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Major Earthquakes on the Southern San Andreas Fault Modulated by Lake Filling Events

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4 Abstract

We use new geologic and paleoseismic data to demonstrate that the past 6 ma-5 jor earthquakes on the Southern San Andreas Fault (SSAF) correlate with 6 high-stands of the ancient Lake Cahuilla, a \sim 236 km³ body of water adjacent 7 to the SSAF. In order to investigate possible causal relationships, we computed 8 time-dependent Coulomb stress changes due to variations in the lake level over 9 the last \sim 1100 years. Simulations were performed using a fully coupled 3-D 10 finite element model incorporating a poroelastic crust overlying a viscoelas-11 tic mantle. We find that the Coulomb stress perturbations on the SSAF are 12 positive (i.e., promoting failure) throughout the lake loading history. For a 13 plausible range of lake ages and material properties of the Earth's crust, the 14 estimated stress perturbations are of the order of 0.5 MPa, likely sufficient for 15 triggering. Stress perturbations are dominated by pore pressure changes, but 16 are enhanced by the poroelastic "memory" effect whereby increases in pore 17

pressure due to previous lake high stands do not completely vanish by dif-18 fusion and constructively interfere with the undrained response in subsequent 19 high stands. Our preferred model suggests that the lake loading complemented 20 the interseismic stress accumulation on average by as much as 16-44%. The 21 destabilizing effects of lake inundation are enhanced by a non-vertical fault 22 dip, presence of a fault damage zone, and lateral pore pressure diffusion. Our 23 model may be applicable to other regions where hydrologic loading, either 24 natural or anthropogenic, was associated with significant seismicity. 25

26 Main

The Southern San Andreas Fault (SSAF) is the only historically quiescent segment of the San 27 Andreas Fault system (Fig. 1), believed to pose the largest seismic hazard in California¹⁻³. 28 The last major earthquake on the SSAF occurred around 1726⁴. The modern open interval of 29 \sim 300 years is well in excess of the average recurrence interval of 180±40 years over the last 30 millennium⁴⁻⁶. Previous studies suggested that at least some large events on the SSAF may have 31 occurred during high-stands of ancient Lake Cahuilla⁷⁻⁹. Lake Cahuilla was a prehistoric lake 32 that formed from periodic meandering of the Colorado River north into the Salton Trough, and 33 had an estimated maximum volume of 236 km³, ref.⁶. Filling and draining of the lake subjected 34 the SSAF system to surface loads corresponding to ~ 100 meter-deep body of water⁷⁻⁹. We use 35 new paleoseismic and geologic data, as well as numerical models, to investigate the relationship 36 between lake loading and major earthquakes that occurred on the SSAF over the last ~ 1100 37 years. 38

³⁹ Paleoseismic evidence is critically important in determining both the timing of large earth-⁴⁰ quakes and lake history. Previous work at the Coachella Paleoseismic Site, just below the ⁴¹ ancient Lake Cahuilla shoreline (Fig. 1), suggested five probable and two possible earth-

quakes over the last millenium⁵. Recently, Rockwell et al.¹⁰ reinterpreted the stratigraphy at 42 the Coachella site, building upon the results developed for the past two full lake inundations, 43 which consider both the historical record and drowned stump ages⁶. Three different lake level 44 models, varied by different sampling assumptions, were constrained by over 122 radiocarbon 45 dates across several paleoseismic sites in the ancient Lake Cahuilla inundation zone. The re-46 sulting nearly 2000 year history of ancient Lake Cahuilla is based on an extensive compilation 47 and review of radiocarbon samples throughout the lake basin, and on the unique character of 48 some stratigraphic unit interpretations. Distinct organic layers, measuring 1-2 cm in thick-49 ness and originally described as "soils", were reinterpreted to be the result of lake inundations. 50 These organic-rich layers record where organic material floated to near the shoreline and were 51 incorporated into a lake sequence during rising lake waters. Supporting this interpretation is 52 the fact that organic soils are not present in the hyper-arid environment of the Lake Cahuilla 53 basin and that such organic layers have only been identified at or below the shoreline of Lake 54 Cahuilla^{7,10,11}. Similar observations have also been made along the Laguna Salada Fault that 55 last ruptured in 1892¹². 56

The new paleoseismic data used to infer the history of lake filling and desiccation¹⁰ can 57 also be used to place tighter constraints on the timing of past earthquake events. We converted 58 the ¹⁴C radiocarbon ages to calendar years C.E. using an algorithm OxCal v. 4.4 and the his-59 tory of atmospheric ¹⁴C concentrations^{13,14}. The OxCal algorithm produces probability density 60 functions (PDFs) of each date by convolving uncertainties in the radiocarbon measurements. 61 An ordered sequence based on the stratigraphic layering provides additional constraints to the 62 radiocarbon age PDFs. The statistically consistent earthquake PDFs are calculated based on the 63 interpretation of their timings in conjunction with the full sequence. The details of our interpre-64 tation of the paleoseismic record are provided in the Supplementary Materials. Figure 2 shows 65 the calculated PDFs for major seismic events that occurred on the SSAF over the last ~ 1100 66

⁶⁷ years. Also shown in Fig. 2 is the preferred history model of Lake Cahuilla¹⁰ that includes the
⁶⁸ following six high-stands: 1731-1733 CE (Lake A), 1618-1636 CE (Lake B), 1486-1503 CE
⁶⁹ (Lake C), 1192-1241 CE (Lake D), 1007-1070 CE (Lake E), 930-966 CE (Lake F). Compared
⁷⁰ to previous estimates of the earthquake dates⁵, the new earthquake history reveals a remarkable
⁷¹ correlation between the large earthquakes and lake filling events (Fig. 2).

