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# 1 Erosion influences the seismicity of active thrust faults

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10

11 **Assessing seismic hazards remains one of the most challenging scientific issues in Earth**  
12 **sciences. Deep tectonic processes are classically considered as the only persistent**  
13 **mechanism driving the stress loading of active faults over a seismic cycle. Here we show via**  
14 **a mechanical model that erosion also significantly influences the stress loading of thrust**  
15 **faults at the timescale of a seismic cycle. Indeed, erosion rates of about  $\sim 0.1$  to  $20 \text{ mm.yr}^{-1}$ ,**  
16 **as documented in Taiwan and in other active compressional orogens, can raise the**  
17 **Coulomb stress by  $\sim 0.1$  to  $\sim 10$  bar on the nearby thrust faults over the inter-seismic phase.**  
18 **Mass transfers induced by surface processes in general, during continuous or short-lived**  
19 **and intense events, represent a prominent mechanism for inter-seismic stress loading of**  
20 **faults near the surface. Such stresses are probably sufficient to trigger shallow seismicity or**  
21 **promote the rupture of deep continental earthquakes up to the surface.**

22

23 **INTRODUCTION**

24           The evolution of the Earth's topography is dictated by the interactions between tectonics,  
25 climate and surface processes (i.e. erosion and sedimentation). Whether this evolution influences  
26 tectonic deformation during mountain building has been widely debated. It is now well accepted  
27 that surface evolution can drive the localization and intensity of tectonic deformation over  
28 geological times<sup>1-3</sup> (1-10 Myr). At intermediate time scales (10 kyr - 1 Myr), erosion and the  
29 subsequent isostatic rebound can favour slip along specific fault planes<sup>4-7</sup>. However, the link  
30 between surface processes and the stress loading of faults during the seismic cycle (0.1-1 kyr),  
31 and in turn the associated deformation mechanisms, remains unsubstantiated.

32           Faults represent the main mechanical discontinuities of the elastic-brittle Earth's upper  
33 crust. They accommodate tectonic deformation by slipping, mostly during earthquakes<sup>8</sup>. These  
34 seismogenic faults are rooted down dip in viscous shear zones<sup>8-10</sup>. It is generally accepted that,  
35 during the inter-seismic phase (i.e. prior to an earthquake), continuous viscous flow in these deep  
36 shear zones leads to the elastic stress loading of active faults closer to failure, and that during the  
37 co-seismic phase (i.e. during an earthquake), failure and slip occur along the previously locked  
38 fault planes, followed by post-seismic stress relaxation<sup>4,8</sup>. Fault failure is commonly defined by  
39 the mean of the Coulomb stress change,  $\Delta CFF = \Delta\tau + \mu' \cdot \Delta\sigma_n$ , a function of the fault effective  
40 friction  $\mu'$ , the shear  $\Delta\tau$  (positive in the direction of slip) and normal  $\Delta\sigma_n$  (positive if the fault is  
41 unclamped) stress changes<sup>8,11</sup>. Earthquakes can be triggered by tectonic stresses, but also by  
42 Coulomb stresses due to episodic and short-lived events such as hydrologic<sup>12</sup> or snow loading<sup>13</sup>,  
43 nearby earthquakes<sup>14-17</sup> and slow-slip events<sup>18</sup>.

44           Here, we show that surface processes significantly contribute to the Coulomb stress  
45 loading of thrust faults during the seismic cycle. To illustrate and then demonstrate our point, we  
46 consider a mountain range in Taiwan where the rates of erosion<sup>19</sup> and tectonic deformation<sup>20,21</sup>

47 are extremely high and amongst the best-documented in the world. We then investigate how  
48 erosion influences the stress loading of thrust faults using a simple model for the seismic cycle.

