The seismic cycle at subduction thrusts: 1. Insights from laboratory models

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[1] Subduction megathrust earthquakes occur at the interface between the subducting and overriding plates. These hazardous phenomena are only partially understood because of the absence of direct observations, the restriction of the instrumental seismic record to the past century, and the limited resolution/completeness of historical to geological archives. To overcome these restrictions, modeling has become a key-tool to study megathrust earthquakes. We present a novel model to investigate the seismic cycle at subduction thrusts using complementary analog (paper 1) and numerical (paper 2) approaches. Here we introduce a simple scaled gelatin-on-sandpaper setup including realistic tectonic loading, spontaneous rupture nucleation, and viscoelastic response of the lithosphere. Particle image velocimetry allows to derive model deformation and earthquake source parameters. Analog earthquakes are characterized by “quasi-periodic” recurrence. Consistent with elastic theory, the interseismic stage shows rearward motion, subsidence in the outer wedge and uplift of the “coastal area” as a response of locked plate interface at shallow depth. The coseismic stage exhibits order of magnitude higher velocities and reversal of the interseismic deformation pattern in the seaward direction, subsidence of the coastal area, and uplift in the outer wedge. Like natural earthquakes, analog earthquakes generally nucleate in the deeper portion of the rupture area and preferentially propagate upward in a crack-like fashion. Scaled rupture width-slip proportionality and seismic moment-duration scaling verifies dynamic similarities with earthquakes. Experimental repeatability is statistically verified. Comparing analog results with natural observations, we conclude that this technique is suitable for investigating the parameter space influencing the subduction interplate seismic cycle.


1. Introduction

[2] The majority of global seismic moment occurs at convergent margins. In particular, most of $M \geq 8.0$ earthquakes are generated along the gently dipping fault plane (i.e., the subduction megathrust, Figure 1) representing the frictional interface between subducting and overriding plates [e.g., Scholz, 1990; Pacheco and Sykes, 1992; Dixon and Moore, 2007; McCaffrey, 2007, 2008]. These earthquakes are associated with the ongoing downdip motion of the subducting lithosphere, resulting in the elastic deformation of the adjacent plates and the transfer of stress across the locked part of the subduction thrust into the upper plate. If the accumulated stress exceeds the strength of the plate interface, an abrupt fault slip occurs, and part of the stored elastic energy radiates as seismic waves. The entire process, including the stress accumulation phase, the relief by the earthquake itself and postseismic relaxation processes, is known as the “subduction seismic cycle” [e.g., Wang, 2007]. In this context, the word “cycle” does not imply any periodicity of characteristic recurrence interval or earthquake size, but rather a repeating spatio-temporal pattern. Methods of studying the subduction seismic cycle include remote sensing, geophysical, and geological observational studies of convergent margins [e.g., Gagnon et al., 2005; Grapenthin and Freymueller, 2011] as well as numerical [e.g., Wang, 2007; Kaneko et al., 2010] and analog modeling [e.g., Rosenau et al., 2009].

[3] Although the general stages of the subduction seismic cycle are known, many critical points remain obscure. For example, the maximum earthquake magnitude recorded at
Figure 1. Schematic cross-section of a subduction zone.

different subduction zones in the instrumental and historical records is highly variable [e.g., McCaffrey, 2008; Heuret et al., 2011], highlighting the possibility of a variable propensity for hosting mega-events [Marzocchi et al., 2011]. Some subduction zones exhibit high seismic moment release rate (e.g., South Chile or East Alaska), whereas it is very low at others [e.g., East Aegean; Hyndman, 2007]. Although the subduction velocity seems to be one of the most important controlling factors for interplate seismicity [e.g., Bird et al., 2009; Heuret et al., 2011], the mechanisms governing the frequency and size of megathrust earthquakes are not yet clear [e.g., McCaffrey, 2008]. The amount of interplate locking varies both regionally and locally [e.g., Moreno et al., 2010; Freymueller et al., 2008] as well as potentially temporarily, and identifying the parameters controlling this variability is still a challenge.

Our knowledge of the parameters that control the subduction seismic cycle and the maximum seismic energy release is limited due to the following constraints: (a) the seismogenic zone behavior can only be studied using indirect methods, because the fault is buried and located mainly offshore. This provides only a snapshot of a long-term, time-dependent process; (b) the pattern of strain accumulation along the subduction thrust may not be constant throughout the seismic cycle; (c) instrumental seismic records of the past several decades do not provide enough temporal perspective on the recurrence of major earthquakes [e.g., Satake and Atwater, 2007; McCaffrey, 2008]. The problem is not solved by the use of historical/geological records, which lack completeness and resolution; and (d) our understanding of the behavior of the subduction thrust fault is based on a patchwork of merged observations at different convergent margins that are presently at different phases of the seismic cycle, causing uncertainties.

Despite unavoidable assumptions that limit the applicability field of results, analog and numerical modeling may provide important insights into small-scale/short-term and large-scale/long-term features of earthquake physics. Examples include investigations on the physical processes governing dynamic rupture nucleation, growth, and arrest [e.g., Ben David et al., 2010; Nielsen et al., 2010], the factors controlling seismogenic slip at a point on a fault [e.g., Xia et al., 2004; Lykotrafitis et al., 2006], and recurrence behavior [e.g., Rubinstein et al., 2012a]. Analog models with realistic geometries (i.e., wedge-shaped) suitable for studying subduction interplate earthquakes fall into the following categories: (a) purely elastic foam-rubber experiments [e.g., Brune, 1996] and (b) granular elasto-plastic models [e.g., Rosenau et al., 2009].

Foam-rubber models of shallow-angle thrust faulting were designed for investigating the basic physics governing the sliding along frictional interfaces during the coseismic stage in the presence of a free surface [Anooshehpoor and Brune, 1994, 1999; Brune and Anooshehpoor, 1997; Brune et al., 1993]. In these models the slow wave speed of the material allows the process of frictional sliding to be captured without high-speed photography. However, these models are affected by unavoidable edge effects and nonlinearities in the rheological response, inhibiting the scalability of the experimental results to nature [Rosakis et al., 2007].

The first scaled laboratory models of subduction megathrust earthquakes have been realized by Rosenau et al. [2009], who used granular analog materials (i.e., a mixture of rice, rubber pellets, and sugar) with frictional and elastic properties that are consistently scaled from the natural prototype. Because of the capability of the model to simulate elastic and plastic deformation, this approach bridges the short-term seismic cycle and the long-term tectonic evolution of convergent margins [Rosenau et al., 2009] and allows to study the subduction forearc tectonic evolution over multiple seismic cycles and inherent feedback [Rosenau and Oncken, 2009]. The ability to experimentally simulate long time-series of self-consistent seismic cycle deformation (i.e., hundreds of cycles) allowed to further explore scaling of earthquake effects like tsunamis in a physics-based probabilistic manner [Rosenau et al., 2010]. Because of lack of temporal resolution in monitoring the experiments, however, these models were not suitable to investigate the dynamics of single ruptures. Thus, an analog model that combines the features of the subduction seismic cycle with experimental visualizations of rupture propagation along frictional interfaces is still missing. This type of model would be beneficial in the light of new observational techniques in nature and the occurrence of several recent mega-thrust events. These events delivered prime examples (i.e., rupture time-series of unprecedented quality) that need to be explained by new theories and experiments.

In this work, we present a new experimental technique to study the behavior of multiple subduction earthquake cycles. This technique is part of a larger modeling attempt that takes advantage of complementary analog (this paper) and numerical modeling approaches [van Dinther et al., 2013, paper 2]. The main advantage of analog modeling lies in the fact that a properly constructed model evolves in response to the physics of the system under the applied experimental conditions. However, laboratory experiments face technical limitations in monitoring and repeatability, in the limited availability of analog materials to reproduce complex rheologies, and in the exact control of the boundary conditions. In contrast, numerical models have the advantage of a straightforward quantitative approach and a considerable flexibility in selecting geometry, material and system parameters as well as their ability to precisely define boundary conditions. However, large computational capacities are necessary to simulate realistic geological features using complex rheologies and/or three-dimensional settings. In particular, such models (both numerical and analog) should be considered as simplified representations of nature for the difficulty in understanding and simulating the frictional behavior. The aim of our combined approach is to simulate multiple subduction interplate earthquake cycles with a simple analog setup (e.g., neglecting temperature, off-fault plasticity, fluid pressure, sediment metamorphism, slab elasto-plastic deformation), and validating it with a numerical model.
Here we introduce the novel analog modeling approach to reproduce the subduction interplate earthquake cycle, including realistic tectonic loading, spontaneous nucleation of the frictional instabilities (i.e., analog earthquakes), and viscoelastic relaxation of the lithosphere. We first describe the laboratory methods adopted to simulate subduction thrust earthquakes, including scaling criteria, material properties, and experimental monitoring. We then present the general characteristics of a reference experiment supported by a series of eight additional models. We provide measurements of analog earthquake source parameters in relation to natural events, aiming to demonstrate that the model is able to reproduce features of the interplate seismic cycle that are common to many convergent margins. Cyclic, deformation during the seismic cycle, faulting parameters, rupture behavior, and seismic moment-duration scaling are analyzed in detail by comparing experimental results with natural data. Finally, we include details on the variability and repeatability of the experimental results to strengthen the reliability of the novel approach.