Such a correlation is likely more than a coincidence, as hydrologic loads are known to stim-72 ulate seismicity^{15–19}. Possible mechanisms include increases in pore pressure, flexural stresses, 73 and poroelastic effects^{20,21}. Seismicity may be directly affected by poroelastic stress perturba-74 tions due to lake loading at the Earth's surface, but often enhanced seismicity occurs on longer 75 time scales following reservoir impoundment due to the time-dependent diffusion of pore fluids 76 to greater depths^{15,19,22}. Increases in pore fluid pressure, in general, bring faults closer to failure, 77 as evidenced by reservoir-induced seismicity, and may potentially advance the timing of large 78 events^{19,23,24}. Another factor relevant to triggering of seismicity is the presence of fluid path-79 ways due to rock damage. Faults are commonly associated with damage zones^{25,26} resulting 80 in enhanced permeability that decays with distance away from the fault core^{27–31}. Fault dam-81 age zones can therefore substantially facilitate fluid transport to seismogenic depths^{32,33}. There 82 is evidence for an extensive SSAF damage zone from field observations at surface exposures, 83 borehole data, and Earth tidal analysis^{34,35}. Models of lake loading therefore need to account 84 for the fully coupled poroelastic response and the heterogeneous hydro-mechanical properties 85 of the fault zone and surrounding ambient crust. 86

The evolution of faults toward (or away from) failure is commonly described in terms of the Coulomb failure stress (ΔCFS)^{15,36,37} (see Methods). Both the magnitude and sign of the estimated ΔCFS is sensitive to several parameters in the context of poroelastic deformation, including: (i) the fault zone permeability, width, and connectivity to regions where large seismic events nucleate. Higher permeability values allow for pore pressure to diffuse faster, increasing

pore pressure and ΔCFS at depth^{38,39}. A larger fault zone width and pore connectivity may 92 also increase pore pressure and ΔCFS on the fault⁴⁰. (ii) The fault dip angle^{2,41}, the varying 93 geometry of the lake, and the concomitant surface load²⁰. The lake load produces regions of 94 relative compression and extension within the seismogenic crust (Fig. S1). Regardless of the 95 fault attitude and sense of slip, increases in the fault-normal compressive stress are expected 96 to inhibit failure, and decreases in compressive stress are expected to promote failure. (iii) The 97 undrained pore pressure effect, without which the total pore pressure will be underestimated^{24,42} 98 (Fig. S1). 99

To quantify the effects of lake loading, we built a fully-coupled, three-dimensional finite 100 element model of the Salton Trough that accounts for time-varying surface loads, a realistic 101 fault geometry, crustal poroelasticity, and viscoelastic relaxation in the ductile substrate (Fig. 102 S2) (see Methods). We performed a large number of numerical simulations in which we varied 103 material properties of the host rocks and damage zone to constrain a plausible range of pore 104 pressure and stress evolution due to ancient Lake Cahuilla (Tab. S1). The temporal evolution of 105 the lake load, which is ascribed as both a pore pressure and vertical stress boundary condition 106 at the Earth's surface, is constrained by the latest geologic data¹⁰ (Fig. 2). Each lake episode 107 consists of a unique filling, high-stand, desiccation, and consolidation history. As a preliminary 108 crosscheck, we considered a one-dimensional analytic solution for a periodic fluctuation of a 109 surface load in a poroelastic half-space¹⁵ (see "Periodic loading of a poro-elastic half-space" 110 in the Supplementary Materials). For a reasonable choice of model parameters, the analytic 111 solution reveals a cumulative increase in pore pressure at depth through multiple inundation 112 events. This "memory" effect is due to the fact that the maximum pore pressure increase at 113 depth is delayed due to diffusion and does not completely vanish before the next flooding event. 114 The timing of the superposition depends on the depth, permeability, and wavelength of the lake 115 load. One important caveat is that the memory effect only emerges when one considers multiple 116

117 lake cycles.

Across our entire model domain, the impoundment of ancient Lake Cahuilla, based on the 118 refined ~ 1100 year lake history for each different fault permeability model, invokes a coupled 119 deformation transient that we resolve with ~ 200 adaptive time steps. The increase in pore pres-120 sure within the basement is highly dependent on the assumed parameters of the fault zone, in 121 particular the damage zone permeability. For a given loading history, higher values of perme-122 ability give rise to greater pore pressures at depth (Fig. S3). Additionally, similar to the analytic 123 solution (Fig. S4; see section "Periodic loading of a poro-elastic half-space" in Supplementary 124 Material), we observe the memory effect of the pore pressure at depth, whereby subsequent 125 lakes can contribute to higher pore pressure due to the diffusive time lag of a previous lake 126 superimposing on the next (Fig. 3). The respective variations in the Coulomb stress are shown 127 in Fig. 4a. 128

The calculated values of ΔCFS resolved on a potential failure plane (Figs. S1 and S5) 129 exhibit a strong sensitivity to the fault geometry, and in particular the fault dip angle. We 130 assume an average strike of 313° based on the fault trace (Fig. 1)⁴³ and a fault dip of 60° 131 NE, which is constrained by geodetic and seismic data^{2,41,44,45}. ΔCFS values are inversely 132 proportional to the fault dip angle: a steeper fault experiences a smaller change in the Coulomb 133 stress, and a more shallowly dipping fault experiences a larger change in the Coulomb stress 134 (Fig. S5). Another factor affecting ΔCFS is a time-dependent flexure due to the varying 135 surface load and the associated viscoelastic relaxation in the underlying ductile substrate. For a 136 range of possible substrate viscosities suggested by previous work²⁰, the effects of viscoelastic 137 relaxation on ΔCFS are relatively minor (Fig. S6). Our simulations indicate that variations in 138 the pore pressure are the dominant contributor to the total ΔCFS (Figs. S7 and S8). Models 139 that used a simplified representation of only the elastic crust suggested higher values of ΔP but 140 lower values of the effective ΔCFS (Figs. S7 and S8). We also find positive ΔCFS outside 141

the lake boundary, albeit smaller in magnitude, which is a consequence of lateral pore pressure
diffusion (Fig. 2).