49

## 50 **RESULTS**

### 51 **Coulomb stress changes induced by erosion in Taiwan**

52 Our first model quantifies the Coulomb stress change  $\Delta CFF$  generated by erosional  
53 unloading, as constrained from fluvial suspended sediment load measured over the 30 yr prior to  
54 the 1999  $M_w7.6$  Chi-Chi earthquake in central Taiwan<sup>19</sup> (Fig. 1 and Methods). The 3D velocity  
55 field  $\mathbf{v}$ , strain rate  $\dot{\boldsymbol{\epsilon}}$  and stress rate  $\dot{\boldsymbol{\sigma}}$  tensors induced by erosion are computed in an elastic half-  
56 space using a Boussinesq approach (Fig. 1C and Methods). We use simplified geometries for  
57 active thrust faults located in the foothills of Taiwan<sup>22</sup> (see Methods) and assume a dip angle  $\alpha$  of  
58  $30^\circ$ , a 15 km deep brittle–ductile transition<sup>22</sup> and an effective friction  $\mu'$  of 0.5 to compute  
59 Coulomb stress changes per unit time (or loading rates)  $\Delta CFF$  due to erosional unloading on  
60 these faults. We find a maximum value of  $\sim 4 \times 10^{-3}$  bar.yr<sup>-1</sup> for the Coulomb stress change  $\Delta CFF$   
61 induced by erosional unloading on the Liuchia fault system (number 8 on Figure 1) in  
62 southwestern Taiwan. Despite a low topographic relief, this area has the highest erosion rates  
63 documented in Taiwan (up to 24 mm.yr<sup>-1</sup>), which are proposed to be controlled locally by a low  
64 substrate strength, a high storminess and a high seismic moment release rate<sup>19</sup>. However, most of  
65 the thrust faults located in the foothills still display a significant  $\Delta CFF$  of  $\sim 0.5 \times 10^{-3}$  bar.yr<sup>-1</sup>,  
66 including the Chelungpu fault (number 3) that ruptured during the Chi-Chi earthquake.  
67 Integrated over a seismic cycle duration<sup>23</sup> of  $\sim 500$  yr, a  $\Delta CFF$  of 0.5 to  $4 \times 10^{-3}$  bar.yr<sup>-1</sup> due to  
68 erosional unloading gives a net Coulomb stress change of 0.25–2.0 bar. Similar values of  
69 Coulomb stress change are documented elsewhere to contribute significantly to the stress loading

70 and dynamics of active faults<sup>12-18</sup>. This suggests that erosional unloading can significantly  
71 influence the short-term dynamics of faults.

72 Erosional unloading modifies the Coulomb stress change on a fault plane in two ways: 1)  
73 it decreases the normal stress and unclamps the fault, and 2) it increases the tangential stress (fig.  
74 2a). The increment of stress on a fault plane is proportional to the amount of erosion, but  
75 decreases with the square of the distance  $r$  between the fault plane and where erosion occurs (fig.  
76 2b). Therefore, the amplitude of erosional Coulomb stress loading on a fault is sensitive 1) to the  
77 effective friction  $\mu'$ , which modulates the effect of erosion on the normal stress, and 2) to the  
78 fault dip angle  $\alpha$ , which decomposes the stresses into fault normal and tangential components  
79 and controls the distance of the fault plane to the surface. For instance, a higher effective friction  
80  $\mu'$  of 0.8 and a lower dip angle  $\alpha$  of  $15^\circ$  therefore result in increasing the induced  $\Delta CFF$  up to  $\sim 1$   
81  $\times 10^{-2}$  bar.yr<sup>-1</sup> for the Liuchia fault system (Fig. 2c). In addition, the stresses modelled here are  
82 invariant with the Young modulus of the material as we are considering a linear elastic material  
83 subjected to a surface pressure load (and not to a surface displacement).

#### 84 **Stresses induced by erosion during the seismic cycle**

85 The above computations consider inter-seismic erosion rates calculated from data  
86 acquired during the 30 yr preceding the Chi-Chi earthquake<sup>19</sup>. Even though its amplitude relative  
87 to co-seismic rock uplift is debated, co- and post-seismic erosional unloading represents a major  
88 contribution to erosion in seismic areas<sup>24-27</sup>. In mountain belts with hillslopes close to failure, co-  
89 seismic ground motion and acceleration can induce a significant amount of landslides<sup>28</sup>. The  
90 sediments produced by these landslides are then transported by rivers mainly during subsequent  
91 floods. This post-seismic landscape relaxation phase has a documented potential duration<sup>25,26</sup> of  
92 years to decades, one order of magnitude shorter than a complete seismic cycle. Therefore, the

93 contribution of co- and post-seismic erosion to the stress loading of active faults also needs to be  
94 evaluated.