2. Experimental Approach

In this study, we selected a viscoelastic material, Pig Skin 2.5 wt% gelatin, as a rock analog because it has the following qualities: (a) proper rheological properties to simulate downscaled lithospheric rock behavior under laboratory conditions in a natural gravity field [Di Giuseppe et al., 2009], and (b) the advantage of being transparent, enabling the tracking of passive tracers included in the media and allowing nonintrusive, continuous monitoring of the internal velocity field during the evolution of the model. We use gelatin-on-sandpaper and gelatin-on-plastic interfaces to simulate velocity-weakening and velocity-strengthening frictional behavior of the subduction thrust fault, respectively. Monitoring of the experiment is performed using the Particle Image Velocimetry (PIV) technique [e.g., Adrian, 1991], and the derivation of source parameters is done using an ad hoc algorithm.

2.1. Rheological Properties of the Analog Materials

To simulate the viscoelastic behavior of the lithosphere, we selected Pig Skin 2.5 wt% after an extensive study of the rheological and physical properties of a wide range of materials. Pig Skin 2.5 wt% has the right properties to scale for length, density, stress, and viscosity of lithospheric rocks in a natural gravity field [Di Giuseppe et al., 2009]. In particular, this material satisfies two necessary conditions for a viscoelastic solid: (1) \( G' \) and \( G'' \), which are standard rheological quantities defining the stored and dissipated energy, respectively [e.g., Nelson and Dealy, 1993], are approximately the same order of magnitude (i.e., the elastic deformation counterbalances the viscous deformation); and (2) \( G' > G'' \) (see Di Giuseppe et al. [2009] for a detailed discussion). Pig Skin gelatin is prepared by preliminary heating and stirring a 2.5 wt% distilled water solution at 60 °C and cooling it at 10 °C for 12 h [Di Giuseppe et al., 2009] to obtain a firm texture. The rheology of the material has been characterized using a 50 mm parallel-plate rheometer (Anton Paar MCR 301) equipped with a Peltier element to control the temperature during the measurements. To avoid possible sample dehydration problems during long measurements, the Peltier element was saturated using a custom water bubbler.

Gel state (i.e., solid-like) Pig Skin 2.5 wt% gelatin has a shear modulus ranging from \( 10^2 \) to \( 10^4 \) Pa, depending on the age of the material (Figure 2a). Constraining the viscosity of Pig Skin gelatin in its gel state is more difficult when using classical rheometric techniques. However, frequency sweep tests conducted at low frequencies provide a lower bound of \( 7 \cdot 10^7 \) Pa · s for viscosity.

2.2. Frictional Properties

The frictional properties of gelatin-on-sandpaper (i.e., an analog of the seismogenic zone of the subduction thrust) and gelatin-on-plastic (i.e., an analog of the updip and downdip aseismic parts of the subduction thrust) have
been investigated using a linear spring-block-like device in which a digital dynamometer (AFG-Mecmesin) continuously measures the shear force exerted during the sliding of a gelatin sample [Corbi et al., 2011].

Here we recall the main frictional properties, including the static friction coefficient, \( \mu_s \), and the friction rate parameter, \( a-b \), which are related to the potential seismic moment release of a given interface. \( \mu_s \) is determined by measuring the amount of force that resists motion of the gelatin sample under an applied normal pressure (i.e., from the slope of the regression line between shear and normal stresses, normalized by the area of the sample on which these forces act).

The \( a-b \) parameter is defined as follows:

\[
a - b = d\mu_s / d[\ln(V)]
\]

where \( \mu_s \) is the steady state friction coefficient and \( V \) is the sliding velocity. Negative values of \( a-b \) are indicative of velocity-weakening behavior (i.e., ruptures may nucleate and can easily propagate along the interface), whereas positive values of \( a-b \) characterize a velocity-strengthening behavior (i.e., no rupture nucleation, and inhibited rupture propagation).

Sandpaper was selected from a range of materials studied in an extensive tribological investigation focused on roughness, normal pressure, and loading rate dependence of friction dynamics [Corbi et al., 2011]. Interface roughness has been quantified in terms of amplitude, \( R_{mh} \), and wavelength, \( \lambda \), of geometrical perturbations with respect to a reference baseline. The ranges of investigated roughness are 0.10–0.01 mm and 0.71–0.03 mm for the amplitudes and wavelengths of protrusions, respectively.

Because the static friction depends on the real contact surface, we observed the highest static friction with the smoothest sandpapers [e.g., Marone and Cox, 1994], according to the adhesion frictional model proposed by Bowden and Tabor [1950].

Frictional behavior is primarily controlled by the roughness of the interface. Regular stick-slip instabilities occur only for contact surfaces with \( R_{mh} > 0.018 \) mm. Slip can nucleate and easily propagates (i.e., \( a-b < 0 \)) for \( R_{mh} \) values of 0.03–0.100 mm; otherwise, the propagation is rapidly inhibited (i.e., \( a-b > 0 \)). The dynamic regime switches to stable sliding and velocity-strengthening behavior for \( R_{mh} 0.03 \) mm. A nonlinearity of \( a-b \) exists because the amplitude and wavelength of protrusions act as two competing parameters. The former seems to tune the magnitude of the slip event [e.g., Ruff and Kanamori, 1983], and the latter controls the capability to transmit the triggered deformation along the rupture plane and is linearly related to the critical slip distance [Ohnaka, 2003]. Pressure and driving rate seem to play secondary roles, although they are not negligible. Independent of the roughness of the contact surface, stable sliding is always achieved for the lowest normal loads and the highest sliding velocities [Corbi et al., 2011].

Stable slip at low normal load appears to be consistent with a conditionally stable regime according to rate and state friction framework. The effect of sliding velocity is probably linked to the finite lifetime of the adhesive bonds and their potential to restick after the occurrence of a slip pulse [Baumberger et al., 2002].

In particular, the sandpaper used in this study is characterized by protrusion amplitudes of 0.038 ± 0.004 mm and interpeak distances of 0.143 ± 0.009 mm. The gelatin-on-sandpaper system has \( \mu_s \) of 0.19 ± 0.05. The system shows stick-slip dynamics for a loading rate of 0.6 cm/min [Corbi et al., 2011], and a velocity-weakening behavior (\( a-b = -0.028 \)) has been observed (Figure 2b) [Corbi et al., 2011].

The friction of gelatin-on-plastic is characterized by stable sliding for all tested experimental conditions (i.e., a loading rate of 0.2 to 2 cm/min and normal pressures of 381–991 Pa). \( \mu_s \) for this system is 0.02 ± 0.01, and \( a-b \) is 0.027 (Figure 2b).

### 2.3. Scaling

The scaling procedure is crucial for designing an analog model that simulates geodynamical processes on a convenient geometric and temporal scale. According to similarity theory, the scaled model should be geometrically, kinematically, dynamically, and rheologically similar to the natural prototype [e.g., Hubbert, 1937; Ramberg, 1981; Weijermars and Schmeling, 1986]. We identify the most important physical quantities of the investigated system (i.e., length, velocity, and force) and provide a dimensionless number (i.e., the scaling factor) for each parameter that represents the ratio and tuning between the model and natural quantities (Table 1).

