Earth and Planetary Science Letters 482 (2018) 81–92

Contents lists available at ScienceDirect

Earth and Planetary Science Letters

www.elsevier.com/locate/epsl

Seismic behaviour of mountain belts controlled by plate convergence rate

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ARTICLE INFO

Article history:
Received 17 February 2017
Received in revised form 16 October 2017
Accepted 26 October 2017
Available online xxxx
Editor: R. Bendick

Keywords:
seismicity
mountain belts
convergence rate
Gutenberg–Richter law
numerical modelling

ABSTRACT

The relative contribution of tectonic and kinematic processes to seismic behaviour of mountain belts is still controversial. To understand the partitioning between these processes we developed a model that simulates both tectonic and seismic processes in a continental collision setting. These 2D seismo-thermo-mechanical (STM) models obtain a Gutenberg–Richter frequency–magnitude distribution due to spontaneous events occurring throughout the orogen. Our simulations suggest that both the corresponding slope (b value) and maximum earthquake magnitude ($M_{W\,max}$) correlate linearly with plate convergence rate. By analyzing 1D rheological profiles and isotherm depths we demonstrate that plate convergence rate controls the brittle strength through a rheological feedback with temperature and strain rate. Faster convergence leads to cooler temperatures and also results in more larger seismogenic domains, thereby increasing both $M_{W\,max}$ and the relative number of large earthquakes (decreasing b value). This mechanism also predicts a more seismogenic lower crust, which is confirmed by a transition from uni- to bi-modal hypocentre depth distributions in our models. This transition and a linear relation between convergence rate and b value and $M_{W\,max}$ is supported by our comparison of earthquakes recorded across the Alps, Apennines, Zagros and Himalaya. These results imply that deformation in the Alps occurs in a more ductile manner compared to the Himalayas, thereby reducing its seismic hazard. Furthermore, a second set of experiments with higher temperature and different orogenic architecture shows the same linear relation with convergence rate, suggesting that large-scale tectonic structure plays a subordinate role. We thus propose that plate convergence rate, which also controls the average differential stress of the orogen and its linear relation to the b value, is the first-order parameter controlling seismic hazard of mountain belts.

1. Introduction

Mountain building is the most spectacular manifestation of continental dynamics. The fact that mountain ranges are able to maintain their topography over tens of millions of years, is the evidence of how tectonic forces deform the continental crust (Avouac, 2015). Frequent earthquakes are the physical response of such crustal deformation, driven by convergence of continental plates.

Mitigating the impact of these earthquakes in mountain belts is crucial as part of these regions are densely populated. Moreover, understanding the partitioning of seismic and aseismic fault slip is central to seismotectonics, as it ultimately determines the seismic potential of faults. Although we cannot predict when and where the next earthquake will occur, we may forecast the maximum credible size and frequency of future events using the Gutenberg–Richter law (Gutenberg and Richter, 1944). This power law describes the frequency–magnitude distribution of earthquakes and its slope, the so-called b value, is commonly used to describe the relative occurrence of small and large events (e.g., Schorlemmer et al., 2005). From a physical mechanism perspective, local spatial variability of b and corresponding seismic hazard have been interpreted to relate to differences in the rheology of the crust (e.g., Pasquale et al., 2010), tectonic regime (e.g., Gulia and Wiemer, 2010; Doglioni et al., 2015), differential stress (e.g., Spada et al., 2013; Scholz, 2015), style-of-faulting (e.g., Schorlemmer et al., 2005), and degree of heterogeneity (e.g., Singh et al., 2009). On a larger regional scale, studies over the past decades have suggested that variations in seismicity of the mountain chains are related to variations in geologic structure (e.g., Bollinger et al., 2004; Grenerczy et al., 2005; Tatar et al., 2002). However, the preeminent physical parameter in controlling the seismic behaviour in orogenic mountain belts is still unclear.

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https://doi.org/10.1016/j.epsl.2017.10.053

Mountain building processes acting on a time-scale of millions of years have been widely studied both using numerical models (e.g., Faccenda et al., 2008) and analogue experiments (e.g., Davis et al., 1983). These models have been used to investigate the effect of different physical characteristics on the evolving geometry and dynamics of a collisional margin. However, because of the relatively large computational time step (Δ_t =1000 yr), transient features and short-term processes, such as earthquakes, are not included. On the other hand, dynamic rupture simulations physically consistently model the propagation and arrest of a rupture for a set of unknown initial conditions—such as the stresses—and strength and the fault geometry (e.g., Dalguer, 2012). Alternatively, seismic cycle models simulate cycles of recurring interseismic and coseismic periods, either by kinematically defining slip or stress drop or by dramatically allowing rupture nucleation and propagation (e.g., Lapusta and Barbot, 2012). However, these models can only assume planar faults and they not account for anelastic deformation. Consequently, such models cannot address the complex fold-and-thrust belt structures and the long-term mountain building processes. Therefore, to improve our physical understanding of how mountain belts behave in terms of seismicity, we need an orogen-scale approach that takes into account the evolution of complex mountain belt architectures from the long-term tectonic history.

To overcome these modelling limitations we use a physics-based self-consistent seismo-thermo-mechanical (STM) modelling approach, which captures the dynamics of both long-term orogen-scale deformation and short-term seismic behaviour. We analyze results of a series of numerical models of continent–continent collision where different plate convergence rates are systematically tested to study its effects on seismicity style. We characterise this style by analysing the Gutenberg–Richter distribution and corresponding b value obtained from the synthetic catalogue of each numerical simulation. To understand these results we analyze the rheological feedback between plate convergence rate and the brittle strength of crustal rocks. Then, we verify our numerical hypothesis calculating similar quantities for natural examples of collisional orogens (i.e., Alps, Apennines, Zagros and Himalaya).