95 To assess the relative contribution of inter-seismic erosion, co-/post-seismic erosion and  
96 tectonics to the Coulomb stress loading of faults, we develop a simple model of the seismic cycle  
97 that accounts for the effect of both erosion and tectonics (Fig. 3 and Methods). We assume a  
98 steady-state landscape over the seismic cycle, i.e. rock uplift rates  $\dot{U}$  are balanced by erosion  
99 rates  $\dot{E}$  over this time scale. Because of the response-time of the geomorphic system to climate or  
100 tectonic perturbations and because of their stochastic properties<sup>29</sup>, this assumption is probably  
101 not valid in most settings. However, it offers a simple and self-consistent approach for modelling  
102 first-order surface processes during the seismic cycle. A Boussinesq approach is used to compute  
103  $\mathbf{v}$ ,  $\dot{\boldsymbol{\epsilon}}$  and  $\dot{\boldsymbol{\sigma}}$ , while the tectonic stresses and uplift (and therefore erosion) are calculated using  
104 dislocations embedded in an elastic half-space<sup>30</sup>. The effects of tectonic deformation during the  
105 inter- and co-seismic phases are accounted for by slip on a deep shear zone and on a shallow  
106 brittle fault, respectively<sup>9,31</sup>. This seismic cycle model is valid when considering a fault that is  
107 fully locked during the inter-seismic phase, as it is proposed for the thrust faults located in the  
108 western foothills of Taiwan<sup>32-34</sup>.

109 For comparison with faults in the foothills of Taiwan, we define a reference model with  
110 a fault trace length of 80 km and a dip angle of 30°, while keeping all the other mechanical  
111 properties identical. We also impose a slip velocity  $V_{\text{inter}}$  of 40 mm.yr<sup>-1</sup> on the shear zone during  
112 the inter-seismic phase, and  $V_{\text{co}}$  of 40 mm.yr<sup>-1</sup> on the associated brittle fault during the co-  
113 seismic phase. This model setup provides only a rough approximation of the seismic cycle at the  
114 scale of the whole western foothills of Taiwan. Indeed, deformation is partitioned between  
115 several active thrust faults in this area that are probably rooted down at depth into a single

116 decollement with a total slip of  $\sim 40 \text{ mm.yr}^{-1}$  (refs. 21, 33). In addition, because our goal is to  
117 quantify the co- and post-seismic erosional unloading rates during the landscape relaxation phase  
118 following large earthquakes<sup>25,26</sup>, we compute co-seismic slip velocity averaged over the seismic  
119 cycle rather than co-seismic instantaneous displacement. Note that these two approaches are  
120 strictly equivalent in an elastic model.

121 In our modelling, inter- and co-seismic rock uplift (and erosion) rates are similar and up  
122 to  $\sim 20 \text{ mm.yr}^{-1}$  (Fig. 3). We assume no time modulation of inter- and co-seismic erosion and  
123 both modeled erosion rates are applied over the entire duration of one seismic cycle. Despite  
124 similar erosion rates, co-seismic erosional unloading induces fault Coulomb stress loading  
125  $\Delta CFF_{E-CO}$  of up to  $\sim 8 \times 10^{-2} \text{ bar.yr}^{-1}$  at very shallow depth ( $< 1 \text{ km}$ ), which is about 30 times the  
126 maximum stress loading induced by inter-seismic erosion  $\Delta CFF_{E-INTER}$ . Indeed, the maxima of  
127 co-seismic uplift of the surface (and erosion) occurs over the shallow portion of the fault,  
128 therefore at a shorter distance to the fault plane than the maxima of inter-seismic erosion (Fig. 3).  
129 Despite a rapid decrease with depth,  $\Delta CFF_{E-CO}$  is still greater than  $\sim 0.5 \times 10^{-2} \text{ bar.yr}^{-1}$  at a depth  
130 of 5 km. Coulomb stress loading  $\Delta CFF_{E-INTER}$  induced by inter-seismic erosion displays two  
131 local maxima of  $\sim 0.3 \times 10^{-2} \text{ bar.yr}^{-1}$ , one located on the deeper part of the fault underneath the  
132 maximum of inter-seismic erosion, and the second one close to the fault tip due to its proximity  
133 with the surface.

134 We then compare fault Coulomb stress loading rates induced by erosion to those induced  
135 only by tectonics during the inter-seismic phase  $\Delta CFF_{T-inter}$ . In terms of amplitude, Coulomb  
136 stress loading rates due to tectonics are up to two or four orders of magnitude greater than those  
137 related to erosion, in particular for the deeper part of the fault. However, at shallower depths ( $< 5$   
138 km), Coulomb stress loading rates due to co-seismic erosion and tectonics are of the same order