The wedge model is designed using a model/nature scale factor, \( L^a \), of 1.57 × 10^6 (i.e., 1 cm in the model corresponds to 6.4 km in nature). For models performed in a natural gravity field, the stress scaling factor, \( \sigma^a \), can be obtained according to the following relation:

<table>
<thead>
<tr>
<th>Parameters</th>
<th>Similarity</th>
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<tbody>
<tr>
<td>Quantity</td>
<td>Symbol</td>
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<tr>
<td>Model parameters</td>
<td>gravitational acceleration</td>
</tr>
<tr>
<td>length</td>
<td>( v )</td>
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<tr>
<td>time (interseismic)</td>
<td>( T_i )</td>
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<tr>
<td>time (coscinetic)</td>
<td>( T_c )</td>
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<tr>
<td>velocity (interseismic)</td>
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<td>velocity (coscinetic)</td>
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<td>stress</td>
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<td>Materials properties</td>
<td>density</td>
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<td>shear modulus</td>
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<td>viscosity</td>
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where \( \rho^* \) is the nature/model density scale factor. Because Pig Skin 2.5 wt% density is \( \approx 1000 \text{ kg/m}^3 \) [Di Giuseppe et al., 2009], we obtain a stress scale, \( \sigma^* \), of \( 5.42 \cdot 10^{-7} \) (i.e., 1 Pa in the model is equal to 1.85 MPa in nature) for an average crustal wedge density of 2900 kg/m³.

[28] The physical quantity controlling the elasticity of the analog material is the shear modulus, which has the dimension of stress and, as a result, scales with the same scaling factor, \( G^* = 5.42 \cdot 10^{-7} \). Pig skin 2.5 wt% has a shear modulus in the \( 10^3 - 10^5 \text{ Pa} \) range, depending on the aging of the material [see Di Giuseppe et al., 2009], and therefore scales to a natural prototype with \( G = 1.85 \cdot 10^9 - 1.85 \cdot 10^{10} \text{ Pa} \).

[29] The subduction seismic cycle is a complex process that includes extremely variable deformation rates. Because the experimental interseismic/coseismic time ratio is very small (see paragraphs 3.1 and 3.4 for details on interseismic and rupture duration, respectively) with respect to the nature, it is not admissible to use a single time scale that accounts for both the coseismic and interseismic stages. To overcome this limitation, we use a dual time scale (i.e., one for the coseismic stage and another for the interseismic stage), similar to Rosenau et al. [2009].

[30] The coseismic stage is characterized by an instantaneous elastic response in which inertial force may play an important role. Therefore, the coseismic time scale, \( T_{c^*} \), is set keeping the Froude number (i.e., the ratio of a body’s inertia to gravitational forces) constant, which leads to the following relation:

\[
T_{c^*} = \sqrt{L^*} \tag{3}
\]

(i.e., 1 s in the model is equal to approximately 800 s in nature).

[31] During slow deformation inertia is negligible, but viscous behavior becomes dominant, and consequently, the interseismic time scale, \( T_{i^*} \), is established using the following relation [e.g., Weijermars and Schmeling, 1986]:

\[
T_{i^*} = \eta^* / \sigma^*, \tag{4}
\]

where \( \eta^* \) is the model/nature ratio of viscosity. For a model viscosity of \( 3 \cdot 10^9 \text{ Pa} \cdot \text{s} \) and a natural viscosity of \( 5 \cdot 10^{21} \text{ Pa} \cdot \text{s} \), we obtain an interseismic time scale of \( 1.11 \cdot 10^{10} \) (i.e., 1 s in the model is equal to 286 yr in nature).

[32] Interseismic and coseismic stage velocities are derived as the ratios between \( L^* \) and \( T_{i^*} \) (i.e., \( 1 \cdot 10^{-4} \text{ m/s} \) in the model is equal to 22 cm/yr in nature) and between \( L^* \) and \( T_{c^*} \) (i.e., \( 1 \cdot 10^{-4} \text{ m/s} \) in the model is equal to 0.1 m/s in nature), respectively.

[33] The static friction coefficient, \( \mu_s \), and the friction rate parameter, \( a-b \), are dimensionless and, thus, scale with a factor of 1.

### 2.4. Experimental Setup

[34] The experimental apparatus (Figure 3) consists of a Plexiglas box that is 60 cm long, 40 cm high and 34 cm wide, in which an undeformed gelatin wedge is placed. The gelatin wedge is representative of a ~380 km long and 70 km deep section of forearc lithosphere. Along-strike (i.e., \( y \)-direction) variations of frictional properties are not imposed as in a quasi-two-dimensional setup. A rigid steel basal conveyor plate, which is the analog of the subducting plate, is driven with a velocity of \( 10^{-4} \text{ m/s} \) using a screw jack connected to a computer-controlled stepping motor. The dip of the basal plate is set to \( 10^\circ \). A rigid vertical backstop is placed at the rear of the gelatin wedge.

[35] The analog megathrust develops at the interface between the gelatin and the basal conveyor plate. The velocity-weakening seismogenic zone of the subduction thrust is simulated using sandpaper at the gelatin-plate interface and extends from 31 to 47 cm from the backstop in horizontal distance. This distance corresponds to a 100 km wide (in map view) seismogenic zone (90 to 190 km from the trench) that spans a depth range of 15 to 34 km. These dimensions are close to the average geometry extracted from worldwide subduction zone statistics [Heuret et al., 2011]. Several physical explanations for the updip and downdip limits of the seismogenic zone have been proposed. A recent hypothesis for explaining the puzzling origin of the updip limit is strain localization, in combination with precipitation of calcite and quartz [Saffer et al., 2012]. The downdip limit is likely related to the brittle-ductile transition of crustal material [i.e., Brace and Kohlstedt, 1980] or by the intersection of the slab with the forearc mantle wedge [Peacock and Hyndman, 1999]. We simulate these velocity-strengthening regions using an upper and lower velocity-strengthening plastic sheet covering the conveyor plate. The plastic sheets are attached to the base of the experimental setup to ensure...
that the depth range of the analog seismogenic zone remains stable during the experimental run.

2.5. Monitoring and Resolution

[32] A transparent analog medium seeded with highly reflective, neutrally buoyant markers and a suitable illumination system are key components for tracking deformation in analog models. We seeded Pig Skin gelatin, which has a suitable transparency in the gel state [Di Giuseppe et al., 2009], with 63 to 125 μm diameter fluorescent microspheres (Cospheric - UYVGPMs). Microspheres emit a green luminescence when excited by 365 nm wavelength light. To reduce possible buoyancy effects (the density difference between microspheres and gelatin is 0.03 g/cm³), the seeding procedure was performed just before the gelatin was cooled to 10°C. The amount of passive tracers employed was extremely small (0.004 wt%) to prevent any effect on the rheological properties of the gelatin. The setup included a 60 cm long ultraviolet (UV) lamp that lights a 2 mm thick planar volume situated in the center of the model. This results in a nonintrusive method for measuring model internal deformation that limits possible biases due to sidewall friction.

[33] The models were monitored using two CCD cameras (ALLIED-PIKE), both of which acquired a side view sequence of high-resolution images (1600 × 1200 pixels², 8 bit, 256 gray levels) digital images of the longitudinal cross-section illuminated (x-z plane; Figure 3). One camera imaged the entire model with a frame rate of 7.5 fps and was used to derive the long-term model deformation. The second camera focuses the seismogenic zone at 15 fps, aiming to increase both the spatial and temporal resolutions. The second camera has been prerequisite for deriving at a high enough resolution source-parameter and dynamic rupture imaging.

[34] Because simultaneous monitoring of the upper surface (x-y plane; Figure 3) velocity field is not possible in a single model due to interference with the UV lamp, additional model runs have been performed with illumination of the shallow horizontal plane exclusively. In such cases, we used a single camera at 7.5 fps.

[35] Images were analyzed by means of the PIV method using MatPIV, an open-source MATLAB code [Sveen 2004]. PIV is a widely used optical method for extracting velocity fields in viscous and granular flows [e.g., Adrian, 1991; Adam et al., 2005; Rosenau et al., 2009].

[36] Each image was sampled by overlapping interrogation windows, i.e., small image regions. The Fast Fourier Transform was employed to compute the displacement of each window across two consecutive frames. PIV, finally, provides the average displacement field of tracer particles within the interrogation window. An initial 128 × 128 pixel² discretization with an overlap of 50%, subsequently refined to 64 × 64 pixels², guaranteed both robust statistics within the interrogation windows and an adequate spatial resolution for both velocity components. The two displacement components in each window were converted to velocities by using the pixel size and the time interval between two consecutive frames, i.e., the inverse of frame rate.

[37] Particle image velocimetry is subject to sources of error. The overall evaluation of the accuracy is affected by the acquisition step through the techniques of velocity extraction, and the accuracy of the velocity measurement is strongly linked to the image quality [Raffel et al., 2007]. During the acquisition step, all factors contributing to accurate velocity measurements were carefully controlled. In particular, we optimized the experimental procedure for proper tracer density and homogeneity, low image background, and homogeneous lighting. The camera used for image acquisition was equipped with low-aberration optics, and the entire system was accurately calibrated. The accuracy of the subpixel estimators of PIV algorithms is on the order of 1/10 to 1/20 of a pixel. Given the camera calibration datum for the reference experiment, the resulting spatial resolution is 12 mm and accuracy is 25 μm for camera 1. A spatial resolution of 3.2 mm with a sensitivity of 7 μm is obtained adopting camera 2.