We find that the magnitude of shear stress ($\Delta \tau_s$), effective normal stress ($\Delta \bar{\sigma}_n$), pore pres-144 sure (ΔP) , and ΔCFS increase throughout the sequence of flooding cycles. Most importantly, 145 ΔCFS remains positive on the fault surface during each cycle (Fig. 4a). When the lake is at 146 high-stand, shear stress is positive (i.e., encouraging failure) in parts of the fault closest to the 147 lake center, and slightly negative outside of the lake. In contrast, the normal stress is negative 148 directly below the lake due to the increased vertical compression during lake impoundment. 149 However, the effective normal stress stays positive in most parts of the fault due to increases in 150 pore pressure. The modeled compression directly below the lake simply moderates the mag-151 nitude of the Coulomb stress in that region. Models with lower permeability display a similar 152 evolution, although the spatial extent of pore pressure diffusion is smaller. 153

Figure 4 depicts evolution of the maximum ΔCFS for a point on the fault plane at the 154 representative seismogenic depth (7 km) for a range of models (M1-M5, see Table S1) with 155 different permeabilities of the fault damage zone^{21,44}. In each model, as the lake fills, the ΔCFS 156 increases rapidly due to the initial lake impoundment (i.e., undrained response). During the 157 desiccation phase of the lakes, the overall ΔCFS decreases, but remains positive. The models 158 at the high end of the assumed damage zone permeability reveal a relatively rapid diffusion of 159 pore pressure from the surface to seismogenic depths. The increased rate of ΔCFS during the 160 lake high-stands is indicative of the higher permeability. Additionally, the results depict how the 161 short time interval between high-stands in the first 3 lakes contributes to an overall cumulative 162 increase in the Coulomb stress for the second (E) and third (D) lakes. The remnant diffusion of 163 pore pressure due to the first (F) lake coincides with the impoundment of the subsequent lake, 164 producing a larger overall ΔCFS . This additive effect is even more pronounced for lower fault 165 zone permeabilities. Depending on the rate of diffusion, depth of interest, and time interval 166

¹⁶⁷ between lakes, the effect of previous lakes remains during subsequent impoundments. This ¹⁶⁸ memory effect is more pronounced for Model 3 and Model 4 due to lower permeability at depth ¹⁶⁹ (Fig. S3). All models, including Model 5, which does not include a fault damage zone, produce ¹⁷⁰ positive ΔCFS values greater than ~0.1 MPa, large enough for earthquake triggering ^{36,46,47}.

One important comparison is the relative contribution of stress from tectonic loading to 171 that of lake loading. We estimate the rate of tectonic loading using a geodetically constrained 172 dislocation model assuming a secular slip rate of 18 mm/yr⁴⁸, which is in agreement with the 173 most recent constraints on the long-term geologic slip rate⁴⁹ (and ref. therein). The estimated 174 tectonic stressing rate at seismogenic depth is $\dot{\sigma}_{13} = 16$ kPa/yr (see Methods). For simplicity, 175 we assume that all of the interseismically accumulated stress is released during large seismic 176 events, and superimpose the Coulomb stress change ΔCFS calculated using our model for the 177 impoundments due to ancient Lake Cahuilla, and the tectonic loading. From the offset between 178 the stress loading rate and the ΔCFS modulation we performed a grid search across all possible 179 $\pm 1\sigma$ earthquake timings to arrive at the largest and smallest mean contributions of the lake load 180 to the total stress (Fig. 4b, see Table S2-S4). 181

Our high-end fault permeability model (M1) assumes permeability $k = 10^{-14} \text{ m}^2$ estimated 182 from the water level tidal response in boreholes³⁵. The fault permeability in M1 is high com-183 pared to that of other fault zones and on the high end of laboratory values. Thus, our preferred 184 model (M2) uses $k = 10^{-15} \text{ m}^2$ which is more consistent with regional measurements and lab-185 oratory data^{34,50,51}. According to the latter model, loading from ancient Lake Cahuilla comple-186 mented the interseismic stress accumulation by as much as 16-44% (Fig. 4b; see Table S2-S4). 187 Larger fault locking depths⁵², smaller earthquake nucleation depths⁵³, and elastic moduli of 188 the host rocks⁵⁴ would result in a lower tectonic stressing rate, and a correspondingly higher 189 contribution of the lake loading. The latter would also be amplified by higher values of the co-190 efficient of friction, and non-negligible memory effects (see Fig. 4a; also⁵⁵) from earthquakes 191

that occurred more than ~1000 years ago. Additionally, the relative contribution of hydrologic stress perturbations could be greater if one relaxes the simplifying assumption that stresses accumulated during the interseismic periods equal coseismic stress drops (Fig. 4b). Therefore, our models likely provide a conservative estimate for the modulation of stress on the SSAF by the filling and desiccation of ancient Lake Cahuilla.

Our results demonstrate how improved paleoseismic and paleolake records, together with 197 advanced models of hydrologic loading of a heterogeneous poroelastic crust, provide new in-198 sights into relationships between water level variations of ancient Lake Cahuilla and the past 6 199 major earthquakes on the SSAF. We find that increases in lake level result in positive Coulomb 200 stress changes on most of the SSAF, bringing it closer to failure. Positive ΔCFS values are also 201 seen on sections of the SSAF that are outside of the lake due to lateral diffusion of pore pressure 202 along a permeable fault zone. This indicates that our model predictions are not strongly sen-203 sitive to uncertainties in the location of nucleation sites of past (as well as future) large events 204 on the SSAF. The pore pressure "memory" effect amplifies the contribution of successive lakes 205 provided that intervals between inundations do not significantly exceed the characteristic dif-206 fusion time within the seismogenic layer. Our results also suggest an intriguing possibility: 207 that the current 300 yr-long quiescent period on the SSAF is modulated by a gradual decrease 208 in water level since the last high stand of ancient Lake Cahuilla, and its current remnant, the 209 Salton Sea. While the maximum ΔCFS resolved on the fault remains positive through time, 210 at certain fault locations our model predicts negative ΔCFS for sufficiently long dry periods, 211 including the present 300 yr-long open interval (Figs. S9-S15). The negative ΔCFS is pri-212 marily due to the flexure of the upper crust caused by a decreased lake level, and, to a smaller 213 degree, viscoelastic relaxation below the brittle-ductile transition. Further desiccation of the 214 Salton Sea in the future will continue to have a stabilizing effect on the SSAF, but increase the 215 stress to be released in a future event (or a sequence of events). Our model may be also applica-216

²¹⁷ ble to other areas where active seismogenic faults are subject to hydrologic loads from natural
²¹⁸ or anthropogenic sources^{18,19,56}.