139 of magnitude, and the ratio  $\Delta CFF_{E-co}/\Delta CFF_{T-inter}$  even reaches  $\sim 20$  close to the surface (Fig.3).  
140 At the contrary, Coulomb stress loading rates associated with inter-seismic erosion, which is  
141 maximum on the deeper and shallower part of the fault, do not dominate tectonic stresses. The  
142 ratio  $\Delta CFF_{E-inter}/\Delta CFF_{T-inter}$  only reaches  $\sim 0.1$  at intermediate depths (5-10 km) and  $\sim 0.6$  close to  
143 the surface. Because the upper crust displays a very long stress relaxation time associated to high  
144 effective viscosities, these Coulomb stress changes induced by erosion can be accumulated over  
145 the time scale of a seismic cycle ( $\sim 500$  yr). On the other hand, increasing the Young modulus  
146 from the reference value (10 GPa) by a factor of 2 (20 GPa) or 5 (50 GPa) results in increasing  
147 the Coulomb stress change induced by tectonics  $\Delta CFF_{T-inter}$  by a factor of 2 or 5 and decreasing  
148 the ratios  $\Delta CFF_{E-inter}/\Delta CFF_{T-inter}$  and  $\Delta CFF_{E-co}/\Delta CFF_{T-inter}$  by the same factor (Fig. 3h). However,  
149 these results still demonstrate 1) that erosion can contribute significantly to thrust fault stress  
150 loading during the seismic cycle, and 2) that erosion can even be one of the dominant stress  
151 loading mechanisms for the shallower parts of thrust fault planes.

## 152 **Model sensitivity analysis**

153 To assess the sensitivity of our results to the model parameters, we design a set of models  
154 similar to the reference model but with varying values of the effective friction  $\mu'$  (0.1 to 0.9), the  
155 Young modulus  $E$  (10, 20 and 50 GPa) and the fault and shear zone dip angle  $\alpha$  (15 to 45°). For  
156 the sake of simplicity we keep the depth of the brittle–ductile transition at 15 km, which in turn  
157 implies that the surface area of the brittle fault increases when decreasing  $\alpha$ . Using this  
158 modelling approach, erosion rates during the inter-seismic  $\dot{E}_{inter}$  and the co-seismic  $\dot{E}_{co}$  phases are  
159 only sensitive to  $\alpha$  (Fig. 4a-b). While  $\dot{E}_{inter}$  remains approximately constant around 13 mm.yr<sup>-1</sup>,  
160  $\dot{E}_{co}$  increases from 11 to 22 mm.yr<sup>-1</sup> when  $\alpha$  increases from 15 to 45°.



161 The resulting Coulomb stresses  $\Delta CFF_{E\text{-inter}}$  and  $\Delta CFF_{E\text{-co}}$  on the fault plane are in turn  
 162 sensitive to 1) the dip angle  $\alpha$ , which controls both the distribution and amplitude of  $\dot{E}_{\text{inter}}$  and  $\dot{E}_{\text{co}}$   
 163 and the distance between the Earth surface and the fault plane, and 2) the effective friction  $\mu'$  by  
 164 amplifying the influence of normal stress on the Coulomb stress (Fig. 4c-d). The maximum  
 165 values of  $\Delta CFF_{E\text{-inter}}$  and  $\Delta CFF_{E\text{-co}}$  obtained on the fault plane show similar distributions in the  
 166 parameter space.  $\Delta CFF_{E\text{-inter}}$  is minimum ( $\sim 1 \times 10^{-3} \text{ bar.yr}^{-1}$ ) for a low effective friction ( $\mu' \leq 0.2$ )  
 167 combined to a high or low dip angle ( $15^\circ$  or  $45^\circ$ ), while it increases when increasing  $\mu'$  and  
 168 reaches a maximum ( $\sim 2.7 \times 10^{-3} \text{ bar.yr}^{-1}$ ) for  $\alpha$  around  $25^\circ$ .  $\Delta CFF_{E\text{-co}}$  is minimum ( $\sim 1.5 \times 10^{-2}$   
 169  $\text{bar.yr}^{-1}$ ) for a low effective friction ( $\mu' \leq 0.2$ ) combined to a high or low dip angle ( $15^\circ$  or  $45^\circ$ ),  
 170 while it increases when increasing  $\mu'$  and reaches a maximum of  $\sim 4 \times 10^{-2} \text{ bar.yr}^{-1}$  for  $\alpha$  around  
 171  $30^\circ$ .