2.6. Source-Parameter Derivations

[38] Earthquake source parameters are derived by analyzing the velocity field of the entire image data set (see Figure 4 for a flow-chart of the required processing steps). First, the velocity
field was derived using the PIV method applied to the full range of sequential images. Then, we extracted the horizontal and vertical components of the velocity field along a profile 1 cm above the analog subduction thrust. This choice is forced by the necessity to exclude from the image monitoring the subducting plate motion. A new time series was successively created by selecting the profile peaks for each time (Figure 5, red line).

[39] We used an ad hoc algorithm to identify local maxima in slip velocity, hereinafter referred to as the picking algorithm (i.e., the fpeak algorithm running under MATLAB), was used to discriminate velocity peaks (i.e., analogous to seismic events) from the low-background-characteristic velocity (i.e., analogous to the interseismic stage). The algorithm requires a characteristic minimum velocity (i.e., the coseismic velocity threshold) and a chosen sensitivity (i.e., the minimum number of frames before and after velocity peaks). A histogram counting the number of frames with a given maximum velocity is used as a tradeoff curve in which the optimal value for discriminating interseismic velocity versus coseismic velocity is given by the inflection of the curve (Figure A1a). We used 0.015 cm/s as the velocity threshold. Note that the same value is used for all seven accompanying experiments. To set the picking algorithm sensitivity, we calculated the number of picked events for sensitivities in the 1–30 frames range (Figure A1b) and found that the number of identified events is stable for sensitivities larger than 16 frames. Slightly faster or slower velocity thresholds (0.01 and 0.02 cm/s) provide the same number of picked events. We conclude that a sensitivity larger than 16 (20 in this case) provides a stable sampling characterized only by large events. Due to this decision, only peaks separated by periods of slow velocities longer than 2.6 s are considered. This approach, therefore, possibly includes aftershocks and foreshocks into one unique event.

Figure 5. Horizontal velocity component time series measured along a linear section located 1 cm above the subduction thrust fault. Maximum velocity above the seismogenic zone is highlighted in red. Grey lines are relative to the entire section velocities, and provide insight in the measurement noise. Velocity peaks (full green points) faster than a fixed velocity threshold (black line) are interpreted as analog earthquakes (see Appendix A1 for details on setting the velocity threshold).

Figure 6. (a) Spatio-temporal slip rate distribution and source parameter derivation. Contour of coseismic velocity threshold (black line) and slip rate (grey shading with red 0.05 cm/s isolines) in the space/time plane are used for identifying the hypocenter position and spatial rupture limits highlighted by the red point and empty red squares, respectively. (b) Spatio-temporal cumulative slip evolution. Red lines highlight 0.05 cm step contour. Pixel size of both figures is proportional to the spatial and temporal resolution of the image analysis technique.
Space-time plots of slip rate were produced for each event identified by the picking algorithm (e.g., Figure 6a). The onset of the stick-slip instability and the position of the rupture front as a function of time throughout the rupture process can be imaged with high discretization by identifying the position of the coseismic velocity threshold in these plots. This procedure allows for the tracking of model velocities during all stages of the experimental rupture (i.e., initiation, propagation, and arrest) and represents an important advance compared to earlier models. For each event, space-time plots of cumulative slip evolution were produced by summing particle displacements during the coseismic stage (Figure 6b). Note that, here, “slip” is the one-sided coseismic displacement because the displacement of the subducting plate is not measured.

3. Results

This section summarizes the observations and analyses of nine models conducted under the same experimental conditions (i.e., the same materials, geometric/kinematic conditions, experimental duration of 800 s). For simplicity, we refer to model #1 as the reference model. Seven supplementary models are used to constrain the variability of the experimental results and test the repeatability of our reference model. The resulting catalog of analog interplate earthquakes includes 260 events. The model surface deformation is acquired from an additional model (model #9) performed using the top-view setup.

3.1. Cyclicity of the Deformation Pattern

Here we discuss time series of the horizontal velocity component 1 cm above the subduction thrust (Figure 7a) to describe and verify the response to frictional behavior of the models.

After an initial 40 s period, in which the gelatin wedge adjusts to the experimental apparatus and the stresses are build up due to convergence (note that this phase has no natural equivalent and is neglected in the following analysis), the model exhibited two spatial domains, one in which cyclicity is detectable, one in which deformation is smaller than detection threshold.

Figure 7b shows the distribution of the recurrence intervals for the reference model and the entire dataset (blue bars and black line, respectively). Both distributions have the same recurrence interval mode of 19 s, verifying statistical significance of the observed model periodicity.

3.2. Wedge Deformation During the Analog Subduction Seismic Cycle: Coseismic and Interseismic Stages

The subduction earthquake cycle alternates between stages of slow strain accumulation and sudden fast motion. Here we use top- and side-view velocity fields to demonstrate that the analog model shares similarities to the natural interplate earthquake cycle as observed using, e.g., geodetic...
observations. For the sake of clarity, we focus exclusively on the analog interseismic and coseismic stages. We skip the post-seismic stage for spatial resolution limitations.

Figure 8a shows particle motion of the model surface during the interseismic stage. Marker velocities are generally directed rearward. Specifically, these velocities taper toward zero close to the model boundaries and reach their maximum in the order of tens of micrometers per second in the frontal and central area (i.e., 12 to 22 cm along the trench and 37 to 48 cm from the backstop). The area of maximum horizontal velocity is thus shifted a few centimeters closer to the trench with respect to the analog seismogenic zone. It is important to note that the velocity along the x-axis tends to be zero approaching the trench area. This relation is due to the presence of the shallow low basal friction (i.e., due to the presence of the plastic sheet) and reflects an increase of basal backslip dislocation (i.e., the amount of plate convergence during the interseismic interval for a kinematically locked megathrust) at the seismogenic zone.

Particle velocities at each time step were used to calculate the horizontal (i.e., trench-normal) displacement. Horizontal displacement was normalized by the amount of plate motion to provide an estimation of "plate locking" (Figure 8a). This mirrors the amount of horizontal plate convergence transferred to the upper plate during the interseismic stage. The percentage of plate locking is 0% to ~40% with a spatially heterogeneous distribution. The highest values of plate locking consistently occur above the seismogenic zone which is highly locked during the interseismic period (Figures 8a and 8c).

The side-view velocity field during the interseismic stage confirms the rearward motion of the model (Figure 8c). Rearward motion is particularly evident above the analog seismogenic zone. The maximum interseismic velocity of ~0.004 cm/s is located 45 cm from the backstop. A slightly downward direction of the velocity vectors occurs at 45 cm from the backstop. This trend changes to a slightly upward motion at 25–35 cm from the backstop.

The side-view velocity fields were used for calculating the vertical surface displacement (i.e., resulting model topography) and the amount of slip and backslip during both the interseismic and coseismic stages. In particular, topography is calculated summing incremental vertical displacement at the model surface, while backslip is obtained summing incremental horizontal displacement at 1 cm above the analog subduction thrust.

Accordingly, the interseismic stage is characterized by subsidence of the area updip of the seismogenic zone near the trench (Figure 8b). Rearward subsidence gives way to uplift in a 25 cm wide zone above and landward of the seismogenic zone. The inflection point from subsidence to uplift is located 15 cm from the trench. The 20 cm close to the backstop area remains basically flat. The amplitude of the maximum observed uplift and subsidence is 0.07 cm. The basal backslip varies from 0 cm to 0.06 cm (Figure 8d). The maximum backslip is located 45 cm from the backstop,
near the updip limit of the seismogenic zone. The backslip distribution tapers to zero at the trench and within 30 cm of the backstop. This condition reflects the presence of the nonsubducting, low-friction plastic sheet in these areas.

During the coseismic stage (Figures 8e–8h), the markers located at the surface move toward the trench (i.e., seaward) at least one order of magnitude faster than during the interseismic stage (Figures 8e and 8g). The fastest velocities, located along the model centerline at 40 to 50 cm from the backstop, slow to approximately zero both at the trench and near the backstop. We observe a maximum horizontal (i.e., trench-orthogonal) displacement of 0.09 cm (Figure 8e). Consistently, the highest coseismic surface displacements correspond to the areas of highest interseismic plate locking.