219 Methods

²²⁰ Calculation of the Coulomb stress changes. The change in ΔCFS is defined as ^{15,36,37}:

$$\Delta CFS = \Delta \tau_s + \mu (\Delta \sigma_n + \Delta P) \tag{1}$$

where $\Delta \tau_s$ is the change in shear stress, μ is the coefficient of friction, $\Delta \sigma_n$ is the change in nor-221 mal stress (increases in compression are deemed negative) and ΔP is the change in pore fluid 222 pressure (increases in fluid pressure are deemed positive). The sign convention indicates that a 223 positive change in the effective stress $\bar{\sigma_n} = \sigma_n + P$, as well as a positive change in shear stress 224 $\Delta \tau_s$ promote failure. In this paper, we assume a typical value for static friction of μ =0.6^{57–60}, 225 although it may vary considerably depending on a rock type and ambient conditions^{61–63}. The 226 Coulomb stress changes ΔCFS are computed by resolving the normal and absolute shear stress 227 components using a full stress tensor, the prescribed fault geometry, and sense of slip. Calcula-228 tions were performed using a Matlab toolbox Abaqus2Matlab⁶⁴. 229

230

Fully-coupled 3-D finite element models. The governing equations for a fully coupled linear poroelastic three-dimensional medium are defined as^{65–67}:

$$G\nabla^2 u_i + \frac{G}{1 - 2v} \frac{\partial^2 u_k}{\partial x_i \partial x_k} = \alpha \frac{\partial p}{\partial x_i} - F_i$$
⁽²⁾

$$\alpha \frac{\partial \epsilon_{kk}}{\partial t} + S_{\epsilon} \frac{\partial p}{\partial t} = \frac{k}{\mu} \nabla^2 p + Q \tag{3}$$

²³³ Where G is the shear modulus, u the displacement, v the Poisson's ratio, α the Biot-Willis ²³⁴ coefficient, F the body force, k the permeability, μ the fluid viscosity, S_{ϵ} the constrained spe-²³⁵ cific storage, ϵ_{kk} the dilatancy (trace of the strain tensor), p the pore pressure, and Q the fluid ²³⁶ source⁶⁷. Equations (4)-(5) are nearly identical to the classic equations for linear elasticity and ²³⁷ diffusion of pore pressure, except for the coupling of pore pressure in the conservation of mo-²³⁸ mentum equation (4) and the fluid flow coupled to strain by the requirement of fluid continuity (5). Analytic solutions to the system (4)-(5) are restricted to a few highly idealized cases. We
solve the respective equations numerically using the three-dimensional finite element software
Abaqus-Simulia⁶⁸.

To account for spatial heterogeneity and anisotropy of the hydraulic properties of the lake 242 sediments, upper crust and the fault zone, we construct a three-dimensional hydro-mechanical 243 model based on parameters constrained by previous studies. Tompson et al.⁶⁹ developed and 244 arranged the hydrostratigraphy of the Salton Trough sediments into three broad classes in order 245 of increasing depth: Brawley, Palm Springs and Imperial Formations. The composition of these 246 formations vary from sandstones to shales, shaley sandstones, conglomeratic and arkosic sand-247 stones (see Table S1). The Salton Trough sediments directly overlie a basement formation and 248 upper mantle layer. The elastic parameters of the basement formation are based on inference 249 from seismic wave velocities, and hydrological parameters are based on crustal depth dependen-250 cies which produce material properties close to those of Westerly granite (Table S1)^{67,70,71}. A 251 crustal thickness of 18 km is assumed based on seismic observations of the region⁷². We mod-252 eled the upper mantle both as an elastic (Sup. S6,S7) and a viscoelastic layer. The viscoelastic 253 mantle is modeled similarly to previous studies as a simple linear Maxwell material^{20,73–75}. We 254 use a non-linear geometry option in Abaqus (NLGEOM) based on the results of Hampel et al. 255 (2019), who recommended it for models involving isostatic effects 76 . 256

The fault damage zone is modeled in accordance with field hydrogeologic estimates of SAF permeability³⁵. Xue et al. (2016) estimated a relatively high permeability (10^{-14} m^2) using earth tidal analysis compared with inferred permeability in the region and laboratory values^{34,50,51}. We use this estimate of permeability as a high-end value and explore a range of fault damage zone permeabilities between the high-end value and permeability of the ambient crust (i.e., no contrast in permeability between the fault zone material and the host rocks; Table S1). Observations also show that the effective shear modulus of rocks within a fault zone may be lower than that of the host rock by as much as a factor of $2^{25,26,52,77}$. In our model the shear modulus of rocks in the fault zone is taken to be a factor of 2 smaller than the shear modulus of the host rocks (Table S1).

The numerical domain was developed and discretized in Abaqus/CAE⁶⁸. The numerical 267 domain has horizontal dimensions of 600 km x 600 km and a depth of 50 km, with the y-268 axis corresponding to North for the Universal Transverse Mercator (UTM) 11 North zone (Fig 269 S2). The finite element mesh consists of nearly 2 million first-order linear tetrahedral elements. 270 Characteristic element sizes vary from 30,000 m in the far-field to less than 100 m on the fault 271 and lake edges. The SSAF fault trace is based on the USGS Quaternary fault database⁴³ We 272 model a 60° northeast dip for the SSAF, which is consistent with geodetic slip models and 273 tomographic observations^{45,48}. We consider a damage zone extent of 100 m on both sides of the 274 SSAF fault core, for the total width of 200 m^{26,34,77}. The lake shoreline is based on coordinates 275 used by Luttrell et al. (2007) to maintain consistency in model comparisons (D. Sandwell, 276 personal communication)²⁰. The three-dimensional body of the lake is simplified to resemble 277 its average water head. Therefore, the lake shoreline is embedded into the model consistent to 278 its relative elevation change $(97.2 \text{ m})^6$. 279

The evolution of the poroelastic effects require careful consideration to model boundary 280 and initial conditions. We assume initial conditions are at equilibrium stress and pore pres-281 sure⁷⁸ (chapter 9). These conditions imply that the model only considers the effect of the lake 282 and does not include any loading from tectonic stress (the latter is considered in Methods S4). 283 It should be noted that gravity is included in the viscoelastic mantle models with stress and pore 284 pressure calculated as relative change from geostatic equilibrium. This is an important consid-285 eration as neglecting gravity in the viscoelastic model may overestimate surface displacements 286 on time scales that exceed the characteristic relaxation time^{79,80}. The bottom and sides of the 287 model are fixed in the surface normal direction, and free of shear stress (a "roller" condition). 288

The bottom of the poroelastic domain is considered to be insulated (zero flux boundary condition), while the sides and top of the model, are considered to be permeable, and subject to vanishing pore pressure (p = 0). The top surface of the model is stress-free outside of the lake area. The lake extent is the only boundary condition that varies through time.

