, 1992; Linde et al., 1996; Crescentini et al., 1999; McGuire and Jordan, 2000) that the inertial forces involved in dynamic rupture are negligible and little or no seismic wave energy is radiated. The slippage is then “quasi-static” (Gomberg et al., 2010). Slow earthquakes not only suggest that faults can sustain ruptures over a wide range of time scales but also that the slowness can be in the velocity of the rupture propagation, in a long rise time, that is, a long time for slip to achieve 63% of its peak value (low slip rate), or both Crescentini, et al., 1999).

The size of a deformation event can be determined by a characteristic length λ_c and a characteristic duration t_c. The ratio of these dimensions defines a characteristic rupture velocity (Beroza and Jordan, 1990)

$$v_c = \frac{\lambda_c}{t_c}$$

Ordinary earthquakes propagate close to shear-wave velocities, whereas other less common events have characteristic velocities orders of magnitude less than fast elastodynamic ruptures. Fault slip events are classified as slow earthquakes when the ratio of low-frequency to high-frequency radiated energy is considerably and anomalously large (McGuire and Jordan, 2000). In other words, the spectra of slow-slip seismic signals contain less high-frequency energy and more low-frequency energy (Gomberg et al., 2010). The preferential radiation of low-energy results from a smoother than normal moment rate function, generally ascribed to the low particle slip velocity or rupture velocity of the event (McGuire and Jordan, 2000). It has also been observed that slow events represent the superposition of an ordinary, fast earthquake, radiating energy at high-frequencies, and a smooth transient of longer duration, which dominates at low frequencies (Jordan, 1991).

3.1 Non- volcanic tremor related to slow-slip events

Due to their frequency content, slow earthquakes can be studied using near-field strainmeters, geodetic data and global seismic networks (Beroza and Jordan, 1990). The most common non-earthquake signals that accompany the occurrence of geodetically observed slow slip are termed “non-volcanic tremor”. This type of pulsating signal is quasi-continuous and displays frequencies between 1 – 15 Hz (Figure 3). Such signals have been reported in the forearc region of Japan and on the deeper part of the Cascadia subduction zone interface (Rogers and Dragert, 2003). Recent studies have found that tremor tracks the propagation of slip on the plate interface and therefore, are termed by some researchers as episodic tremor and slip (ETS) (Rogers and Dragert, 2003). Tremor patterns show that slip fronts within the plate interface propagate along the strike of the subduction zone, triggering seismic radiation along an irregular surface (Ghosh et al., 2010). Some studies have suggested that large-scale slow-slip events represent the coherent superposition of slips that radiate tremor and other seismic slow-slip signals. In addition, some researchers propose that fluid pressure increase may reduce the effective normal stress and enable slip of the plate interface. This increases permeability and allows pressurized fluid to escape, contributing to the tremor signal already in progress (Shelley, et al., 2006). Also, near lithostatic pore fluid pressures may reduce the effective normal stress on source faults that exhibit slow-slip behaviors (Audet et al., 2009).

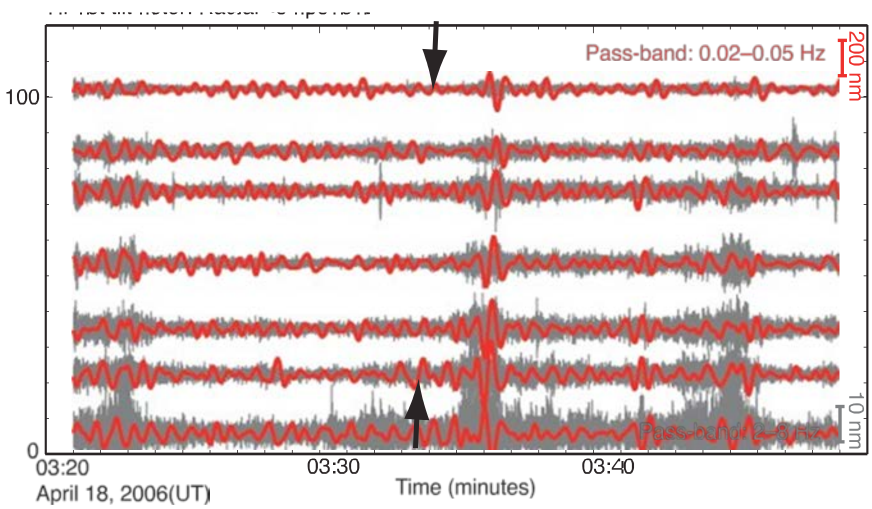

Figure 3. Example of non-volcanic tremor containing low-frequency events (shaded areas in 3b) (Ito, et al., 2007)

3.2 The Cascadia subduction zone

Slow-slip phenomena have been observed in the Cascadia subduction zone, which is the region affected by the convergence of the Juan de Fuca, Explorer, Gorda and North American plates (Figure 4). The observed phenomena are related to manifestations of a significant mode of fault slip that occurs down dip the locked zone. These events have practical implications for predicting recurrence, location and size of future earthquakes in the region.