Surface topography, side-view velocities, and slip show the opposite pattern of that found during the interseismic periods. In particular, the model topography is characterized by an uplifted area 15 cm updip of the seismogenic zone followed by subsidence 23 to 45 cm from the backstop (Figure 8f). The inflection point from subsidence to uplift coincides with the location of highest measured slip. The slip distribution along the plate interface slightly asymmetrically tapers to zero at the rupture edges (Figure 8h). In the imaged analog event, the maximum slip occurs at the updip limit of the seismogenic zone.

### 3.3. Hypocenter Location, Rupture Width, and Slip Variation

The hypocenter location, one of the key source parameters, is important as a reference point to represent the spatial distribution of the slip [e.g., Chu et al., 2011]. The majority (~90%) of hypocenters are located inside the velocity-weakening area (Figure 9a). However, in the few remaining cases, the hypocenter is less than 2 cm from the upper limit of the seismogenic zone. In 86% of the cases, the hypocenters are located in the deeper half of the rupture (the downdip hypocenter position with respect to the rupture width is approximately 0.3 to 0.4; Figure 9a), and 95% of them are also deeper than the peak slip-rate location (see Figure 4d of the companion paper and Figure 6a). Consistently, the rupture preferentially propagates updip. In general, 80% of the rupture extends updip with respect to the hypocenter position. This observation is statistically consistent throughout the entire data set.

The hypocenters appear to be randomly located throughout the seismogenic zone and are mostly unrelated to the hypocenter location of the previous event. However, series of 3 to 4 events are “repeaters” characterized by hypocenters located in the same position (e.g., events 31–34 and 41–43; Figure 9a). On other occasions, sequences of 3 to 5 “triggered” events with progressively shallower hypocenters occur (e.g., events 6–8 and 19–23; Figure 9a).

Rupture area controls earthquake size and scales with earthquake magnitude [e.g., Wells and Coppersmith, 1994; Mai and Beroza, 2000; Blaser et al., 2010; Strasser et al., 2010]. Moreover, the position of the downdip and updip limits of an earthquake is important for the proximity of seismic sources to inland cities and the tsunamigenic potential of subduction earthquakes.

Here we focus exclusively on the along-dip extent of analog ruptures (i.e., the rupture width) due to experimental limitations (i.e., the analog model is quasi-2D). Analog interplate events in the reference model are characterized by rupture widths of 0.9 cm to 23.8 cm (Figure 9a). Equivalent natural rupture widths range between 5 km and 150 km. Figure 9a shows a simplified synoptic view of the downdip extent of the reference model sequence. The updip limit of all events is located within the seismogenic zone. Few events (9%) are characterized by updip rupture limits that reach, but do not cross, the base of the seismogenic zone. For the majority of the events (95%), the rupture extends to well within the shallower velocity-strengthening area. The rupture, however, never breaks the surface. Only two events are characterized by updip and updip limits confined to the seismogenic zone.

The average slip of an earthquake is another parameter controlling the magnitude of a seismic event. Analog earthquakes are characterized by an average “one-sided”

**Figure 9.** (a) Line-source slip distributions (grey shading), rupture extent (limited by empty red squares), and hypocenters (full red dots) for the reference model events. The black dashed lines mark the limits of the velocity-weakening seismogenic zone. (b) Mean slip versus rupture width for the reference model events showing a clear correlation between source parameters ($R^2 = 0.49$). The linear regression line, as well as the 95% confidence bounds are highlighted by red and grey lines, respectively.
slip ranging between 0.005 and 0.05 cm with a mean of 0.03 ± 0.01 cm. The equivalent natural slip ranges from 32 m to 320 m. The absolute maximum slip (0.08 cm) is located in the shallower part of the rupture of event number 36 (Figure 9a). The slip distribution of analog earthquakes systematically exhibits a taper with a rather pronounced asymmetry (Figure 8h). In particular, most events are characterized by peak slip locations within the shallower half of the rupture (Figure 9a). Analog earthquakes are characterized by a positive linear relation between the mean slip and rupture width consistent with crack theory (Figure 9b) [e.g., Marone and Richardson, 2006].

3.4. Seismic Moment vs. Duration

[60] The most important physical measure for earthquake size is the seismic moment, $M_w$, which is calculated as

$$M_w = G \cdot D \cdot W \cdot L,$$

where $G$ is the rigidity of the rock, $D$ is the average slip, and $W$ and $L$ are rupture width and length, respectively [Hanks and Kanamori, 1979]. This formulation, which requires knowledge of both the downdip and trench-parallel rupture extent, is not directly applicable to our model, because we do not simulate the third dimension in the presented quasi-2D setup. To extrapolate our data to a third spatial dimension, we apply a procedure that consists of using existing rupture width-magnitude scaling relationships [Wells and Coppersmith, 1994; Mai and Beroza, 2000; Blaser et al., 2010; Strasser et al., 2010]. Empirical relationships of the form

$$M_w = a + b \log_{10}(X)$$

are available from regression analysis based on large numbers of events in nature ($a$ and $b$ are coefficients that vary with rupture dimension $X$). Our choice is supported by the following observations: (a) the rupture width is the most robust observable in our experiments (taking into account both the updip and downdip limits of the rupture, an error of 0.26 cm is estimated based on the PIV interrogation window size; see Figure 6a); (b) the experimental average slip appears to be too large (scaling from 32 m to ~320 m) and would, therefore, provide unrealistic magnitudes, larger than 10; (c) the rupture width-magnitude scaling relationships are robust for various kinds of faulting in nature [Wells and Coppersmith, 1994; Mai and Beroza, 2000] and with special focus on the subduction environment [Blaser et al., 2010; Strasser et al., 2010].

[61] The smallest rupture width of our reference model is 0.9 cm, which corresponds to 5 km in nature. According to $M_w-W$ scaling relationships [Wells and Coppersmith, 1994; Blaser et al., 2010; Strasser et al., 2010], this rupture corresponds to a $M_w = 5.6$ earthquake. The largest rupture width of the reference model is 23.8 cm, corresponding to 150 km in nature. This rupture width has a large moment magnitude variability corresponding to $M_w = 8.3$, $M_w = 8.8$, and $M_w = 9.2$, according to the empirical relationships of Strasser et al. [2010], Blaser et al. [2010], and Wells and Coppersmith [1994], respectively. In the following, we exclusively use the intermediate $M_w-W$ scaling of Blaser et al. [2010]. In particular, we refer to their preferred relation for “orthogonal reverse-slip” events with a rupture width in the 2–240 km range:

$$M_w = -(a/b) + (1/b) \log_{10} W,$$

where the $a$ and $b$ coefficients are equal to ~1.86 and 0.46, respectively. This results in experimental moment magnitudes from 5.6 to 8.8 (with an average of 8.2 ± 0.5).

[62] The seismic moment is then calculated following Hanks and Kanamori [1979]:

$$\log_{10}(M_w) = 1.5 M_w + 9.1.$$

For analog earthquakes in the reference model, we calculate seismic moments ranging from $4 \cdot 10^{17}$ to $1.9 \cdot 10^{22}$ Nm and from $3 \cdot 10^{16}$ to $2.3 \cdot 10^{23}$ Nm, considering the entire data set. [63] A clear empirical relationship between earthquake duration and seismic moment has been found in nature [e.g., Abercrombie, 1995] and is used to further validate our approach. In particular, duration scales as the cube root of seismic moment for regular earthquakes and linearly for slow earthquakes [e.g., Ide et al., 2007; Houston, 2001].

[64] We compared analog earthquake moment to duration with scaling laws derived for regular and slow earthquakes [Ide et al., 2007] to relate the experimental mode of slip propagation to natural end-member models (Figure 10). The durations of analog earthquakes (ranging from 0.07 s to 2.2 s for the reference model) were upscaled in accordance with the coseismic time-scaling factor $T_c^*$. Interestingly, analog earthquakes are located between regular and slow earthquakes in the moment-duration space. In particular, these analog earthquakes are located in the field of shallow subduction and tsunami events (Figure 10) [cf. Peng and Gomberg, 2010; Bilek, 2004], even if a slightly longer duration than regular earthquakes is observed (upscaled analog durations range from 1–29 min). The slope of the least-squares fit of experimental data is similar to the slope of regular earthquakes. This observation suggests a mechanical similarity between analog earthquakes and earthquakes in

![Figure 10](image-url)
nature, that is, earthquakes operate as dynamically running shear cracks following unique source scaling relations.

3.5. Experimental Variability and Repeatability

[65] Providing details on the variability and repeatability of the experimental results is important for demonstrating the robustness of the experimental approach and for speculating about possible natural variability.

