Fault interaction and stress triggering
<table>
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<th>Interaction type</th>
<th>Perturbation effects</th>
<th>Spatial scale</th>
<th>Temporal scale</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dynamic</td>
<td>- Rupture propagation; - Arrest</td>
<td>1 – 60 Km</td>
<td>1 – 20 s</td>
</tr>
<tr>
<td>Static</td>
<td>- Earthquake triggering; - Off – faults aftershocks; - Sesimicity rate change</td>
<td>1 – 60 Km</td>
<td>minutes – few years</td>
</tr>
<tr>
<td>Post – seismic</td>
<td>Long – term stress changes</td>
<td>10 – 1000 Km</td>
<td>few years – centuries</td>
</tr>
</tbody>
</table>
Following the Coulomb’ s failure assumption we define a Coulomb Failure Stress as ( e. g. Jaeger and Cook, 1969):

\[ CFS = \|T\| + \mu (\sigma_n + p_{fluid}) - C \]

where:

- $\|T\|$ is the shear traction modulus,
- $\mu$ is the coefficient of friction,
- $\sigma_n$ is the normal stress (positive in tension),
- $p_{fluid}$ is the pore fluid pressure,
- $C$ is the cohesion.

Assuming $\mu$ and $C$ constant over time, we have the Coulomb Failure Stress change:

\[ \Delta CFS = \Delta \|T\| + \mu (\Delta \sigma_n + \Delta p_{fluid}) \]

where it has been assumed an isotropic failure plane.
\( \Delta \text{CFS} \) is used to evaluate if one earthquake brought another earthquake closer to, or farther from, failure:

\[ \Delta \text{CFS} > 0 \Rightarrow \text{fault plane loaded} \Rightarrow \text{closer to failure} \]

\[ \Delta \text{CFS} < 0 \Rightarrow \text{fault plane relaxed} \Rightarrow \text{farther from failure} \]

(Stress Shadow)

Neglecting the spatial dependence in tractions, are:

\[
\begin{align*}
T(t) &= T(0) + \Delta T(t) \\
\sigma_n(t) &= \sigma_n(0) + \Delta \sigma_n(t) \\
p_{\text{fluid}}(t) &= p_{\text{fluid}}(0) + \Delta p_{\text{fluid}}(t)
\end{align*}
\]

Therefore we can write:

\[
\Delta \text{CFS}(t) = \| T(0) + \Delta T(t) \| - \| T(0) \| + \mu (\Delta \sigma_n(t) + \Delta p_{\text{fluid}}(t))
\]

\( \Delta \|T\| \) is the change in shear stress due to the first earthquake and it is resolved in the slip direction of the second earthquake;

\( \Delta \sigma_n \) is the change in normal stress due to the first earthquake and it is resolved in the direction orthogonal to the fault plane of the second earthquake.
## Stress changes approaches (after Harris, 1998)

<table>
<thead>
<tr>
<th>Method</th>
<th>Parameters Required</th>
<th>Successes</th>
<th>Problems</th>
<th>Author(s)</th>
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<tbody>
<tr>
<td>Static Coulomb failure stress</td>
<td>AUCS(1), Δu, Δτ, Δτ</td>
<td>may predict rupture lengths, given fault geometry</td>
<td>does not explain long delays (more than tens of seconds) between subevents; needs more testing</td>
<td>Harris et al. [1994], Harris and Doe [1995], Hill et al. [1995], Gomberg and Bock [1994], Spudich et al. [1990], Cotton and Crotwell [1997].</td>
</tr>
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<td>(elastic)</td>
<td></td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Static rate and state</td>
<td>Δu, Δτ, Δτ, A, B, H, t, τ</td>
<td>may predict aftershock duration</td>
<td>needs more testing; rate-awaited parameters defined in the laboratory but not known for the Earth</td>
<td>Durieuxch [1994], Durieuxch and Rivery [1994], Rivery and Monnet [1996], Genis and Ballesteros [1998], Gomberg et al. [this issue], Harris and Simpson [this issue], and Inda et al. [this issue]</td>
</tr>
<tr>
<td>Dynamic rate and state</td>
<td>Δu, Δτ, Δτ, A, B, H, t, τ</td>
<td>may explain remote triggering</td>
<td>needs more testing; still need to define rate-awaits parameters in the Earth, initial terms not yet included in models</td>
<td>Durieuxch [1997] and Gomberg et al. [1997], this issue</td>
</tr>
<tr>
<td>Static Coulomb failure stress</td>
<td>Δu, Δτ, Δτ, A, B, H, t, τ</td>
<td>may explain time delays between mainshocks and subsequent events, also</td>
<td>needs more testing, also needs more geodetic data to confirm viscoselastic parameters</td>
<td>Dingwea et al. [1988], Rishi [1988], Goff et al. [1993], Tavakoli et al. [1996], Public and Sacks [1997], Prof and Lin [this issue]</td>
</tr>
<tr>
<td>(viscoelastic)</td>
<td></td>
<td>irregular recurrence intervals</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Fluid flow</td>
<td>Δu, Δτ, Δτ, A, B, H, t, τ</td>
<td>may explain time delays between mainshocks and subsequent events</td>
<td>may not be successful at predicting both the spatial and temporal aftershock pattern</td>
<td>Li et al. [1987], Hauk et al. [1987], Noc et al. [1997], etc., Seiber et al. [1997], this issue</td>
</tr>
</tbody>
</table>