172 Because the Coulomb stresses induced by tectonics  $\Delta CFF_{T\text{-inter}}$  are also sensitive to the  
 173 Young modulus  $E$ , the ratios  $\Delta CFF_{E\text{-inter}}/\Delta CFF_{T\text{-inter}}$  and  $\Delta CFF_{E\text{-co}}/\Delta CFF_{T\text{-inter}}$  are in turn  
 174 sensitive to  $\alpha$ ,  $\mu'$  and  $E$ . Increasing the Young modulus from 10 to 20 or 50 GPa increases  
 175  $\Delta CFF_{T\text{-inter}}$  and decreases  $\Delta CFF_{E\text{-inter}}/\Delta CFF_{T\text{-inter}}$  and  $\Delta CFF_{E\text{-co}}/\Delta CFF_{T\text{-inter}}$  by a factor of 2 or 5,  
 176 respectively.  $\Delta CFF_{E\text{-inter}}/\Delta CFF_{T\text{-inter}}$  remains lower than 1 only for all the models tested,  
 177 independent of the Young modulus  $E$ . At the contrary,  $\Delta CFF_{E\text{-co}}/\Delta CFF_{T\text{-inter}}$  displays a large  
 178 domain in the parameter space with values greater than 1 (i.e. with at least one element of the  
 179 fault plane dominated by stresses induced by erosion). For  $E=10$  GPa, only models with low  $\mu'$   
 180 ( $< 0.5$ ) and high  $\alpha$  ( $> 35^\circ$ ) are dominated by tectonic stresses ( $\Delta CFF_{E\text{-co}}/\Delta CFF_{T\text{-inter}} < 1$ ), while for  
 181  $E=50$  GPa, most of the models with  $\alpha$  greater than  $20\text{-}30^\circ$  are dominated by tectonic stresses.  
 182 Therefore, Coulomb stresses induced by erosion during the seismic cycle represent a significant

183 contribution to fault stress loading, even though their amplitude depends on the properties of the  
184 fault (dip angle, effective friction) and of the medium (Young modulus).

185

## 186 **DISCUSSION**

187 Based on erosion data from Taiwan<sup>19</sup>, we have demonstrated that the elastic Coulomb  
188 stresses induced by erosion are of the order of  $\sim 0.5 \times 10^{-3} \text{ bar.yr}^{-1}$  on the thrust faults located in  
189 the western foothills and reach a maximum of  $\sim 4 \times 10^{-3} \text{ bar.yr}^{-1}$  on the Liuchia fault. These  
190 results are consistent with the outcomes from a simple model of the seismic cycle of a thrust  
191 fault that accounts for the effect of both erosion and tectonics using fault properties and a slip  
192 velocity close to the ones inferred for Taiwan<sup>20-22</sup>. Coulomb stresses induced by inter-seismic  
193  $\Delta CFF_{E\text{-inter}}$  and co-seismic  $\Delta CFF_{E\text{-co}}$  erosion and averaged over the duration of a seismic cycle  
194 ( $\sim 500 \text{ yr}$ ) reach values of up to  $\sim 3 \times 10^{-3} \text{ bar.yr}^{-1}$  and  $\sim 8 \times 10^{-2} \text{ bar.yr}^{-1}$ , respectively. On the  
195 shallower part of thrust faults ( $< 5 \text{ km}$  deep), the ratio of the Coulomb stresses induced by co-  
196 seismic erosion  $\Delta CFF_{E\text{-co}}$  to the ones induced by tectonic loading  $\Delta CFF_{T\text{-inter}}$  is about equal to 1  
197 for a Young modulus  $E=10 \text{ GPa}$  ( $\sim 0.2$  for  $E=50 \text{ GPa}$ ) and even reach a maximum of  $\sim 20$  closer  
198 to the surface ( $\sim 4$  for  $E=50 \text{ GPa}$ ). In addition, assuming that co-seismic erosion happens only  
199 during a period 10 to 100 times shorter<sup>25,26</sup> ( $\sim 5\text{-}50 \text{ yr}$ ) than the complete seismic cycle<sup>23</sup> ( $\sim 500$   
200  $\text{yr}$ ),  $\Delta CFF_{E\text{-CO}}$  increases up to  $80 \times 10^{-1}$  to  $8 \text{ bar.yr}^{-1}$  and the ratio  $\Delta CFF_{E\text{-co}}/\Delta CFF_{T\text{-inter}}$  to 10 or  
201 100 for  $E=10 \text{ GPa}$  (2 to 20 for  $E=50 \text{ GPa}$ ) during the first  $\sim 5\text{-}50$  years following a large  
202 earthquake. Large earthquakes with a potential negative mass balance (i.e. erosion greater than  
203 uplift), such as the Wenchuan earthquake<sup>24,27</sup>, could induce even higher rates of  $\Delta CFF_{E\text{-co}}$  than  
204 those predicted by a steady-state model. Our modelling approach imposes that co-seismic  
205 erosion is maximum close to the fault trace and above the shallower part of thrust faults (Fig. 3).