[66] We identified the most important experimental parameters for this purpose: earthquake duration, recurrence interval, and rupture width. These parameters describe the temporal features of interseismic and coseismic stages as well as the size of analog earthquakes.

[67] We use the coefficient of variation, $C_v$, to quantify the variability of experimental results, and use the two-sample Kolmogorov-Smirnov test to investigate the repeatability using eight similar models. $C_v$ is a dimensionless value that describes the dispersion of a selected variable. $C_v$ is defined as the ratio of the standard deviation of a data distribution to its mean. $C_v$ is considered a reasonable measure if it contains only positive values, as is the case here. In general, high $C_v$ values are associated with a broad distribution of the selected parameters. We computed $C_v$ for each of the eight side-view models. We found that our experiments are characterized by $C_v$s for earthquake durations of 0.40–0.95, recurrence intervals of 0.22–1.00, and rupture widths of 0.38–0.81 (Figures 11a, 11b, and 11c, respectively). The mean $C_v$ is 0.70 for the rupture duration, 0.57 for the recurrence interval, and 0.61 for the rupture width.

[68] To test whether there would be a similar data dispersion for a larger data set, we iteratively recalculated $C_v$ for an increasingly large set of samples by merging $n$ models (with $n = 1, \ldots, 8$). The case $n = 1$ provides the same result obtained when only considering model #1 (the reference model). Following this procedure, $C_v$ always reaches a plateau at approximately 0.6–0.7, for both the earthquake duration and rupture width when four models are merged into a set, indicating that the data dispersion does not further increase when the events of more than four models are considered together (Figures 11a and 11c). The recurrence interval does not display similar behavior; rather, $C_v$ linearly increases with an increasing number of events (Figure 11b).

[69] The nonparametric Kolmogorov-Smirnov test can be used to determine whether two data sets are drawn from the same population. In other words, this test evaluates how well two data sets match, demonstrating the repeatability of experimental results. The Kolmogorov-Smirnov statistic searches for the greatest difference, $k$, between the empirical cumulative distribution of the two samples (e.g., the rupture duration data of two distinct models). This statistic is calculated under the null hypothesis that the samples are drawn from the same distribution [Davis, 1973]. The test rejects the null hypothesis when the significance level is equal to or exceeds the p-value, $p$ (i.e., the minimum level of significance for which the null hypothesis is rejected). The significance level was set to 0.05. For each of the selected parameters, we tested all the possible combinations of pairs of models (i.e., 28 tests for each parameter; Figures 11d, 11e, and 11f).

[70] The comparison of rupture durations shows that there is a relatively high consistency of performance between different pairs of models. The average greatest difference between distributions of a pair of data sets, $k$, is 0.32 ± 0.10, and 82% of tests failed to reject the null hypothesis. A similar level of matching was observed for the rupture width. In this case, the average $k$ is 0.34 ± 0.12, and 75% of tests failed to reject the null hypothesis. The recurrence interval indicates the weakest level of similarity between cumulative distributions; the average $k$ is 0.41 ± 0.18, and 57% of tests failed to reject the null hypothesis.

4. Discussion

[71] The new experimental setting presented in this paper allows for the investigation of the coseismic and interseismic
behavior of subduction thrusts. Additionally, we identify four major aspects of our experimental results that deserve to be analyzed in detail: (a) the recurrence behavior, (b) the coseismic and interseismic model responses, (c) the rupture behavior, and (d) the source parameters. To validate our model, experimental results are first compared to previous analog models specifically designed for investigating the seismic cycle at subduction thrusts [Rosenau et al., 2009; Rosenau and Oncken, 2009; Brune, 1996] and accompanying numerical simulations performed in the same setting [van Dinther et al., 2013, paper 2]. To validate the approach we then discuss our findings with respect to natural observations. Although our models oversimplify the natural prototype, the analysis of the experimental results provides interesting insights into the complex behavior of subduction thrusts.

4.1. Recurrence Behavior

[73] The identification of stick-slip events with regular recurrence intervals allows us to compare our physically self-consistent model behavior to simple deterministic models of stress accumulation and release that have been proposed for describing earthquake cyclicity. The simplest model is the “characteristic earthquake model” [Reid, 1910], which postulates that a specific fault ruptures at regular intervals, generating events of similar size and rupture area. Noncharacteristic models include time- and slip-predictable models [Shimazaki and Nakata, 1980]. The time-predictable model implies the presence of a constant threshold strength at which the fault ruptures. This model assumes that slip during a large earthquake reduces the shear stress on the fault more than slip during a small event. Therefore, the time to the next earthquake can be predicted from the size of its predecessor. Alternatively, in the slip-predictable model, the fault does not rupture at a fixed shear stress, but each earthquake reduces the shear stress on the fault to a fixed minimum level. This model cannot be used to predict when an earthquake will occur, but predicts the minimum magnitude of the next earthquake. Accordingly, the size of an earthquake is correlated with the length of the preceding recurrence interval.

[74] Approximately 40% of the experimental recurrence intervals are clustered within a 5 s time window around the average value (Figure 7b). This first-order observation, focused only on recurrence intervals, seems to promote a “quasi-periodic recurrence” model. Further investigation, that includes event size, is used to test the applicability of the characteristic, slip- or time-predictable model to the experimental data. We compared analog earthquake size with the preceding or subsequent interseismic period. For our data low linear correlation coefficients ($R^2 < 0.06$) occur when correlating parameters that determine event size (mean slip and rupture width) with the preceding and subsequent recurrence times. The majority of analog earthquakes are scattered and display no clear trend. We conclude that neither time-predictability nor slip-predictability can be applied to the experimental behavior; nor does a characteristic model fit the data. In the latter case both slip- and time-predictable models should fit equally well. Rather, our models tends to follow a cyclicity governed by a quasi-periodic recurrence interval.

[75] An additional test for the applicability of the characteristic earthquake model to analog earthquakes is performed by analyzing the coefficient of variation, $C_v$, of the recurrence time. Clustered, random or quasi-periodic events would be characterized by a $C_v > 1$, $C_v \sim 1$, or $C_v < 1$, respectively [Kuehn et al., 2008]. The observation of a $C_v$ in the 0.2–0.6 range for the entire data set seems to support quasi-periodic recurrence as most representative of the analog model behavior. A stronger tendency toward the characteristic earthquake is observed when perfectly elastic systems are subject to both fixed loading rates and elastic energy-release conditions, as in spring-block models [e.g., Baumberger et al., 1994], Couette systems [Galeano and Rubio, 1994], and ring shear tests [e.g., Rosenau et al., 2008].

[77] The factors (e.g., subduction velocity, plate interface locking) controlling the frequency of large events at convergent margins remain one of the key questions in modern seismology. A mult centennial return time of major earthquakes has been constrained by both geologic observations [e.g., Goldfinger et al., 2003] and statistical simulations [McCaffrey, 2008], whereas our analog models provide recurrence intervals that are roughly one order of magnitude larger (~750–10000 years). Such a discrepancy might be
explained by (a) the uncertainties in estimating the lithospheric viscosity and elasticity—mainly in nature—and the resulting imprecision in estimating the experimental interseismic time scale factor and slip, respectively; (b) the 2D-like character of the analog model implies that an event has to pass through the section that is illuminated by the UV lamp to be captured. This contributes to detect a smaller number of events than those occurring in the entire naturally three-dimensional, experimental interface.

4.2. Coseismic and Interseismic Model Response

[78] The analog model successfully simulates one of the most important features of the seismic cycle at subduction zones: the elastic rebound associated with interplate thrust earthquakes as recorded in geologic and geodetic data [e.g., Natavidiya et al., 2004; Wang, 2007]. Accordingly, we observed periods of slow rearward motion alternating with short phases characterized by fast “seaward” motion (Figure 8).

[79] During the coseismic stage, the highest deformation rates and largest displacements occur in the seaward-most part of the wedge (Figures 8e and 8g). However, small displacements (i.e., 0.01 cm) are observed far inland, up to 45 cm from the analog trench (~300 km in nature, which is about the location of the volcanic arc in subduction zones). This finding agrees with observations during megathrust earthquakes (e.g., the 2004 Mw 9.2 Sumatra event; Chlieh et al., 2007), experimental results obtained in granular elastoplastic models [Rosenau et al., 2009], and the companion numerical simulation [van Dinther et al., 2013, paper 2].