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²⁹⁴ **Modulation of stress during earthquake cycle**. We approximate the secular tectonic load-²⁹⁵ ing using a model of a screw dislocation in an elastic half-space⁸¹ (see equation 2.25 in⁷⁸). The ²⁹⁶ respective interseismic shear strain rate is given by:

$$\dot{\epsilon_{13}} = \frac{-\dot{s}}{4\pi} \left[\frac{x_2 - d_2}{(x_2 - d_2)^2 + x_1^2} - \frac{x_2 + d_2}{(x_2 + d_2)^2 + x_1^2} \right]$$

where \dot{s} is the fault slip rate (taken to be 18 mm/yr,⁴⁸), x_1 is the horizontal coordinate with respect to the center of the fault damage zone, x_2 is the assumed depth of nucleation of large events, and d=10 km is the locking depth of the SSAF^{48,82}.

For typical values of the shear modulus G=28-32 GPa⁴⁸, the stressing rate predicted by 300 equation (6) at the bottom of the seismogenic zone (x₂=6-8 km) is $\dot{\sigma}_{13} = 2G(\dot{\epsilon}_{13}) = 25 - 51$ 301 kPa/yr. However, we note that the distribution of stressing rates throughout the seismogenic 302 layer depends on the details of a transition between the locked and creeping parts of the fault 303 (e.g.,⁵³), as well as the degree of strain localization below the brittle-ductile transition (e.g.,⁸³). 304 A dislocation model likely over-predicts the stressing rate, especially at the bottom of the seis-305 mogenic zone due to a strong stress singularity. Also, the depth distribution of earthquakes in 306 California shows that most earthquakes nucleate at depth of 3-4 km⁵³. Therefore, we use the 307 stressing rate at the free surface ($x_2 = 0$), $\dot{\sigma}_{13} \sim 16$ kPa/yr as a conservative lower bound, and 308 twice this rate, 32 kPa/yr, as an upper bound for the entire seismogenic layer. Given the PDF dis-309 tributions of past events on the SSAF (Fig 2), we sample these PDF distributions 10,000 times 310 using inverse transform sampling to arrive at a mean earthquake age for each of the six events 311

and standard deviation. This allows us to quantify interseismic intervals, and, by multiplying
the latter by the mean stressing rate, the accumulated stress since the last earthquake.

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521 Author Contributions

R.H. constructed the finite element model, performed analysis of the model results, and wrote the manuscript. M.W. managed the study, assisted with building the model, provided access to the modeling software, acquired funding, helped conceive the experiment, and commented on the manuscript. T.R. carried out the paleoseismic analysis, conceived the experiment, and contributed to the manuscript. Y.F. contributed to interpretations of the model results and contributed to the manuscript.

528 Ethics Declaration

529 Among the authors there are no relevant financial or non-financial competing interests to report.

Data Availability

The datasets generated during and/or analysed during the current study will be made available at Zenodo.

533 Code Availability

All relevant Matlab codes and Abaqus model files for reproducing analysis results will be made
 available at Zenodo.



Figure 1 | **Regional Context.** Map of the Salton Trough and present-day Salton Sea (9.5 m mean head) with the historical extent of ancient Lake Cahuilla (97.2 m mean head). The 13 m above sea level shoreline are the same coordinates used by²⁰. Fault traces are from Quaternary fault and fold database⁴³. Green 'X' symbol marks the location of the Coachella Paleoseismic site⁵.



Figure 2 | **Lake and Earthquake History.** Earthquake PDFs superimposed on the relative lake loading history used in this study. Colored PDFs are the reinterpreted and refined earthquake distributions used in this study¹⁰(see "San Andreas Fault, Coachella Site Earthquake History" in Supplementary Material). Coa-1 is blue, Coa-2 is green, Coa-3&4 is pink, Coa-5 is magenta, Coa-6 is yellow, Coa-7 is red, and Coa-8+ is orange. Gray distributions are the previous earthquake distributions reproduced from⁵ including the lower confident Coa-6 and Coa-3.



Figure 3 | Finite Element Model Results. Left: Spatiotemporal evolution of pore pressure for a single lake cycle of ancient Lake Cahuilla. The cross section cuts through the northwestern end of the lake and through the embedded dipping fault. A) Pore pressure immediately increases as the lake begins to inundate. B) At the time of lake highstand the pore pressure has diffused down much of the higher permeable fault but is more slowly varying elsewhere due to the low permeability basement. C) As the lake desiccates higher pore pressures remain in and around the fault. D) Subsequent lakes further increase the previously elevated pore pressure at basement depth. *Right*: The Coulomb stress change (ΔCFS), pore pressure (P), effective stress ($\bar{\sigma}_n$), and absolute shear stress (τ_s) perturbations on the SSAF fault plane for a single time frame in the lake cycle. The magenta line represents the outline of ancient Lake Cahuilla at the surface.



Figure 4 | **Stress Effects of Lake Loading.** *A*) Maximum $\triangle CFS$ (mPa) on the SSAF as a function of time C.E. (years) for 7 km depth. Color lines correspond to models assuming different permeability of the fault zone (see Table S1), from highest (Model 1) to lowest (Model 5, no permeability contrast with the host rocks). *B*) Variations suggested by our preferred model (model 2), assuming an average tectonic stressing rate of 16 kPa/yr (dashed line; also see Methods). Dark and light blue solid lines denote the maximum and minimum stress perturbations, respectively, based on a grid search across all possible earthquake timings ($\pm 1\sigma$). Circles represent the minimum and maximum modulation in the grid search results for a range of σ . Stress perturbations are with respect to the (unknown) background stress at the beginning of simulation. The averaged stress load contribution plotted here is 32.33% for the lowest mean total and 43.70% for the largest mean total. For additional results, see Table S3-5.