*If the aftershock fault planes are not known, then some authors assume optimally oriented faults; this requires knowledge of the background stress directions.
1 – D Spring – slider model
m \ddot{\delta} = k (\delta_0 - \delta) - \tau_f + \Delta \tau, \quad \Delta \tau(t) \text{ perturbazione}

\tau_f = \text{resistenza di attrito}

Reologia: attrito rate- and state-dependent

\theta (\Phi) = \text{variable di stato della superficie}, \quad V = \dot{\delta} \text{ velocità}

\begin{align*}
A - \text{Ruina-Dieterich} \\
\tau_f &= \tau_\theta + \theta + A \ln \left( \frac{V}{V_\theta} \right) \\
\frac{d \theta}{dt} &= -\frac{V}{L} \theta + B \ln \left( \frac{V}{V_*} \right)
\end{align*}

\begin{align*}
B - \text{Dieterich - Ruina} \\
\tau_f &= \tau_\theta - A \ln \left( \frac{V}{V_\theta} \right) + B \ln \left( \frac{\Phi V_*}{V_*} \right) \\
\frac{d \Phi}{dt} &= 1 - \frac{\Phi V}{L}
\end{align*}

Stato del sistema: \( (V(t), d(t), t_f(t)) \)

o condizioni mecc. faglia

approx. g. statica

\( V < V_c = 0.1 \text{ mm/s} \)

Inertia is negligible and the system passes through a sequence of equilibrium states
Fault seismic cycle modeling

\[(\tau_f - \tau_o)/A\]

Steady State Line

Time Intervals (%T )

- 0-10
- 11-20
- 21-30
- 31-40
- 41-50
- 51-60
- 61-70
- 71-80
- 81-90
- 91-100

\[\ln(V/V_0)\]
Analytical stress perturbations

![Graph showing slip velocity vs time with various labels and markers indicating dynamic motion, quasi-static motion, seismic range, aseismic range, and clock advance.]
Analytical stress perturbations
The step and the pulse #1

Step

Pulse

\[ \theta (\text{bars}) \]

\[ V (\mu \text{m/s}) \]

\[ \tau + \Delta \tau (\text{bars}) \]

Time (s)
Analytical stress perturbations
The step and the pulse #2
Realistic stress perturbations

Syntetic seismograms #1

stress time history from
Realistic stress perturbations
Syntetic seismograms #2

![Graph showing stress and slip velocity over time]
Fault interaction by dynamic stress transfer: the case of the 2000 South Iceland seismic sequence

Part I
To evidence the eventual effect of the transient part of the coseismic stress changes due to the 17 June 2000, M 6.6 South Iceland earthquake;

The debate on the triggering potential of transient stress changes is still open;

The observational evidences are difficult and few.
The choice of the events

- The largest events (M ~ 5) occurring in the first five minutes
  - 8s, 26s, 30s, 130s, 226s
- In intermediate - far field
  - 26s, 30s, 130s, 226s
- That reasonably are not secondary aftershocks
  - 26s, 30s, 226s.
The 26 s and 30 s events

• They were not detected teleseismically.

• **26 s (64 km far)**
  – Not detected by DInSAR.
  – Known fault.

• **30 s (77 km far)**
  – Waveforms partially obscured by the first event (mechanism uncertain)
  – Detected by DInSAR and surface effects.
  – August 2003: M 5 event on N-S fault with the same epicenter.

From SIL seismograms the 26 s and 30 s events occurred at the arrival (later than the first) of shear waves traveling at 2.5 km/s at their location.