[80] Our slip distribution is asymmetrical, with the maximum slip near the updip end of the rupture (Figures 8h and 9). The sense of skewness has been proposed to correlate to the stress drop trend in a seismogenic zone [Wang and He, 2008]. A downdip increase in stress drop is responsible for the slip distribution to be skewed “landward” [e.g., Wang and He, 2008; Rosenau et al., 2009] and vice versa [Rudnicki and Wu, 1995; Brune, 1996; Geist and Dmowska, 1999]. In particular, the discrepancy between experimental approaches could be related to the capability to simulate seismic waves and their amplification (resulting in trenchward increasing slip) in a realistic way, as in the dynamic and fully elastic models (e.g., Brune [1996] and this work). In the model of Rosenau et al. [2009], seismic energy could be quantitatively absorbed by the granular nature of the wedge medium (Rosenau, pers. comm. 2012). However, the exact shape of the slip distribution depends also on mechanical details of the rupture zone [e.g., Scholz, 2002; Bilek, 2007].

[81] Natural cases of large subduction interplate events reveal variable slip distributions with maximum slip recorded both in the shallower part (e.g., the 2011 Mw = 9.0 Tohoku-Oki earthquake; Lee et al., 2011) and in the central part of the subduction thrust (e.g., the 2004 Mw = 9.2 Sumatra earthquake; Chlieh et al. [2007]).

[82] The observed maximum coseismic subsidence does not coincide with where the slip tapers to zero, as predicted by elastic dislocation modeling [e.g., Wang, 2007] and observed in the companion numerical simulation [van Dinther et al., 2013, paper 2], but is shifted a few centimeters toward the trench (Figures 8f and 8h). While the observed shift, captured also in the elastoplastic analog models by Rosenau et al. [2009], might have important implications for inversion of surface deformation in nature, its explanation remains obscure. In active convergent margins, the base of the seismogenic zone and, consequently, the location of the maximum subsidence, roughly coincides with the coastline [e.g., Plafker, 1972] and it can reach a maximum amplitude of about ~2 m as in the 1960 Mw = 9.5 Chile earthquake [Plafker and Savage, 1970].

[83] Geologic and geodetic observations helped in identifying the three primary processes that took place in the aftermath of an earthquake: (a) long-term (decadal scale) mantle viscoelastic relaxation [e.g., Pollitz et al., 2008]; (b) short term (days to years) relaxation of shear stresses along the plate interface, which lead to seismic and aseismic afterslip [e.g., Perfetti et al., 2005]; (c) relocking of the subduction thrust fault [e.g., Gagnon et al., 2005]. The resulting postseismic, time-dependent deformation is characterized by a slow seaward motion of some inland areas that occurs simultaneously with outer-wedge landward motion as reported from Chile and Alaska where Mw = 9.5 and Mw = 9.2 earthquakes occurred in 1960 and 1964, respectively. In particular, this pattern reflects the interference of two processes in the seismic cycle: the onset of plate locking at shallow depth and viscoelastic relaxation of the mantle beneath the continent [e.g., Wang et al., 2012]. At present, such evidence cannot be imaged in detail in the analog viscoelastic model also because of the missing upper mantle contribution.

[84] The interseismic stage is characterized by rearward velocities (Figures 8a and 8c). A decrease in horizontal velocities from the locked region to inland areas is a common feature recognized in active convergent margins [e.g., Simons et al., 2011] and elastoplastic analog models [Rosenau et al., 2009], as well as numerical models like the companion numerical simulation.

[85] Consistently, landward motion is associated with uplift above the downdip limit of the seismogenic zone (Figure 8b). These experimental observations are similar, for example, to the vertical deformation pattern inferred through repeated leveling observations at the Nankai margin [Thatcher, 1984].

[86] One of the major difference between viscoelastic and elastoplastic models is in their basal backslip distributions. Our interseismic backslip (Figure 8d) tapers to zero at the updip end of the analog subduction thrust. This behavior mirrors an interseismic slip rate at the subduction thrust that progressively decreases from the plate convergence rate to almost zero in the locked zone [Wang, 2007]. The high interseismic slip rate at the shallowest part of the subduction thrust is caused by updip aseismic velocity-strengthening plastic sheet. Elastoplastic models [Rosenau et al., 2009] are instead characterized by a basal backslip smoothly increasing from the base of the seismogenic zone to the trench. This behavior suggests that the fault is locked all the way up to the trench [Wang, 2007] as a consequence of underthrusting / accretion of material. In the Rosenau et al. [2009] model, shallow slip occurs, therefore, along splay faults in the overriding plate instead of along the plate interface.

[87] The fraction of plate convergence not accommodated by aseismic slip in the interseismic period, referred to as the interseismic coupling, reveals consistent similarities between viscoelastic and elastoplastic models of the subduction interplate earthquake cycle. Figure 8a shows that the plate convergence accumulated in the model varies from ~40% above the seismogenic zone to zero at the inland area.
The variability in interseismic coupling observed at subduction margins remains one of the least understood elements of the strain accumulation process [e.g., Scholz and Campos, 2012]. Based on geodetic data, some subduction zones are considered fully locked, whereas others are only partially locked or are freely slipping [e.g., Dixon and Moore, 2007], with spatio-temporal variability possible within the same subduction zone [e.g., Freymueller et al., 2008; Heuret et al., 2011].

4.3. Rupture Behavior

[88] Monitoring the spatial and temporal behavior of earthquake rupture (i.e., initiation, propagation, and arrest) is necessary for understanding the physics of dynamic rupture propagation, which, in turn, may help us to extrapolate from theory to natural settings, and to thereby better characterize the resulting seismic hazard.

[89] We process the experimental data in a purely kinematic fashion, as is commonly done in seismic or geodetic source inversions [e.g., Ide et al., 2007]. A physics-based rupture propagation model within a quasi-static framework is extensively described in the companion paper (sections 4.4 and 4.6).

[90] The location of rupture initiation (i.e., hypocenter) is a first-order source parameter serving as a reference for the kinematic characterization of the spatio-temporal slip distribution during analogue earthquakes. In nature, the location of the hypocenter is generally more reliable with respect to other earthquake characteristics, such as the location of the largest slip [Ide et al., 2007]. Our experimental results show locations of hypocenters are dominantly located near the base of the analog earthquake rupture (Figures 6a and 9a). This finding is in agreement with results obtained in our numerical simulation (Figure 4c in the companion paper), in elastoplastic analog models of the subduction seismic cycle [Rosenau et al., 2009], in numerical models of crack propagation [Das and Scholz, 1983], and by analyzing selected dip-slip and strike-slip natural events [Mai et al., 2005]. On the basis of 12 finite-source rupture models for subduction events, Mai et al. [2005] suggested that these earthquakes prefer to nucleate in the along-dip center of the fault. Observations of recent megathrust earthquakes reveal that both cases (i.e., deep and intermediate depth nucleation) may occur. The 2004 $M_w=9.2$ Sumatra earthquake was characterized by a hypocenter located in the deepest part of the rupture [Rhee et al., 2007], whereas in the 2010 $M_w=8.8$ Maule and the 2011 $M_w=9.0$ Tohoku-Oki earthquakes, the hypocenters were located in the along-dip center of the rupture [e.g., Moreno et al., 2010; Lee et al., 2011]. Few analog hypocenters (i.e., $<$10%) are located in the shallow velocity-strengthening zone likely originated by the sandpaper/plastic sheet frictional discontinuity.

[91] Once the rupture initiates, it mainly propagates bilaterally with preference in the updip direction. This behavior is consistent with the observation of granular elastoplastic [Rosenau et al., 2009], foam-rubber analog models of inter-plate seismicity [Brune, 1996], and the numerical results obtained in the companion paper (Figure 4c). One possible explanation is that the rupture follows the lithostatic pressure gradient that results from the subduction wedge [Das and Scholz, 1983]. Another possibility is that rupture asymmetry results from a strong bimaterial contrast in elastic properties for gelatin and steel on each side of the analog thrust interface. As far as the bimaterial contrast is concerned, the analog setup simulates an upper bound of the natural convergent setting, since the overriding plate is expected to be the more compliant than the subducting one [e.g., Ma and Beroza, 2008]. Observations [e.g., Rubin and Gillard, 2000], numerical models [e.g., Dalguer and Day, 2009], and analog models [Xia et al., 2004] show that if a rupture occurs on the boundary between two solids with different elastic properties, the rupture may propagate in the direction of the slip of the softer rocks (e.g., gelatin in our model).

[92] A possible consequence of asymmetry in analog earthquakes is revealed by hypocenters that are generally deeper than the regions of peak slip (Figure 9a) and peak slip rate (Figure 4d in the companion paper). A similar rupture pattern has been inferred for the 2010 $M_w=8.8$ Maule earthquake and the 2011 $M_w=9.0$ Tohoku-Oki earthquake [Lee et al., 2011; Vigny et al., 2011]. In particular, for the 2010 $M_w=8.8$ Maule earthquake such relationship might result from stress-concentration induced by a downdip gradient in locking [Moreno et al., 2010]. Based on the observation of earthquakes with various kinematics, Mai et al. [2005] suggested that hypocenters are not randomly located on a fault but are close to regions of large slip.