¹ Supplementary Materials

Major Earthquakes on the Southern San Andreas Fault
 Modulated by Lake Filling Events

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⁸ Online data

9 Quaternary Fault Database is available at

10 https://urldefense.com/v3/__https://www.usgs.gov/natural-hazards/earthquake-hazards

11 faults/__;!!Mih3wA!Hr4XqsVg5wISODuGUvfFRLdy-6cECmelfBdNDod9PYtFYKnZ9vFz3RhXYvPuMHYK

- ¹² s2V\$, last accessed July 2020.
- ¹³ Matlab utility used for post processing Abaqus model in Matlab
- 14 https://urldefense.com/v3/__https://abaqus2matlab.wixsite.com/abaqus2matlab_

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15 _;!!Mih3wA!Hr4XqsVg5wISODuGUvfFRLdy-6cECmelfBdNDod9PYtFYKnZ9vFz3RhXYvPuMHYKd4d9ie8U
16 last accessed July 2020.
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$_{18}$ Methods

¹⁹ San Andreas Fault, Coachella Site Earthquake History

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The Coachella paleoseismic site (33.72722°N, 116.16976°W) trenches provide evidence 21 of 6 or 7 lake phases and 5 to 7 earthquakes on the southern San Andreas fault¹. Four 22 types of stratigraphy have been identified at the Coachella site: lake clay/silt, reces-23 sional sand, aeolian sand, and fluvial alluvium. The distinction between aeolian sand 24 and recessional sand may be difficult, as the aeolian sand can be derived entirely from 25 the recessional and shoreline sand deposits. Please refer to Philibosian et al. (2011) 26 Figure 5 for a generalized visual of the stratigraphic section and brief descriptions of 27 each unit². 28

An "organic-rich layer" is described near the top of unit 1L which is strikingly 29 similar to other organic mats from near shoreline sites that are interpreted as the 30 debris that accumulates near the shoreline from the inundation of the lake basin. Unit 31 1L is described as a massive silt layer of lacustrine origin. An alternative interpretation 32 is that unit 1L represents deltaic sedimentation during flooding of the site. In either 33 case, 1L and 1S collectively represent regional Lake A at the Coachella site¹ [see their 34 Supplementary Material]. The most recent event occurred during the highstand of 35 Lake Cahuilla in 1726 ± 7 C.E.^{3,4}. 36

The penultimate lake earthquake, Coa-2, clearly occurs during deposition of unit 2L, which represents the lacustrine phase of Lake B (Philibosian et al., 2011; Rockwell et al., 2022), based on extensive liquefaction (only possible with water present) and upward fault terminations^{1,2}. That places Coa-2 to have occurred during Lake B.

Coa-3 was interpreted to have occurred during Lake C based on weak evidence 41 of sediment (unit 3L) filling a structural trough. Coa-4 was interpreted as occurring 42 before Lake C but between two organic layers in unit 4S. However, the reinterpretation 43 is that these organic layers were deposited during the rising stages of Lake C along with 44 fine sand deposits (probably of deltaic origin) implying that Coa-4 occurred during the 45 early phase of Lake C and that the structural trough formed by Coa-4 was filled by 46 the same lake phase, Lake C^1 . Belle Philibosian agrees with this reinterpretation 47 (Rockwell, personal communication, 2022). This means that the evidence for Coa-3 48 and Coa-4 are the same and that there was only one rupture, which occurred during 49 the filling phase of Lake C but when the water had essentially reached the highstand 50 shoreline (as the Coachella site lies at 9 m, only 4 m below the highstand). 51

Coa-5 was interpreted by Philibosian et al. (2011) as having occurred between 52 deposition of units 5S and 4L, with the units 5S and 5L folded and capped by un-53 deformed unit $4L^2$. Rockwell et al. (2022) interpret the organic unit at the top of 54 unit 5S to represent the inundation and accumulation of organics associated with the 55 drowning phase of Lake D^1 . Coa-5 deforms the organic layer along with the under-56 lying lake deposits, implying that the earthquake occurred after its deposition during 57 the rising stages of Lake D. Supporting this interpretation, the strata of units 5L and 58 5S are plastically deformed, which in this arid environment, almost certainly required 59 the presence of water which will only be present during a lake stand. Hence, Coa-5 60 is interpreted to have occurred during Lake D, and probably during the late filling 61 stage as the event is capped by fine-grained deposition associated with the highstand 62 of Lake D. 63

⁶⁴ Coa-6 is interpreted to have occurred during deposition of unit 5L (Lake E) based on

⁶⁵ apparent upward terminations and a possible colluvial wedge of material derived from
⁶⁶ unit 5L. This evidence was considered weak, but the presented images look convincing,
⁶⁷ so we confirm the occurrence of a rupture during the highstand of Lake E.

Event Coa-7 is interpreted to have occurred between units 6S and 5L based on upward terminations and displaced strata capped by undeformed strata. This would potentially place this earthquake to between lakes E and F as there is no evidence of the presence of water at that time (no documented liquefaction).