<table>
<thead>
<tr>
<th>Event</th>
<th>Origin time</th>
<th>Latitude (°)</th>
<th>Longitude (°)</th>
<th>Depth (km)</th>
<th>ML</th>
<th>MLw</th>
</tr>
</thead>
<tbody>
<tr>
<td>26s</td>
<td>154106.9</td>
<td>63.951±0.004</td>
<td>-21.689±0.008</td>
<td>8.9±1.3</td>
<td>4.91</td>
<td>6</td>
</tr>
<tr>
<td>30s</td>
<td>154111.254</td>
<td>63.937±0.003</td>
<td>-21.94±0.01</td>
<td>3.8±1.3</td>
<td>4.68</td>
<td>5.9</td>
</tr>
</tbody>
</table>
Parameters used to compute the dynamic stress

- Slip distribution from geodetic data (Arnadottir et al. 2003). Right lateral strike slip fault, strike 7° E, dip 86°.
- Rupture history: bilateral Haskell model, rise time: 1-2 s, rupture velocity: 2.5 km/s.
- 2 crustal models with 4 layers:

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>( V_p ) (km/s)</th>
<th>( V_s ) (km/s)</th>
<th>Density (kg/m³)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-3.1</td>
<td>3.3</td>
<td>1.85</td>
<td>2300</td>
</tr>
<tr>
<td>3.1-7.8</td>
<td>6.0</td>
<td>3.37</td>
<td>2900</td>
</tr>
<tr>
<td>7.8-17</td>
<td>6.85</td>
<td>3.88</td>
<td>3100</td>
</tr>
<tr>
<td>&gt;17</td>
<td>7.5</td>
<td>4.21</td>
<td>3300</td>
</tr>
</tbody>
</table>

West of Hengill

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>( V_p ) (km/s)</th>
<th>( V_s ) (km/s)</th>
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</tr>
<tr>
<td>&gt;17</td>
<td>7.5</td>
<td>4.21</td>
<td>3300</td>
</tr>
</tbody>
</table>

East of Hengill
Dynamic stresses at the two hypocenters

- Nord - Sud vertical right - lateral faults
- $\Delta \text{CFF} = \Delta \tau + \mu (1 - B) \Delta \sigma_n$, with $\mu = 0.75$, $B = 0.47$
- Rise time: 1.6 s

26 s aftershock

30 s aftershock
\[ \Delta CFF(t) \text{ at the two hypocenters} \]

Time separation between the events and between stress peaks comparable.
Snapshots of dynamic stress

- Snapshots at different times (23, 27, 31, and 35 seconds)

- Color scale showing stress levels ranging from -0.5 to 0.5 MPa

- Depth indication: 8.9 km
• Stress at each hypocenter is affected by uncertain parameters such as the crustal model, rise time and the hypocentral depth.

• **Crustal model**

<table>
<thead>
<tr>
<th>Parameters sensitivity #1</th>
</tr>
</thead>
<tbody>
<tr>
<td>26 s aftershock</td>
</tr>
<tr>
<td>30 s aftershock</td>
</tr>
</tbody>
</table>

- The origin times (from mainshock) of the two events remain at, or follow closely the second CFF peak for ~ 1 - 2 s rise time.
Parameters sensitivity #2

Rise time

Hypocentral depth

Uncertainties in stress amplitudes.
The fault response

- We study the fault response to the stress changes as evaluated at the two hypocenters with varying the parameters within their uncertainties;

- We use a spring-slider model with rate- and state-dependent friction for variable effective normal stress $\sigma_{neff}$;

- The system is perturbed either in shear stress and normal stress ($\Delta\tau(t)$, $\Delta\sigma_{neff}(t)$);

- We investigate the possibility of instantaneous triggering (during the transient stress perturbation).
The instantaneous trigger

- $h \sim 10$ km linear fault dimension,
- standard values of rheological parameters ( $\mu^* = 0.7$, $L = 1$ mm, $b = 0.01$),
- $v_0 = 2$ cm/yr (spreading rate in the SISZ),
- fault in close to failure conditions (100% steady state $\mu$ unperturbed failure expected at less than 2 yr from June 17, 2000)

The fault tends to fail within 1 s after a peak in CFF, as evaluated at the two hypocenters

\textit{if}

1. the initial effective normal stress $\sigma_0$ is enough low, so that the shear stress perturbation $\Delta \tau$ at that peak is much larger than $a(\sigma_0 + \Delta \sigma)$

2. and the direct effect of friction $a$ is enough low to keep fault unstable ($k/k_{\text{crit}} < 1$) for low values of $\sigma_0$. 
For $a \leq 0.003$ and $\sigma_0 \approx 20$ bar, we obtained instantaneous trigger within 1 second after the second peak of CFF, as expected for the two aftershocks in the SISZ.