206 This result contrasts with most of the observed distribution of earthquake-triggered landslides,  
207 with a maximum of landslide density close to the epicentral area and therefore generally above  
208 the deeper part of thrust faults<sup>28</sup>. Therefore, depending on the location of large earthquake  
209 hypocenters along the fault plane, our estimates for the contribution of co-seismic erosion to  
210 fault stress loading might likely be overestimated.

211 However, our results emphasize that short-lived and intense erosional events associated  
212 with efficient sediment transport, such as typhoons, could suddenly increase the Coulomb stress  
213 of underlying faults. On longer time-scales, climatic changes or transition from fluvial- to  
214 glacial-dominated surface processes could also lead to high transient erosion rates and therefore  
215 to transient increases of the fault stress loading due to erosion<sup>4-7</sup>. The mechanism proposed in  
216 this study is limited neither to convergent settings nor to erosion only, as sediment deposition on  
217 the hanging wall of normal faults could also lead to a significant increase in Coulomb stress<sup>5</sup>.  
218 Moreover, some less active areas, such as intra-continental faults or old orogens still experience  
219 intense erosion and episodic seismic activity<sup>6,7</sup>. In the absence of major tectonic deformation,  
220 surface processes could significantly contribute to the stress loading of faults in these areas, even  
221 when considered independently of the stresses induced by isostatic rebound.

222 In summary, our results demonstrate that surface processes represent a significant  
223 contribution to the Coulomb stress loading of faults during the seismic cycle. In terms of  
224 deformation, these additional stresses on the shallower part of fault planes can induce and trigger  
225 shallow earthquakes, as illustrated by the seismicity triggered by the large 2013 Bingham  
226 Canyon mine landslide<sup>35</sup>, or potentially favour the rupture of large deeper earthquakes up to the  
227 surface as, for instance, during the Chi-Chi earthquake<sup>36</sup>. This offers new perspectives on the  
228 mechanisms influencing stress transfers during the seismic cycle, as well as on seismic hazard

229 assessment in areas experiencing rapid erosion. More generally, Coulomb stress loading of faults  
230 induced by surface processes over short time scales provides an additional positive feedback  
231 between climate, surface processes and tectonics.

232

## 233 **METHODS**

### 234 **Seismic cycle model**

235 The deformation model computes the velocity field  $\mathbf{v}$ , strain  $\dot{\boldsymbol{\epsilon}}$  and stress  $\dot{\boldsymbol{\sigma}}$  rate tensors induced  
236 by surface erosion in a 3D elastic half-space based on the Boussinesq approximation<sup>37</sup>. With  
237 respect to this approximation, we assume that the model surface is horizontal (Fig. 1). The effect  
238 of such assumption is here limited as the topography of the western foothills of Taiwan is  
239 globally lower than 1 km. The model is discretized by cubic cells with a 100 m-resolution, with a  
240 Young modulus of  $E=10$  GPa, a Poisson ratio of  $\nu=0.25$ , and a rock density of  $\rho=2800$  kg.m<sup>-3</sup>.  
241 Velocity, stresses and strain induced by tectonics are simulated by the mean of triangular  
242 dislocations<sup>30,31</sup> accounting for the slip 1) of a viscous deep shear zone during the inter-seismic  
243 phase and of 2) a frictional fault located in the shallow elastic-brittle crust during the co-seismic  
244 phase. The imposed averaged tangential velocities of the shear zone  $V_{\text{inter}}$  and of the fault  $V_{\text{co}}$  are  
245 both equal to 40 mm.yr<sup>-1</sup>. Coulomb stress changes are then computed by projecting the stresses  
246 due to surface processes and to tectonics on the fault plane that is discretized with a resolution of  
247 100 m. The extent of fault planes and the dip angle data of the western foothills of Taiwan were  
248 simplified from ref. 22. Each fault trace was simplified to a line segment that best reproduces the  
249 real fault trace geometry.