A seventh event, Coa8+, is interpreted to have occurred in unit 7S, which Phili-72 bosian et al. (2011) describe as "coarse sand, gravel, and rounded cobbles form cross-73 cutting channel deposits" which are obviously fluvial in $origin^2$. Unit 7L is bedded 74 fine sand and is only exposed at the very base of the section. It is interpreted by 75 Philibosian et al. (2011) as "likely lacustrine" but a lake affiliation is not clearly 76 demonstrated. An alternative interpretation is that unit 7L is a deltaic section that 77 was buried by a large flood event during rising lake level. The rationale for this is that 78 unit 7S is several meters thick and a date a charcoal sample from unit 7S returned an 79 age of 1185 ± 30 rcyBP (sample Sb4m14A-c), which is consistent with Lake F or the 80 dry period immediately preceding Lake F and not the significantly older Lake G. 81

Furthermore, the unit does not display the oxidation that is characteristic for sed-82 iments of Unit G, which were exposed at the surface for nearly 1 ka. Hence, we 83 interpret the charcoal date as indicating that L7 may actually be the lower part of 84 Lake F (deltaic rising lake phase). If correct, an earthquake occurred during the early 85 depositional phase of lake F, followed by a large depositional event and then more 86 lacustrine sediments. The well-sorted sand of the base of unit 7S appears to extend 87 down through unit 7L and may represent a feeder pipe, with the lowest part of unit 88 7S possibly a liquefaction deposit. 89

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92 <u>Periodic loading of a poro-elastic half-space</u>

93

As a preliminary, we consider an analytical solution for a periodic fluctuation of water level over a poroelastic half-space⁵. For a half-space extending in the positive direction of z (vertical) axis from z = 0 to $z = \infty$ with a periodic water load of amplitude p_s at the surface given as both vertical stress and pore pressure boundary conditions $\sigma_{zz}(z=0,t) = -p(z=0,t) = p_s exp(iwt)$ leads to the following solution⁶:

$$\bar{p}_z = \gamma p_s + (1 - \gamma) p_s exp(-z\sqrt{(w/2c)}) exp(-iz\sqrt{(w/2c)})$$
(1)

Where $\gamma = \frac{B(1+v_u)}{3(1-v_u)}$ is the loading efficiency and c is the diffusivity. If we assume 99 incompressible grains and fluid ($\gamma = 1$) then the response is entirely the vertical stress 100 associated with water load and constant at all depth (undrained response) (Fig. S1). 101 The other end-member ($\gamma = 0$) assumes that the pore pressure is uncoupled from the 102 applied stress and is given by the pore pressure diffusion solution (drained response) 6,7 103 (equation 4-14) (Fig. S1). Moreover, equation (2) is linear with respect to p_s . If p(0,t)104 is known, the solution for p(z,t) may be parameterized in the frequency domain with 105 the Green's function that carries a pressure disturbance for frequency w at z = 0 to 106 z > 0: 107

$$p(z,t) = \int G(z,w)H(w)e^{-iwt}dw,$$
(2)

108 where

$$G(z,w) = \gamma + (1-\gamma)exp(-z\sqrt{w/2c})exp(-iz\sqrt{w/2c}),$$

and H(w) is the Fourier transform of the surface (z = 0) history of the pressure:

$$p(0,t) = \int H(w)e^{-iwt}dw$$

Therefore, if we know the lake loading history at the surface, p(0,t), we can simply take its Fourier transform, multiply this pore pressure distribution with the Green's function (3) in the frequency domain, and take the inverse Fourier transform of the product to arrive at the solution to the pore pressure distribution for all depths and times (Fig. S4).

Given a quasi-periodic lake loading history¹, the distribution of pore pressure re-115 veals an interesting effect. While the shape of the pore pressure distribution beyond 116 the Earth's surface is entirely dependent upon the poroelastic constants in Eq. (1), for 117 reasonable values of poroelastic constants of rocks comprising the upper ctust (Table 118 S1), the high frequency of the surface lake levels reveal a gradual increase of pore pres-119 sure magnitude. This effect would not be present in a "single lake" simulation, as the 120 amplitude of the maximum P would just decay exponentially. Instead, with multiple 121 lakes, the "memory" pore pressure effect is the result of a superposition of each inde-122 pendent Fourier component of p(0,t). For example, the diffusive (drained) component 123 of the poroelastic response of Lake F (Fig S4) lags its surface response at depth. This 124 is readily observed at greater depth (9 km) where essentially only the instantaneous 125 (undrained) response to the surface load is felt through the entire load of a single lake 126 cycle. Consequently, the lag of the maximum (drained) response of pore pressure can 127 coincide with the maximum (undrained) response of pore pressure for subsequent lakes 128 thus leading to larger pore pressures compared to a single lake response. Therefore, 129 with multiple lakes there is an inherent temporal dependence of the maximum pore 130 pressure. A point in our finite element model (model 2) below the lake at a depth 131 of 7.2 km that lies within our fault is more coincident to the one-dimensional model 132 at shallower 1 km depth (Fig. S4). This 1D solution serves to conceptualize the im-133 portant delayed memory effects of variable pore pressure. While the analytical 1-D 134 solution is a useful first-order approximation, especially for locations directly below 135 the lake, it fails to account for other potentially important factors affecting stress on 136 the fault such as crustal heterogeneities, viscoelastic relaxation, and lateral pore fluid 137 diffusion. 138

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Figure S1: Example of the undrained and drained effect from a lake load. At t = 0 the undrained effect is felt nearly instantaneous throughout the poroelastic medium beneath the lake. As time progresses this effect attempts to equillibriate at depth. At t = 0 the drained effect is not felt except for thesruface poroelastic medium and the bottom of the lake. As time progresses this effect increases pore pressure as diffusion drives fluid from the surface down. Furthermore, as the lake load is applied areas of compression form immediately beneath lake while areas of extension are formed near the edges.



Figure S2: 3D FEM model domain. The model mesh contains ~ 2 million tetrahedron elements. The light blue color represents the extent of Ancient Lake Cahuilla. The prescribed vertical load is hydrostatic, to the lake average water head (97.2 m). The solid red line is the SSAF fault trace. The fault zone is modeled as a slab dipping to the north-east at 60^{o} ¹¹, with the assumed thickness of 200 km^{12–14}.



Figure S3: Maximum Pore Pressure across the SSAF fault plane for every time step. Each model is based on the variable fault permeability with Model 1 as the highest permeable and Model 5 as no damage zone (Table S1).



Figure S4: 1D analytical model of pore pressure for a variety of different depths (blue) with surface lake level pore pressure (black). FEM Model2 at 7.2 km depth (green line) shows the effect of 3D diffusion with a high-permeability fault damage zone embedded in a lower permeability host rock. The FEM model at 7.2 km resembles pore pressure in the 1D analytical case at 1 km, demonstrating how a fault damage zone can transmit pore pressure to depth effectively. ($\gamma = 0.1685$; $k_{fault} = 1e-15$ [m²]; $k_{host/1Dmodel} = 1e-18$ [m²]).