For $a = 0.003$ and $\sigma_0 > \gamma$ 20 bar, $1 < \gamma < 10$ (increasing with the amplitude of the second peak of $\Delta \tau$) the trigger is not instantaneous (failure time > 4 hours).

Mean failure time $\approx 26$ s
The 26 and 30 s events occurred near one of the important geothermal areas of Iceland;

They were negligibly affected by static stress changes;

They followed closely a peak of positive CFF;

These results favour the hypothesis of dynamic triggering;

Dynamic models of fault responses can explain observations for low values of effective normal stress (near lithostatic pore pressure).
Fault interaction by dynamic stress transfer: the case of the 2000 South Iceland seismic sequence

Part II
The values of the tensor $\Delta \sigma_{ij}$ are calculated on the 26 s fault plane up to 2.78 Hz, in a total of 12 × 8 “receivers”, located in nodes uniformly spaced 1650 m in the strike direction and at depths of 0 m, 1650 m, 3300 m, 4950 m, 6550 m, 8100 m, 9900 m and 11550 m.

\[
\begin{align*}
\mathbf{T}^{(\mathbf{n})} &= n_j \sigma_{ij} \\
\mathbf{T}^{(\mathbf{n})} &= \mathbf{T}^{(\mathbf{n})} + \Sigma^{(\mathbf{n})} \\
T_j^{(\mathbf{n})} &= n_j \sigma_{ij} - n_j (n_i \sigma_{ik} n_k) \\
\Sigma_j^{(\mathbf{n})} &= n_j (n_i \sigma_{ik} n_k)
\end{align*}
\]

For our shear rupture:
\[
\mathbf{\hat{n}} \parallel \mathbf{x}_2 \equiv (0,1,0) \\
\mathbf{T}^{(\mathbf{n})} = (\sigma_{21}, 0, \sigma_{23})
\]
The spatial sampling of the receiver grid is not sufficient to correctly resolve the dynamic processes occurring during the rupture nucleation and propagation (Bizzarri and Cocco, 2003; 2005), as well as the temporal discretization.

We develop an algorithm that employs a $C^2$ cubic spline to interpolate $\Delta \sigma_{ij}$ in space and in time.

![Graphs showing original and interpolated values at $t = 26.37$ s.](image)
At time $t$, in each fault node, the dynamic load is:

$$L_i = f_{ri} + T_{0i} + \Delta \sigma_{2i}$$

($i = 1$ and 3).

$T_{0i}$ are the components of the initial traction ($T_0(x_1,x_3) = \tau_0(x_1,x_3)(\cos(\varphi_0),0,\sin(\varphi_0))$).

$f_{ri}$ are the components of the load (namely the contribution of the restoring forces, $f_r$) exerted by the neighboring points:

$$f_{ri} = (M^+ f^+_i - M^- f^-_i)/(M^+ + M^-),$$

where $M^+$ and $M^-$ are the masses of the “+” and “−” half split–node of the fault plane $\Sigma$ and $f^+$ is the force acting on partial node “+” caused by deformation of neighbouring elements located in the “−” side of $S$ (and viceversa for $f^−$).

$\{\Delta \sigma_{2i}\}$ are coupled to the components of the fault friction $T_i$ via

$$\frac{d^2}{dt^2} u_1 = \alpha [L_1 - T_1]$$

$$\frac{d^2}{dt^2} u_3 = \alpha [L_3 - T_3]$$

where $\alpha \equiv \mathcal{A} ((1/M^+) + (1/M^-)), \mathcal{A} = \Delta x_1 \Delta x_3$. $T_i$ express on the governing law.
Observational constraints

1) Perturbed rupture time $t_r = 25.9 \pm 0.1$ s

2) Hypocenter $(63.951 \pm 0.004$ °N, $21.689 \pm 0.008$ °W, $8.9 \pm 1.3$ Km) ↔ on fault coordinates of $(16500 \pm 450, 8900 \pm 1300)$ m (Antonioli et al., 2005)

3) From the aftershocks distribution shown in Hjaltadottir and Vogfjord (2005) we consider the seismic part of the fault ($A$) limited in latitude between $63.890$ °N and $63.951$ °N (in the case of Nord–South fault this corresponds to [9700, 16500] m in strike direction) and limited in depth between 5400 m and 7400 m