### 250 **Elastic Boussinesq model**

251 We here consider the displacements, stress and strain components generated by a point load  $F$  at  
 252 the surface of a 3D semi-infinite elastic solid of coordinates  $x$ ,  $y$  and  $z$ , with  $z$  being positive  
 253 downward. Let's define  $r = \sqrt{(x - x_0)^2 + (y - y_0)^2 + z^2}$  the distance to the point load of  
 254 location  $(x_0, y_0, z)$ ,  $\Delta x = x - x_0$  and  $\Delta y = y - y_0$ . The elasticity of the model is described by  
 255 the Lamé's first  $\lambda$  and second parameters  $\mu$ , which are related to the Young modulus  $E$  and the  
 256 Poisson ratio  $\nu$  by  $\lambda = E\nu / ((1 + \nu)(1 - 2\nu))$  and  $\mu = E / (2(1 + \nu))$ . For  $z \geq 0$ , the  
 257 displacement components are<sup>37,38</sup>:

$$U_x(x, y, z) = \frac{F}{4\pi} \left( \frac{\Delta x}{(\lambda + \mu)r(z + r)} - \frac{\Delta x z}{\mu r^3} \right)$$

$$U_y(x, y, z) = \frac{F}{4\pi} \left( \frac{\Delta y}{(\lambda + \mu)r(z + r)} - \frac{\Delta y z}{\mu r^3} \right)$$

$$U_z(x, y, z) = \frac{F}{4\pi} \left( \frac{(\lambda + 2\mu)}{\mu(\lambda + \mu)r} + \frac{z^2}{\mu r^3} \right)$$

258 Assuming infinitesimal deformation, the symmetric Cauchy strain tensor  $\varepsilon$  is then obtained by  
 259 differentiating the displacement vector,  $\varepsilon_{ij} = 1/2 (dU_i/dx_j + dU_j/dx_i)$ . For an isotropic  
 260 medium, the stress components are then given by the following equation,  $\sigma_{ij} = \lambda \delta_{ij} \varepsilon_{kk} + 2\mu \varepsilon_{ij}$ ,  
 261 where  $\delta_{ij}$  is the Kronecker delta. The 6 stress components are<sup>37,38</sup>:

$$\sigma_{xx}(x, y, z) = \frac{F}{2\pi} \left( \frac{3\Delta x^2 z}{r^5} + \frac{\mu(\Delta y^2 + \Delta z^2)}{(\lambda + \mu)r^3(z + r)} - \frac{\mu z}{(\lambda + \mu)r^3} - \frac{\mu \Delta x^2}{(\lambda + \mu)r^2(z + r)^2} \right)$$

$$\sigma_{yy}(x, y, z) = \frac{F}{2\pi} \left( \frac{3\Delta y^2 z}{r^5} + \frac{\mu(\Delta x^2 + \Delta z^2)}{(\lambda + \mu)r^3(z + r)} - \frac{\mu z}{(\lambda + \mu)r^3} - \frac{\mu \Delta y^2}{(\lambda + \mu)r^2(z + r)^2} \right)$$

$$\sigma_{zz}(x, y, z) = \frac{F}{2\pi} \left( \frac{3z^3}{r^5} \right)$$

$$\sigma_{xy}(x, y, z) = \frac{F}{2\pi} \left( \frac{3\Delta x \Delta y z}{r^5} - \frac{\mu \Delta x \Delta y (z + r^2)}{(\lambda + \mu)r^3(z + r)^2} \right)$$

$$\sigma_{xz}(x, y, z) = \frac{F}{2\pi} \left( \frac{3\Delta xz^2}{r^5} \right)$$

$$\sigma_{yz}(x, y, z) = \frac{F}{2\pi} \left( \frac{3\Delta yz^2}{r^5} \right)$$

262 Because the model is linear and elastic, the total displacement, stress and strain components for  
263 any distribution of surface load are then computed by summation of the displacement, stress and  
264 strain components obtained for each individual point load.

### 265 **Taiwan erosion rates**

266 Erosion rates in Taiwan were calculated from fluvial suspended sediment load measured over the  
267 30 yr prior to the 1999  $M_w$ 7.6 Chi-Chi earthquake, considering that the total fluvial sediment  
268 load contains 70% of suspended load<sup>9</sup> (Fig. 1b). Erosion rates were calculated assuming that  
269 suspended sediment load, measured at 130 gauging stations, represents 70% of the river total  
270 sediment load, and that catchment-wide erosion rates correspond to the total sediment load  
271 divided by sediment density and by drainage area<sup>19</sup>. The smoothed erosion map of Figure 1 was  
272 then obtained using a circular averaging window with a radius of 30 km. The spatial resolution  
273 of the erosion map, even though coarse, still allows for resolving potential heterogeneities of  
274  $\Delta CFF$  induced by erosion between different faults and along each individual fault plane.

275

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369

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376

### 377 **AUTHOR CONTRIBUTIONS**

378 P.S. analyzed the data and performed the modelling. All authors contributed equally to the  
379 design of the study and to the writing of the paper.

380

### 381 **ADDITIONAL INFORMATION**

382 The authors declare no competing financial interests. Supplementary information accompanies  
383 this paper. Correspondence and requests for materials should be addressed to P.S.