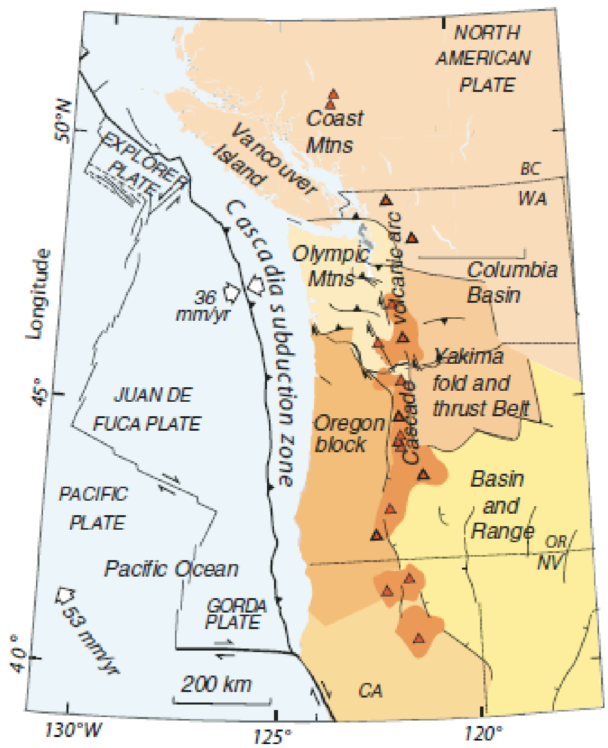

Figure 4. The Cascadia Subduction Zone (Gomberg, 2010)

Slow earthquakes in Cascadia have been recorded by seismic and GPS stations. Slip models derived from the GPS data show that these events occur between 30 and 40 km on the plate interface. Also, studies have revealed a zone of the interface between the North American and Juan de Fuca plates where 50% of the relative plate motion is accommodated by a series of transient slip events (Chapman and Melbourne, 2009). The aseismic slip and seismic generated sources, exhibit patterns that indicate slow-slip fronts that propagate across a heterogeneous surface (Gosh, et al. 2010).

Researchers have found a deep segment on the subducting plate where a stick-slip behavior occurs at much shorter scales. This segment is referred to as the ETS (Episodic Tremor and Slip) zone. The episodes of deep slip are accompanied by seismic tremors and they persist for the duration of the slip events (Figure 5). The importance of the ETS relies in the definition of the boundary of the locked zone that will rupture during the next great earthquake. Another behavior worth noting is that each ETS adds some stress on the locked portion of the subduction zone. Therefore, a great earthquake is more likely to be triggered by these ETS. Chapman and Melbourne have found that the locked zone and future great earthquakes may extend ~60 Km farther inland in Washington. In Southernmost Oregon, the transition from locked to decoupled zone occurs at ~30 km (Burgette et al., 2009)

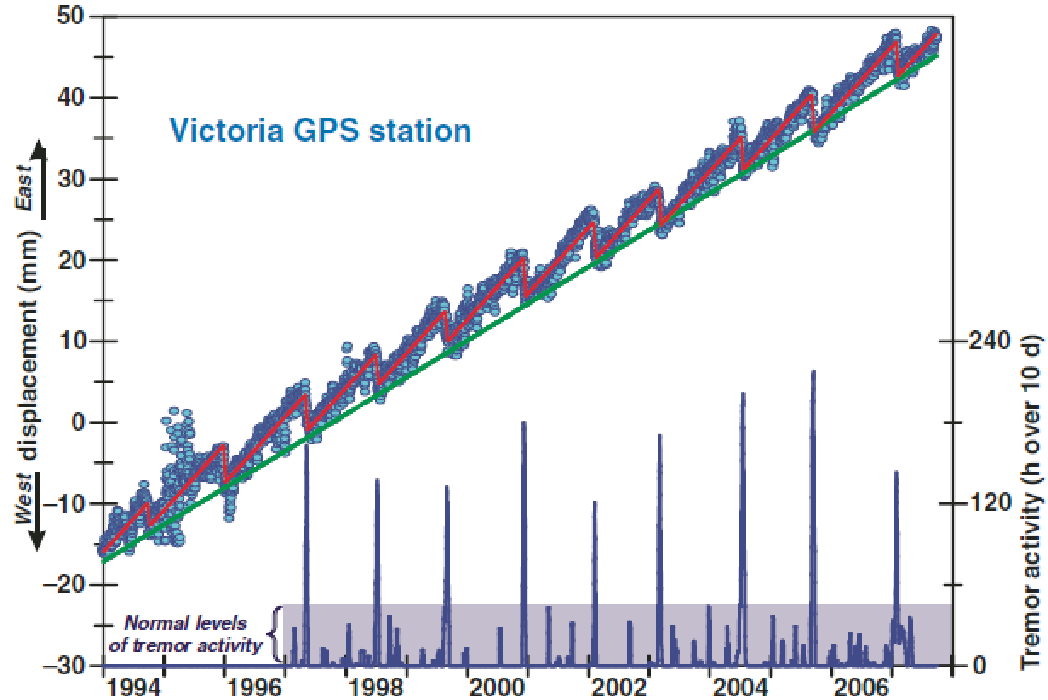

Figure 5. Episodic and Tremor Slip (ETS) (Gomberg, 2010)

Finally, some ETS have been found also in southern Cascadia, with tremor epicenters occurring in a narrow band similar to that in the northern Cascadia. ETS activity detected from 2007 to 2008 in southern Cascadia occurred synchronously with that observed in northern Cascadia (Gomberg, 2010).

4. CONCLUSIONS

The study of slow earthquakes helps to better understand faults and subduction processes. Locations and characteristics of slow earthquakes provide new insights into the behavior of ordinary or “fast” earthquakes. These events are important constituents of hazard mitigation since the location of tremor associated to slow-slips can be used to determine the extension of the locked zone and estimates of future great earthquakes in the area of study.

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