Upper bound estimates:
$M_0 = 1.23 \times 10^{15} A^{3/2} = 6.15 \times 10^{16}$ Nm;
Av. fault slip: $<u>_{A} = M_0/(\rho \nu_s^2 A) = 0.12$ m;
Av. stress drop: $<\Delta \tau>_{A} = 2M_0/(\pi W_A L_A) = 1.44$ MPa

4) $M_w \geq 5$ (Arnadottir et al., 2006; Vogfiord, 2003) ⇒ $M_0 \cong 3.2 \times 10^{16}$ Nm
3-D Results with DR law – homogeneous

Dieterich – Ruina governing law

$$\tau = \mu(v, \Psi)\sigma_n^{\text{eff}} = \left[ \mu + b \ln \left( \frac{v}{v_s} \right) + a \ln \left( \frac{\Psi}{v_s} \right) \right] \sigma_n^{\text{eff}}$$

$$\frac{d}{dt} \Psi = 1 - \frac{\Psi v}{L}$$

Can be neglected (see Antonioli et al., 2005)

$\sigma_n^{\text{eff}} = 2.5 \text{ MPa everywhere; acting only } \Delta\sigma_{21}$

Perturbed rupture times

$$v(x_1, x_3, t) \geq v_l \Rightarrow t_p(x_1, x_3) = t$$

$v_l = 0.1 \text{ m/s, in agreement with Belardinelli et al. (2003); Antonioli et al. (2005); Rubin and Ampuero (2005); Ziv and Cochard (2006)}$

$t_p^{\text{min}} = 23.47 \text{ s @ (20700,2900) m}$

$M_0 = 2.37 \times 10^{19} \text{ Nm}$

Whole fault

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
Results with DR law – homogeneous

Dieterich – Ruina governing law

\[
\tau = \mu(v, \Psi)\sigma_{n, \text{eff}}^{\text{eff}} = \left[ \mu_* + a \ln \left( \frac{v}{v_*} \right) + b \ln \left( \frac{\Psi v_*}{L} \right) \right] \sigma_{n, \text{eff}}^{\text{eff}}
\]

\[
\frac{\text{d} \Psi}{\text{d} t} = 1 - \frac{v \Psi}{L}
\]

Can be neglected (see Antonioli et al., 2005)

Perturbed rupture times

\[v(x_1, x_3, t) \geq v_l \Rightarrow t_p(x_1, x_3) = t\]

\[v_l = 0.1 \text{ m/s}, \text{ in agreement with Belardinelli at al. (2003); Antonioli et al. (2005); Rubin and Ampuero (2005); Ziv and Cochard (2006)}\]

\[t_p^{\text{min}} = 23.47 \text{ s } @ (16500, 2900) \text{ m}\]

\[M_0 = 2.23 \times 10^{19} \text{ Nm}\]

Whole fault

\[\sigma_{n, \text{eff}} = 2.5 \text{ MPa everywhere; acting also } \Delta \sigma_2^2\]

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
Results with DR law – heterogeneous

Velocity strengthening behavior \((a > b)\) for 
\(x_1 < 9700\) m, 
\(x_1 > 16500\) m, 
\(x_3 > 8800\) m

Effective normal stress profile

\[ t_{p\min} = 24.94\ s \ @ \ (13200,7500)\ m \]

\[ M_0 = 2.27 \times 10^{16}\ Nm \]

[9700,16500] m in strike direction

[6400,7500] m in dip direction

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
Instability at $t = t_p^{\min} = 24.94$ s

NO instability

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
Results with RD law – heterogeneous

Ruina – Dieterich governing law

\[ \tau = \left[ \mu_n + a \ln \left( \frac{v}{v_n} \right) + b \ln \left( \frac{\Psi v}{L} \right) \right] \sigma_n^{\text{eff}} \]

\[ \frac{d}{dt} \Psi = -\frac{\Psi v}{L} \ln \left( \frac{\Psi v}{L} \right) \]

Can be neglected

\( t_p^{\text{min}} = 23.44 \text{ s} @ (15700,7900) \text{ m} \)

\( M_0 = 2.02 \times 10^{16} \text{ Nm} \)

[9000,17300] m in strike direction

[6300,8000] m in dip direction

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
In the “virtual” hypocenter

Dieterich – Ruina governing law

\[ v_H^d = 0.01 \text{ m/s } (t = 24.56 \text{ s}) \]

\[ v_H^d = 0.05 \text{ m/s } (t = 24.84 \text{ s}) \]

\[ v_H^d = v_t = 0.1 \text{ m/s } (t = t_p = 24.94 \text{ s}) \]

Failure occurs before traction reaches the residual level.