Figure S5: ΔCFS is dependent on different Fault dip angles. Based on the location of our lake geometry a steeper dip angle yields lower overall ΔCFS while a shallower dip angle yields higher overall ΔCFS . A steeper dip accommodates more compressional stress and is therefore lower while a shallower fault accommodates more extensional stress. Plot presented is based on a preliminary lake model that contains minor differences in ages.



Figure S6: ΔCFS for the fastest relaxation time endmember (20 years), the slowest (200 years), and preffered model viscosity (70 years) (Model 2)¹⁵. Changing the viscosity of the mantle results in minor changes to the maximum ΔCFS .



Figure S7: Model 1, 7 km depth for viscoelastic mantle and elastic mantle. Pore pressure is a dominant signal of the total ΔCFS . The elastic mantle model resolves larger pore pressure values, but lower $\Delta CFS - \Delta P$ values. Plot presented is based on a preliminary lake model that contains minor differences in ages.



Figure S8: Model 4, 7 km depth for viscoelastic mantle and elastic mantle. Pore pressure is still the dominant signal of the total ΔCFS despite having a significantly less permeable damage zone. The elastic mantle model resolves larger pore pressure values, but lower $\Delta CFS - \Delta P$ values. Plot presented is based on a preliminary lake model that contains minor differences in ages.



Figure S9: Test points along the SSAF equivalent to 6 rows at (3,4,5,6,7,8 km depth) and 5 columns as shown as green squares on a reduced cloud of points of the fault. The top left square on the figure (r1,c1) represents row 1 column 1.



Figure S10: 6 row points for (3,4,5,6,7,8 km depth) associated with column 1 from S8 (Model 2).



Figure S11: 6 row points for (3,4,5,6,7,8 km depth) associated with column 2 from S8 (Model 2).



Figure S12: 6 row points for (3,4,5,6,7,8 km depth) associated with column 5 from S8 (Model 2).



Figure S13: 5 column points associated with row 1 (3 km depth) from S8 (Model 2).



Figure S14: 5 column points associated with row 3 (5 km depth) from S8 (Model 2).



Figure S15: 5 column points associated with row 5 (8 km depth) from S8 (Model 2).

Hydrostratigraphy Unit	Post Brawley / Brawley	Palm Springs / Borrego	Imperial	Crystalline Basement	$Mantle^*$	Variable Fault
Depth Range [km]	0-0.5	0.5-3	3-7	7-18	18-50	0-18
Poisson's Ratio	0.25	0.25	0.25	0.25	0.36	0.25
Young's Modulus [GPa]	50	50	50	58.4	140	1/2 host rock
Permeability [m ²]	2e-13	1.2e-12	9e-14	1e-18	/	1e-14,1e-15,1e-16,1e-17
Bulk Modulus of Solid Grains [GPa]	33	33	33	39	/	host rock
Reference	(8)	(8)	(8)	(9)	(7)	(10, 11)

Table S1: *The mantle is modeled as a simple linear viscoelastic material with 3e19 Pa·s viscosity¹⁵. Bulk modulus of permeating fluid (water) is assumed to equal 2.2e9 Pa. Therefore, models 1-4 coincide with fault permeability differences of [1e-14,1e-15,1e-16,1e-17] respectively and model 5 represents a no damage zone so permeability and Young's modulus are identical to those of the host rock.

Model		M1	M2	M3	M4	M5
16 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	36.26	32.33	25.47	18.26	16.11
	Maximum (mean)	49.21	43.70	33.49	24.11	21.99
24 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	24.17	21.55	16.98	12.17	10.74
	Maximum (mean)	32.80	29.14	22.33	16.07	14.66
32 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	18.13	16.16	12.73	9.13	8.05
	Maximum (mean)	24.60	21.85	16.75	12.05	10.99
Avg Stress Drop (4.88 MPa)	$\pm 1\sigma$					
	Minimum (mean)	8.46	7.22	5.24	4.19	3.91
	Maximum (mean)	11.48	9.95	6.97	4.85	4.44

Table S2: For each permeability varying model (M1-M5) the average modulation of lake loading on the tectonic loading cycle for a variety of secular tectonic loading [16,24,32] kPa/yr (slip-predictable model). Additionally we include the modulation for an average stress drop of 4.88 MPa equivalent to ~ 7.4 Mw for each event timing (slip-predictable model). Values are based on all earthquake events Coa-8+ through Coa-1.

Model		M1	M2	M3	M4	M5
16 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	24.86	21.67	16.28	13.05	12.28
	Maximum (mean)	49.43	43.49	32.25	23.69	22.02
24 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	16.57	14.45	10.85	8.70	8.19
	Maximum (mean)	32.95	29.00	21.50	15.79	14.68
32 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	12.43	10.83	8.14	6.52	6.14
	Maximum (mean)	24.71	21.75	16.12	11.85	11.01
Avg Stress Drop (4.88 MPa)	$\pm 1\sigma$					
	Minimum (mean)	8.00	6.87	5.19	4.36	4.10
	Maximum (mean)	11.52	10.07	7.20	5.13	4.73

Table S3: Same as Table S2 for $\pm 1\sigma$ except the calculated mean does not include the largest modulation contribution from the COA 7 event (ie. the largest modulation).

Model		M1	M2	M3	M4	M5
16 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	13.33	10.96	7.48	7.44	7.43
	Maximum (mean)	20.98	17.96	11.99	9.40	9.22
24 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	8.89	7.31	4.98	4.96	4.95
	Maximum (mean)	13.99	11.98	7.99	6.26	6.15
32 kPa/yr (Loading)	$\pm 1\sigma$					
	Minimum (mean)	6.66	5.48	3.74	3.72	3.71
	Maximum (mean)	10.49	8.98	6.00	4.70	4.61
Avg Stress Drop (4.88 MPa)	$\pm 1\sigma$					
	Minimum (mean)	7.84	6.61	4.77	4.27	4.18
	Maximum (mean)	11.28	9.80	6.88	5.12	4.87

Table S4: Same as Table S2 for $\pm 1\sigma$ except the calculated mean does not include the largest modulation contribution from the COA 7 and COA 6 events (ie. the two largest modulations).

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