2. Methods

2.1. Seismo-thermo-mechanical modelling

The employed numerical STM code (van Dinther et al., 2013b) combines conservative finite differences on a fully staggered grid and marker-in-cell techniques (Gerya and Yuen, 2003). The momentum, mass and heat conservation equations are implicitly solved using a visco-elasto-plastic rheology (Gerya and Yuen, 2007) on the non-deforming Eulerian grid. The advection of physical properties including viscosity, plastic strain and temperature is performed with the displacement of Lagrangian markers. The momentum equations include the inertial term to stabilize high-coseismic slip rates at small time steps. The employed visco-elasto-plastic rheology is based on a constitutive relationship between deviatoric stresses and strain rates $\dot{\varepsilon}_{ij}$ applying linear elasticity and non-Newtonian viscosity (Gerya and Yuen, 2007)

$$\dot{\varepsilon}_{ij} = \frac{1}{2G}\frac{D\sigma'_{ij}}{Dt} + \frac{1}{2\eta}\sigma'_{ij} + \begin{cases} 0 & \text{for } \sigma'_{II} < \sigma_{yield} \\ \chi\frac{\partial\sigma'_{II}}{\partial\sigma'_{ij}} = \chi\frac{\partial\sigma'_{ij}}{2\sigma'_{II}} & \text{for } \sigma'_{II} = \sigma_{yield} \end{cases} \quad (1)$$

where G is shear modulus and η is effective viscosity. $D\sigma'_{ij}/Dt$ is the objective co-rotational time derivative solved using a time explicit scheme and $\sigma'_{II} = \sqrt{{\sigma'_{xx}}^2 + {\sigma'_{xz}}^2}$ is the square root of the second invariant of the deviatoric stress tensor, and χ is a plastic multiplier connecting plastic strain rates and stresses (Gerya and Yuen, 2007). Viscous creep is computed in terms of deformation invariants and depends on strain rate, temperature, and pressure (Ranalli, 1995). The viscous component of the deformation is then calculated as a combination of diffusion and dislocation creep. Plastic behaviour is taken into account assuming a non-associative Drucker–Prager yield criterion. Evaluated at each Lagrangian marker, plasticity sets in when the square root of the second invariant of the deviatoric stress tensor reaches the local pressure-dependent yield strength ($\sigma'_{II} = \sigma_{yield}$)

$$\sigma_{yield} = C + \mu_{eff}\ P, \quad (2)$$

where C is the cohesion. An important component in the yield criterion is the friction coefficient. Following the approach in van Dinther et al. (2013a), we apply a strongly rate-dependent friction formulation (e.g., Ampuero and Ben-Zion, 2008) in which the effective friction coefficient μ_{eff} depends on the visco-plastic slip velocity $V = (\sigma_{yield}/\eta_m)\Delta x$. Here η_m is the local viscosity, σ_{yield}/η_m is the square root of the second invariant of the visco-plastic strain rate ($\dot{\varepsilon}_{II}$) and Δx is the Eulerian grid size. This allows us to calculate the effective steady-state friction (μ_{eff}) coefficient at every marker as a function of slip rate V as

$$\mu_{eff} = \mu_s(1-\gamma) + \mu_s\frac{\gamma}{1+\frac{V}{V_c}} \quad (3)$$

$$\gamma = 1 - (\mu_d/\mu_s) \quad (4)$$

where μ_s and μ_d are static and dynamic friction coefficients, respectively. V_c is the characteristic velocity at which half of the friction change has occurred, and γ represents the amount of slip velocity-induced weakening if γ is positive, or strengthening if γ is negative. For all lithologies, the velocity weakening frictional formulation is parameterized using $\gamma = 0.7$ and $V_c = 4.4$ cm/yr, following van Dinther et al. (2013b). Full details of this method are provided in Gerya and Yuen (2007) and van Dinther et al. (2013b).

2.2. Model setup

The initial 2D model setup consists in a 4000 × 250 km computational domain (Fig. 1). Two 1700-km-long continental plates were separated by a 600-km-long oceanic plate. Both continental plates and an oceanic are composed of an upper crust, lower crust and lithospheric mantle. The visco-elasto-plastic thermomechanical parameters of these lithologies are based on a range of laboratory experiments (Table 1). The models use a grid resolution of 2041 × 401 nodes with variable grid spacing. This allow a minimum grid resolution of 400 m in the area subjected to largest deformation. More than 28 million Lagrangian markers carrying material properties were used in each experiment. The model has free slip boundary conditions and a permeable lower boundary with an infinity-like external free slip and external constant temperature conditions at 1670 km depth (Burg and Gerya, 2005). The free surface boundary condition atop the crust is implemented by using a “sticky air” layer (Schmeling et al., 2008) with low density (1 kg/m^3) and viscosity (10^{17} Pa s). The model is kinematically driven using internal boundary conditions located at 1000 and 3000 km (Fig. 1), which allow to impose an initial convergence rate of 10 cm yr^{-1}. The thermal structure of the oceanic lithosphere was calculated from the half space cooling model (Turcotte and Schubert, 2002) for a given plate age of 70 Myr (Fig. 1b). The initial continental geotherm was set using values of 0 °C at the surface and 1300 °C at 142 km depth. The thermal gradient used within the mantle was quasi-adiabatic (0.5 °C/km). Gravitational acceleration of 9.81 m s^{-2} was used in the model. To initiate subduction a weak zone was imposed on the right ocean–continent transition. The weak zone cuts through the whole lithospheric mantle with an angle of ∼30° and is characterised by weak plastic strength (1 MPa) and wet olivine rheological parameters (Ranalli, 1995).