Ruina – Dieterich governing law

RD with \( L = 5 \text{ mm} \):

\[ t_{p \min} = 23.99 \text{ s } @ (14600,7600) \text{ m} \]

\[ M_0 = 1.27 \times 10^{16} \text{ Nm} \]

[9500,16800] m in strike direction

[6500,7700] m in dip direction

RD with \( L = 10 \text{ mm} \):

\[ t_{p \min} = 24.72 \text{ s } @ (13300,7300) \text{ m} \]

\[ M_0 = 2.27 \times 10^{16} \text{ Nm} \]

[9500,16700] m in strike direction

[6000,7400] m in dip direction

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
**Alternative source time functions**

**Bouchon source time function:**

\[ f(t) = \frac{1}{2} \left[ 1 + \tanh \left( \frac{t - t_0}{2} \right) \right] \]

Bouchon, 1981; \( t_0 = 1.6 \) s

**Modified Bouchon source time function:**

\[ f(t) = \frac{1}{2} \left[ 1 + \tanh \left( \frac{t - t_0}{2} \right) \right] \]

corrected from Cotton and Campillo, 1995; \( t_0 = 1.6 \) s

\[ f(t) = \frac{1}{2} \left[ 1 + \tanh \left( \frac{t - t_0}{2} \right) \right] \]

corrected from Cotton and Campillo, 1995; \( t_0 = 3.2 \) s
Alternative source time functions

Bouchon modificata, \( t_0 = 3.2 \) s

\[ t_p^{min} = 26.49 \text{ s} @ (13000,7500) \text{ m} \]

\[ M_0 = 2.30 \times 10^{16} \text{ Nm} \]

[9700,16500] m in strike direction

[6400,7600] m in dip direction

Bouchon modificata, \( t_0 = 1.6 \) s;

\[ \sigma_{n}^{eff} = 4.2 \text{ MPa} \]

\[ t_p^{min} = 25.36 \text{ s} @ (13500,7600) \text{ m} \]

\[ M_0 = 2.59 \times 10^{16} \text{ Nm} \]

[9500,16700] m in strike direction

[6200,8700] m in dip direction

From Bizzarri and Belardinelli (Nov. 2005; subm. to JGR)
We simulate the remote triggering in a truly 3–D fault model with different governing laws;

We generalize the results of Antonioli et al. (2006), providing additional details of the 26 s event: the location of the hypocenter, its failure time, the rupture area and the seismic moment;

The spring–slider and the 3–D model are intrinsically different, but we observe an excellent agreement during the slow nucleation phase…

… during the acceleration, in the 3–D model the dynamic load of the slipping points further decrease the perturbed failure time;