384

### 385 **FIGURE CAPTIONS**

386

387 **Figure 1.** Faults stress loading rates induced by erosion in the foothills of Taiwan. a)  
388 Topography, simplified main thrust fault systems (red lines and numbers from 1 to 8) and  
389 location of the 1999  $M_w$ 7.6 Chi-Chi earthquake epicenter (red star) that ruptured the Chelungpu  
390 fault (number 3). b) Inter-seismic erosion rates prior to the Chi-Chi earthquake, as calculated  
391 from fluvial suspended sediment measurements with a 5-km grid resolution, smoothed at the  
392 catchment scale using a circular moving mean with 30-km diameter<sup>19</sup>. c) Erosion induced

393 Coulomb stress loading rates  $\Delta CFF$  calculated on the fault planes. The horizontal solid black  
394 lines represent scale bars of 50 km.

395

396 **Figure 2.** Mechanism of Coulomb stress loading of a thrust fault by surface erosion. a)

397 Distribution of stress increment  $\Delta\sigma$  (here purely illustrative) induced by a punctual erosion at the  
398 surface, increasing both the tangential  $\Delta\tau$  (driving effect) and the normal stresses  $\Delta\sigma_n$

399 (unclamping effect). The white solid line indicates a scale bar of 2.5 km. b) The resulting fault

400 Coulomb stress change  $\Delta CFF$  decreases with the square of the vertical distance  $r$  between the

401 fault plane and where erosion occurs (assuming  $\alpha=30^\circ$  and 1 m of erosion). c) Sensitivity of

402  $\Delta CFF$  induced by erosion to the fault dip angle  $\alpha$  and effective friction  $\mu'$ , taking the Liuchia

403 fault system as an example. The black solid line indicates a scale bar of 25 km. The model in the

404 dashed green box is equivalent to the one in Figure 1 and  $\Delta\tau$  and  $\Delta\sigma_n$  are reported in

405 Supplementary Figure 1.

406

407 **Figure 3.** Erosional versus tectonic driven Coulomb stress loading of faults during the seismic

408 cycle. Model 3D geometry and modeled surface rock uplift rate  $U$  during the a) inter- and b) co-

409 seismic phases averaged over the entire seismic cycle. The averaged tangential velocities of the

410 shallow fault  $V_{co}$  and of the deep shear zone  $V_{inter}$  are equal to  $40 \text{ mm.yr}^{-1}$ . Erosional Coulomb

411 stress loading rates of the fault obtained by equating erosion rates  $\dot{E}$  to surface uplift rates  $\dot{U}$  for

412 the c) inter-seismic  $\Delta CFF_{E-inter}$  and d) co-seismic  $\Delta CFF_{E-co}$  phases. Ratios of e) inter-seismic and

413 f) co-seismic erosional Coulomb stress loading rates over g) the inter-seismic tectonic Coulomb

414 stress loading rate  $\Delta CFF_{T-inter}$ .  $\Delta\tau$  and  $\Delta\sigma_n$  are reported in Supplementary Figure 2. h) Spatial

415 variation of Coulomb stresses along the fault plane (cross-section  $\Omega-\Omega'$ ) considering different

416 Young moduli ( $E=10, 20$  and  $50$  GPa) that only influences the tectonics stresses and not the  
417 stresses induced by erosion.

418

419 **Figure 4.** Sensitivity of model results to the model parameters. Maximum surface erosion rates  
420 obtained during the a) inter-seismic  $\dot{E}_{\text{inter}}$  and the b) co-seismic  $\dot{E}_{\text{co}}$  periods as a function of the  
421 fault and shear zone dip angle  $\alpha$ . Maximum fault Coulomb stresses induced by erosion during c)  
422 inter-seismic  $\Delta CFF_{E\text{-inter}}$  and d) co-seismic  $\Delta CFF_{E\text{-co}}$  periods, which are function of the dip angle  
423  $\alpha$  and of the effective friction  $\mu'$ , are shown with a plain color map. White lines indicate the  
424 limits of tectonics (black arrow) vs erosion (white arrow) dominated Coulomb stresses in the  
425 model parameter space for varying values of the Young modulus  $E$  (10, 20 and 50 GPa). The  
426 reference model is represented by a white star. We consider that Coulomb stresses induced by  
427 erosion dominates tectonic stresses when at least one element of the fault is dominated by  
428 erosional stresses. For the inter-seismic erosion induced stresses, we here only consider the deep  
429 part of the fault.

430