[93] The analog earthquake rupture front propagates with an average velocity of 13.9 $\pm$ 7.1 cm/s upward and 11.1 $\pm$ 8.1 cm/s downward. The equivalent natural rupture velocities are 0.11 $\pm$ 0.05 km/s and 0.09 $\pm$ 0.06 km/s, respectively. Even if experimental rupture speed is slower than velocities inferred for natural interplate events [e.g., Lee et al., 2011], analog earthquakes are not similar to slow earthquakes (as shown in Figure 10 and described in section 4.4).

[94] The coseismic slip rate of analog earthquakes varies from 0.015 cm/s to 0.320 cm/s. The equivalent natural slip rate ranges from 0.1 m/s to 2.6 m/s, which is comparable to the general range of natural slip rates [e.g., Scholz, 2002] and to the slip rate of the 2011 $M_w=9.0$ Tohoku-Oki earthquake [Lee et al., 2011].

[95] Another important characteristic of earthquakes is the amount of slip, which may reach tens of meters for very large subduction interplate events. For example, the great 1960 $M_w=9.5$ Chile earthquake was characterized by a maximum slip of $\sim$40–50 m [e.g., Barrientos and Ward, 1990; Moreno et al., 2009], and the 2011 $M_w=9.0$ Tohoku-Oki earthquake had a maximum slip of $\sim$50–60 m [Simons et al., 2011; Lee et al., 2011]. The peak and average slip of analog earthquakes scale to roughly one-order-of-magnitude larger values (i.e., tens to a few hundreds of meters). Possible explanations for this discrepancy include the quasi-2D setting [e.g., Bilek and Lay, 2002] and the absence of off-fault plasticity for the analog model [e.g., Dunham et al., 2011; Wang and Bilek, 2011]. Propagation in the lateral third dimension and off-fault plasticity would absorb part of the energy that is used for slip on the analog subduction thrust. Additionally, the strong bimaterial contrast existing in the analog model favors a larger strength and consequent stress drop and thereby contributes positively in providing large slip [Ma and Beroza, 2008].

[96] Analog earthquakes can be further characterized by their rupture mode. Rupture growth can be described according to two competing rupture styles: crack-like versus pulse-like. In the former case, the nucleation region continues to slip...
throughout the earthquake, and the slipping region expands until arresting phases arrive from the borders of the fault. In the latter case, a healing front propagates behind the rupture front [e.g., Marone and Richardson, 2006]. Quantitatively, these modes can be distinguished by analyzing the duration of slip at a certain point on a fault with respect to the duration of the rupture of the entire fault [Rosakis et al., 2007]. In particular, the pulse mode is characterized by slip at a point on the fault that is one order of magnitude shorter than the total event duration [e.g., Heaton, 1982; Hartzell and Heaton, 1983; Heaton, 1990]. Distinguishing between these two rupture modes is a central issue in earthquake source physics because they predict different hazard levels [e.g., Marone and Richardson, 2006].

[97] Figure 6a shows how we characterized the rupture behavior of analog earthquakes. The minimum duration of slip at a point on the fault is larger than 1/10 of the entire rupture process, indicating a crack-like behavior. The spatio-temporal cumulative slip distribution (Figure 6b) also seems to support the idea of growth of the rupture as a crack. However, the patch with a high slip rate is concentrated in a small region that propagates upward just behind the rupture front, suggesting possible rupture growth in a mixed pulse/crack fashion. Similar behavior has also been observed in previous elastoplastic models of subduction interplate seismicity [Rosenau et al., 2009] and in incoherent frictional interface experiments [e.g., Rosakis et al., 2007]. Although a complete characterization of rupture styles is beyond the scope of this paper, we consider this the next challenge to be investigated using this experimental setting.

[98] Detailed observations of rupture growth in large subduction interplate earthquakes are available only for recent mega-events, which have been captured by modern, high-resolution geodetic and seismological monitoring methods. For example, along-strike crack-like rupture propagation has been proposed for the 2001 $M_s = 8.4$ Peru earthquake [Robinson et al., 2006], the 2004 $M_w = 9.2$ Sumatra earthquake [Kruger and Ohrnberger, 2005], and during the initial 75 s of the 2011 $M_w = 9.0$ Tohoku-Oki earthquake [Lee et al., 2011]. Slip pulse behavior has been suggested for the 2010 $M_w = 8.8$ Maule earthquake [Heaton et al., 2011].

4.4. Source-Parameter Scaling

[99] Faulting parameters are systematically related [e.g., Kanamori and Anderson, 1975], and a wide range of scaling laws has been identified depending on seismic moment range [e.g., Shimazaki, 1986], intraplate/interplate events [e.g., Kanamori and Anderson, 1975; Blaser et al., 2010; Strasser et al., 2010], fault kinematics [e.g., Wells and Coppersmith, 1994; Mai and Beroza, 2000], and shape of the rupture [e.g., Scholz, 2002].

[100] We examined the average slip/rupture width relationship, which linearly correlates with a proportionality factor of $1.7 \cdot 10^{-3}$ (Figure 9b). This observation might support the $W$-model [e.g., Scholz, 1982]. However, the absence of the third dimension (i.e., $L$) dimension in the experimental setting does not allow for a conclusive argument for such speculation.

[101] Subsequently, we investigate the seismic moment/duration relationship to test the capability of the model to reproduce similarities in slip behavior with respect to natural end-members (i.e., regular and slow earthquakes). Similar to natural events, we identified a clear dependence of analog earthquake duration on seismic moment. In particular, analog earthquakes are located in a region between regular and slow earthquakes, since upscaled analog earthquakes duration is roughly one order of magnitude larger than megathrust earthquakes in nature (i.e., average duration of 1.15 s correspond to ~15 min in nature). Slow earthquakes show a much larger characteristic duration in the range of hours to months [e.g., Ide et al., 2007].

[102] The experimental data follow a moment/duration scaling law characterized by the trend of regular earthquakes (Figure 10). This suggests that even if the model rupture duration is slightly longer than for regular natural earthquakes, the size/duration proportionality is retained. This observation allows us to speculate that the physics controlling natural and analog earthquakes is similar.

5. Conclusions

[103] A new analog modeling technique has been developed to shed light on the seismic cycle at subduction megathrusts. The quantitative analysis of velocity time series of the gelatin-on-sandpaper analog system allowed us to monitor the model deformation and measure earthquake source parameters.

[104] Our experimental results show that:

[105] (a) The recurrence interval of analog earthquakes can be described as “quasi-periodic” recurrence interval.

[106] (b) The interseismic model deformation shows rearward motion associated with subsidence in the outer wedge and uplift of the coastal area. The coseismic stage is characterized by faster velocities in the seaward direction. During this stage, the model shows a subsidence of the coastal area and uplift in the outer wedge.

[107] (c) Analog earthquakes generally nucleate in the deeper portion of the rupture and propagate preferentially seaward. The majority of events propagate with a crack-like behavior.

[108] (d) There are clear proportionalities between rupture width and slip, and between seismic moment and rupture duration.

[109] (e) The size, duration, and recurrence interval of analog earthquakes are consistent among models performed under the same experimental conditions. These findings support the repeatability of the experimental results.

[110] This work verifies that our analog model simulates the basic physics governing the seismic cycle and relating rupture processes in a simplified but robust way. Qualitative and quantitative comparisons between modeling results and what is known about the subduction earthquake cycle in nature suggest that our experimental approach offers an efficient and versatile tool for investigating the seismic cycle at subduction thrusts. Future efforts will be directed toward increasing the level of complexity of the experimental setting to systematically explore the parameter space influencing the subduction interplate seismic cycle.

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Appendix A: Setting the Coseismic Velocity Threshold and Picking Algorithm Sensitivity

Figure A1: Setting the coseismic velocity threshold and picking algorithm sensitivity. (a) Trade-off curve (grey line) showing the distribution of the number of frames as a function of the maximum horizontal velocity observed at each frame of the reference model. The red point indicates the optimal value for the coseismic velocity threshold, which is 0.015 cm/s. (b) Number of picked events as a function of picking algorithm sensitivity (red line). The number of picked events remains stable for sensitivities larger than 16. A similar result is obtained also for velocity thresholds of 0.01 cm/s and 0.02 cm/s (grey lines).

References


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