Dieterich–Ruina and Ruina–Dieterich laws are valid candidate to model the activation of the Hvalhnúkur fault at 26 s;
✓ On the contrary, with slip–dependent friction laws it is not possible to simulate the activation of the 26 s aftershock;
✓ The agreement with observations increases considering a modified (and more causal) source time function;
✓ If a detailed information of the initial state of the fault, potentially highly heterogeneous, was available the agreement with observations will be even better.
<table>
<thead>
<tr>
<th>Case</th>
<th>$\sigma_{\gamma 0}$ profile</th>
<th>Constitutive law</th>
<th>Heterogeneous rheology</th>
<th>Rupture extension along strike (m)</th>
<th>Rupture extension along dip (m)</th>
<th>Hypocenter location (m)</th>
<th>Origin time (s)</th>
<th>Total seismic moment $M_0$ (Nm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A</td>
<td>2</td>
<td>DR</td>
<td>No</td>
<td>Whole fault</td>
<td>Whole fault</td>
<td>(20700,2900)</td>
<td>23.47</td>
<td>$2.37 \times 10^{19}$</td>
</tr>
<tr>
<td>B</td>
<td>2</td>
<td>DR</td>
<td>No</td>
<td>Whole fault</td>
<td>Whole fault</td>
<td>(16500,2900)</td>
<td>23.47</td>
<td>$2.23 \times 10^{19}$</td>
</tr>
<tr>
<td>C</td>
<td>1</td>
<td>DR</td>
<td>No</td>
<td>[0, 27400]</td>
<td>[6000, 11600]</td>
<td>(15400,6600)</td>
<td>24.08</td>
<td>$1.94 \times 10^{19}$</td>
</tr>
<tr>
<td>D</td>
<td>2</td>
<td>DR</td>
<td>No</td>
<td>Not defined</td>
<td></td>
<td></td>
<td></td>
<td>$1.21 \times 10^{14}$</td>
</tr>
<tr>
<td>E</td>
<td>3</td>
<td>DR</td>
<td>No</td>
<td>[6600, 20000]</td>
<td>[6400, 7500]</td>
<td>(13200,7500)</td>
<td>24.94</td>
<td>$6.43 \times 10^{16}$</td>
</tr>
<tr>
<td>F</td>
<td>3</td>
<td>DR</td>
<td>Yes</td>
<td>[9700, 16500]</td>
<td>[6400, 7500]</td>
<td>(13200,7500)</td>
<td>24.94</td>
<td>$2.27 \times 10^{16}$</td>
</tr>
<tr>
<td>G</td>
<td>3</td>
<td>DR</td>
<td>No</td>
<td>[15700, 35100]</td>
<td>[6000, 7800]</td>
<td>(27300,7500)</td>
<td>23.44</td>
<td>$1.22 \times 10^{17}$</td>
</tr>
<tr>
<td>H</td>
<td>3</td>
<td>RD</td>
<td>Yes</td>
<td>[9000, 17300]</td>
<td>[6300, 8000]</td>
<td>(15700,7900)</td>
<td>23.44</td>
<td>$2.02 \times 10^{16}$</td>
</tr>
<tr>
<td>I</td>
<td>3</td>
<td>RD</td>
<td>Yes</td>
<td>$L = 5$ mm</td>
<td>[9500, 6800]</td>
<td>(14600,7600)</td>
<td>23.99</td>
<td>$1.27 \times 10^{16}$</td>
</tr>
<tr>
<td>L</td>
<td>3</td>
<td>RD</td>
<td>Yes</td>
<td>$L = 10$ mm</td>
<td>[9500, 6700]</td>
<td>(13300,7300)</td>
<td>24.72</td>
<td>$2.17 \times 10^{16}$</td>
</tr>
<tr>
<td>M</td>
<td>3</td>
<td>OY</td>
<td>Yes</td>
<td>Not defined</td>
<td></td>
<td></td>
<td></td>
<td>$1.46 \times 10^{14}$</td>
</tr>
<tr>
<td>N</td>
<td>3</td>
<td>OY</td>
<td>No</td>
<td>Whole fault</td>
<td>Whole fault</td>
<td>(24000,7700)</td>
<td>23.75</td>
<td>$2.49 \times 10^{19}$</td>
</tr>
<tr>
<td>O</td>
<td>3</td>
<td>DR</td>
<td>Yes</td>
<td>[9700, 16500]</td>
<td>[6400, 7600]</td>
<td>(13000,7500)</td>
<td>26.49</td>
<td>$2.30 \times 10^{16}$</td>
</tr>
<tr>
<td>P</td>
<td>3</td>
<td>DR</td>
<td>Yes</td>
<td>[9500, 16700]</td>
<td>[6200, 8700]</td>
<td>(13500,7600)</td>
<td>25.36</td>
<td>$2.59 \times 10^{16}$</td>
</tr>
</tbody>
</table>

**Observational constraints**

[$9700, 16500$] [$5400, 7400$] $(16500 \pm 450, 8900 \pm 1300)$ $25.9 \pm 0.1$ $\approx 3.2 \times 10^{14}$
This slide is empty intentionally.
Support Slides:
Parameters, Notes, etc.

To not be displayed directly. Referenced above.
Figure 1. Geothermal areas in Iceland. The five main exploited high-temperature areas, Svartsengi, Reykjanes, Nesjavellir, Krýsuvík and Óskjuhlíð are shown as well as the four unexploited high-temperature geothermal areas selected for study of natural changes, Krýsuvík, Þraustareyki, Torfajökull and Kverðjöll areas.
<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>( \varphi )</td>
<td>parallelepiped that extends ( x_{1_{end}} = 36.5 ) Km along ( x_1 ), ( x_{2_{end}} = 10 ) Km along ( x_2 ) and ( x_{3_{end}} = 11.6 ) Km along ( x_3 )</td>
</tr>
<tr>
<td>( \Sigma = \Omega \setminus \delta )</td>
<td>{ ( x \mid x_2 = x_2^J = 5000 ) m }</td>
</tr>
<tr>
<td>( \Delta x_1 = \Delta x_2 = \Delta x_3 \equiv \Delta x )</td>
<td>100 m</td>
</tr>
<tr>
<td>Number of nodes</td>
<td>4,289,571</td>
</tr>
<tr>
<td>( \Delta t )</td>
<td>( 1.27 \times 10^{-3} ) s</td>
</tr>
<tr>
<td>Number of time levels</td>
<td>33,650</td>
</tr>
<tr>
<td>( v_r )</td>
<td>0.1 m/s</td>
</tr>
<tr>
<td>( \sigma_n^{eff} )</td>
<td>2.5 MPa</td>
</tr>
<tr>
<td>( \varphi(x_1, x_3, 0) )</td>
<td>( \varphi_0 = 180^\circ )</td>
</tr>
<tr>
<td>( v(x_1, x_3, 0) )</td>
<td>( v_{init} = 6.34 \times 10^{-10} ) m/s ( = 20 mm/yr )</td>
</tr>
<tr>
<td>( \Psi(x_1, x_3, 0) )</td>
<td>( \Psi^{eff}(v_{init}) = 1.577 \times 10^6 ) s ( ( \approx 18.25 ) d )</td>
</tr>
<tr>
<td>( \sigma_n^{eff}(x_1, x_3, 0) )</td>
<td>See Table 3</td>
</tr>
<tr>
<td>( \tau_n(x_1, x_3) )</td>
<td>( \mu^* (v_{init}) \sigma_n^{eff}(x_1, x_3, 0) )</td>
</tr>
<tr>
<td>( \alpha )</td>
<td>0.003</td>
</tr>
<tr>
<td>( b )</td>
<td>0.010</td>
</tr>
<tr>
<td>( L )</td>
<td>( 1 \times 10^{-2} ) m</td>
</tr>
<tr>
<td>( \mu_s )</td>
<td>0.7</td>
</tr>
<tr>
<td>( v_s )</td>
<td>( v_{init} )</td>
</tr>
<tr>
<td>( \sigma_{LD} )</td>
<td>0</td>
</tr>
</tbody>
</table>
Crustal profile (from Vogfjord et al., 2002; Antonioli et al., 2005)

<table>
<thead>
<tr>
<th>Layer #</th>
<th>$v_{P_k}$ (m/s)</th>
<th>$v_{S_k}$ (m/s)</th>
<th>$\rho_{rock_k}$ (Kg/m$^3$)</th>
<th>Up do depth of $x_{3_k}$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>3200</td>
<td>1810</td>
<td>2300</td>
<td>1100</td>
</tr>
<tr>
<td>2</td>
<td>4500</td>
<td>2540</td>
<td>2540</td>
<td>3100</td>
</tr>
<tr>
<td>3</td>
<td>6220</td>
<td>3520</td>
<td>3050</td>
<td>7800</td>
</tr>
<tr>
<td>4</td>
<td>6750</td>
<td>3800</td>
<td>3100</td>
<td>11600</td>
</tr>
</tbody>
</table>
Initial effective normal stress

\[
\sigma_{n_0}^{\text{eff}}(x_3) \equiv \sigma_n^{\text{eff}}(x_1, x_3, t) = \]

\[
\left\{ \begin{array}{l}
\hat{P}^{(\text{litho})}(x_3^*) - \Delta \sigma^{(\text{dev})} - P_{\text{fluid}}^{(\text{hyd})}(x_3) \\
\hat{P}^{(\text{litho})}(x_3^*) - \Delta \sigma^{(\text{dev})} - \left[ \hat{P}^{(\text{litho})}(x_3) - \Delta \sigma^{(\text{dev})} - \sigma_n^{\text{eff}} \right] \\
- \Delta P_2 e^{\frac{x_3 - x_3^*}{h}} + \sigma_n^{\text{eff}} e^{\frac{x_3 - x_3^*}{h}} \end{array} \right\}
\]

\[
\sigma_n^{\text{eff}} = 2.5 \text{ MPa}
\]

\[
\Delta P_2 = \hat{P}^{(\text{litho})}(x_3^*) - \Delta \sigma^{(\text{dev})} - P_{\text{fluid}}^{(\text{hyd})}(x_3^*)
\]

\[
, x_3 \leq x_3^* = 5800 \text{ m}
\]

\[
, x_3^* < x_3 < x_3^* + D^*
\]

\[
, x_3 \geq x_3^* + D^* = 8800 \text{ m}
\]