works against any hypothesis that attributes deep earthquakes in southern Tibet to processes related to subduction.

Large to moderate-sized earthquakes occur at depths of 100 km or more beneath the western Himalayan syntaxis and the western Kunlun Mountains. Such focal depths are likely to be in the mantle, indicating that the uppermost mantle of the continental lithosphere is strong enough to sustain the accumulation of elastic strain required for causing earthquakes. The thickened crust of Tibet appears to vary in thickness by up to 20 km over distances of a few hundred kilometers, so whether every unusually deep earthquake is in the mantle remains uncertain. Nonetheless, a bimodal distribution of focal depths, peaking in the shallow crust and near the Moho, strongly suggests that the two seismogenic regions of the continental lithosphere correspond to maxima in mechanical strength.

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Dynamics of Slow-Moving Landslides from Permanent Scatterer Analysis

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High-resolution interferometric synthetic aperture radar (InSAR) permanent scatterer data allow us to resolve the rates and variations in the rates of slow-moving landslides. Satellite-to-ground distances (range changes) on landslides increase at rates of 5 to 7 millimeters per year, indicating average downslope sliding velocities from 27 to 38 millimeters per year. Time-series analysis shows that displacement occurs mainly during the high-precipitation season; during the 1997–1998 El Niño event, rates of range change increased to as much as 11 millimeters per year. The observed nonlinear relationship of creep and precipitation rates suggests that increased pore fluid pressures within the shallow subsurface may initiate and accelerate these features. Changes in the slope of a hill resulting from increases in the pore pressure and lithostatic stress gradients may then lead to landslides.

Resolving the kinematics of slow-moving, continuously creeping landslides may aid in understanding the mechanics of these hazardous features. The location and extent of some landslides can be determined by geologic mapping, but it has been difficult to develop spatially detailed characterizations of their rates of movement over large areas and time spans (*1*). InSAR (*2*, *3*) can resolve the movement of large (>1 km^2) landslides (*4*); however, coherence problems, inherent error sources, and the spatial and temporal resolution of this method hamper detection and monitoring of landslide features whose rates fluctuate over time. Alternatively, the permanent scatterers InSAR method (PS-InSAR) (*5*–*8*) identifies these scatterers (radar-bright and phase-stable targets such as buildings, utility poles, and rock outcrops) within many (>15) SAR scenes to determine a time series of displacements with high spatial and temporal resolution.

We used 46 scenes acquired by European Remote Sensing Satellites ERS-1 and ERS-2 between 1992 and 2001 to construct a range-change time series for the Berkeley vicinity in the eastern San Francisco Bay area (*9*). Here, the active Hayward Fault (HF) bounds the western margin of the East Bay Hills (EBH), which rise to ~370 m above sea level. The PS-InSAR analysis identified 18,428 PS, which we used in our analysis. Observed range-change rates reflect shallow aseismic right-slip movement along the HF (*8*, *10*, *11*); however, several groups of PS located along the mid-slopes of the EBH display large positive range-change rates (Fig. 1A). To determine the spatial extent and rates of movement of these features, we corrected the range-change rates for the field-measured surface HF slip rates (*12*) and an additional regional, cross-fault offset that may reflect the ~0.4-mm/year uplift of the EBH to the northeast (*13*). The interpolated, adjusted range-change rates resolve at least three patches of large range-rate increases whose locations match those of mapped landslides (*14*) (Fig. 1B). The upslope portions of the slides are commonly located in the vicinity of the HF, whereas downslope portions terminate toward the southwest margin of the EBH. Range-change rates of the southern two landslides were ~5 to 7 mm/year. These rates were measured in the direction of the look-angle of the satellite, so there was insufficient information to resolve all of the components of the displace-

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ment or velocity vectors. However, assuming that most of the displacement along these landslides follows the average 4° downhill slope, the measured range-change rates imply sliding at velocities of 27 to 38 mm/year (*15*).

The time series of range-change rates allowed us to explore the relationship between precipitation and slide movements. Cumulative precipitation and 20-day averaged precipitation rates were calculated using data from the Richmond meteorological station (*16*), about 5 miles north of the landslides (*8*). Between 1992 and 2001, annual precipitation averaged 0.61 m/year, with 95% of the precipitation falling between the months of October and April. At values of seasonal cumulative precipitation less than 0.5 m, higher precipitation intensities led to proportionally increased slide displacements, whereas larger amounts of yearly precipitation resulted in less movement than expected on the basis of dryer years (*8*, *17*).

The 1997–1998 El Niño event saw a ~200% increase in seasonal precipitation; peak 20-day-averaged precipitation intensity exceeded 6 m/year, about 10 times the long-term annual average. Seventeen scenes acquired during the El Niño event and the following year resolve the details of the movements of the landslides in response to the wet year (Fig. 2). In the season preceding the enhanced El Niño precipitation, range-change rates in the vicinity of the northern and middle slides were low and often difficult to distinguish from the surrounding areas, while the southern slide moved between January and July 1997. With the onset of the El Niño rains, we could not detect movement of the mapped landslides even ~3 months into the rainy season. Between November 1997 and April 1998, range-change rates along the middle slide accelerated to maximum values of 35 mm/year. In the following dry season (May to November 1998), movement along the slides was not detectable. During the El Niño, the peak yearly-averaged range-change rates were ~10 mm/year, about 30% higher than typical years. However, the 1997–1998 El Niño seasonal displacements were not as high as would have been predicted from extrapolation of trends observed at lower cumulative precipitation (*8*). Total precipitation during the following year (1999) was only 35% of that during El Niño, which is generally reflected in the 10 to 20 mm/year decrease in the peak range-change velocities of the middle and southern slides during the wet season. The north slide moved ~20 mm/year (greater than during the wetter El Niño time by ~5 to 10 mm/year) during 1999 (*8*, *18*).

The association of landslide motion with high precipitation indicates that near-surface groundwater flow may play a role in the initiation and acceleration of sliding. In particular, increased flow in the near-surface groundwater system and eventual saturation may increase pore pressures, decrease the effective strength of the failure surface, and trigger movement (*19*, *20*). However, the ~3-month time lag observed between the onset of precipitation and acceleration of the slides during the El Niño season suggests that the near-surface groundwater system acts to buffer the effects of intense and sustained precipitation early in the wet season. Specifically, the time associated with saturating the groundwater system appears to discourage failure along the slides early in the wet season and during relatively dry years. The nonlinear response of landslide movement to seasonal precipitation (*8*) suggests that saturation of the near-surface hydrologic system may occur during wet years, which may act in concert with increased sliding resistance at higher velocities to reduce the sensitivity of creep rates to total seasonal precipitation.

Whereas the seasonal acceleration of sliding is closely tied to rainfall, the location of the slides on the slopes of the EBH is determined by slope increases caused by long-term tectonic uplift along the HF. The head scarps of the landslides generally are located upslope of a ~150% increase in average hill-slope angles in the vicinity of the HF. This observation suggests that there is likely a causal relationship between the increases in slope (perhaps driven by uplift northeast of the HF) and the location of these landslides.

To test the hypothesis that the landslides are tied to slope increases near the HF, we focus on the topography of the middle landslide where

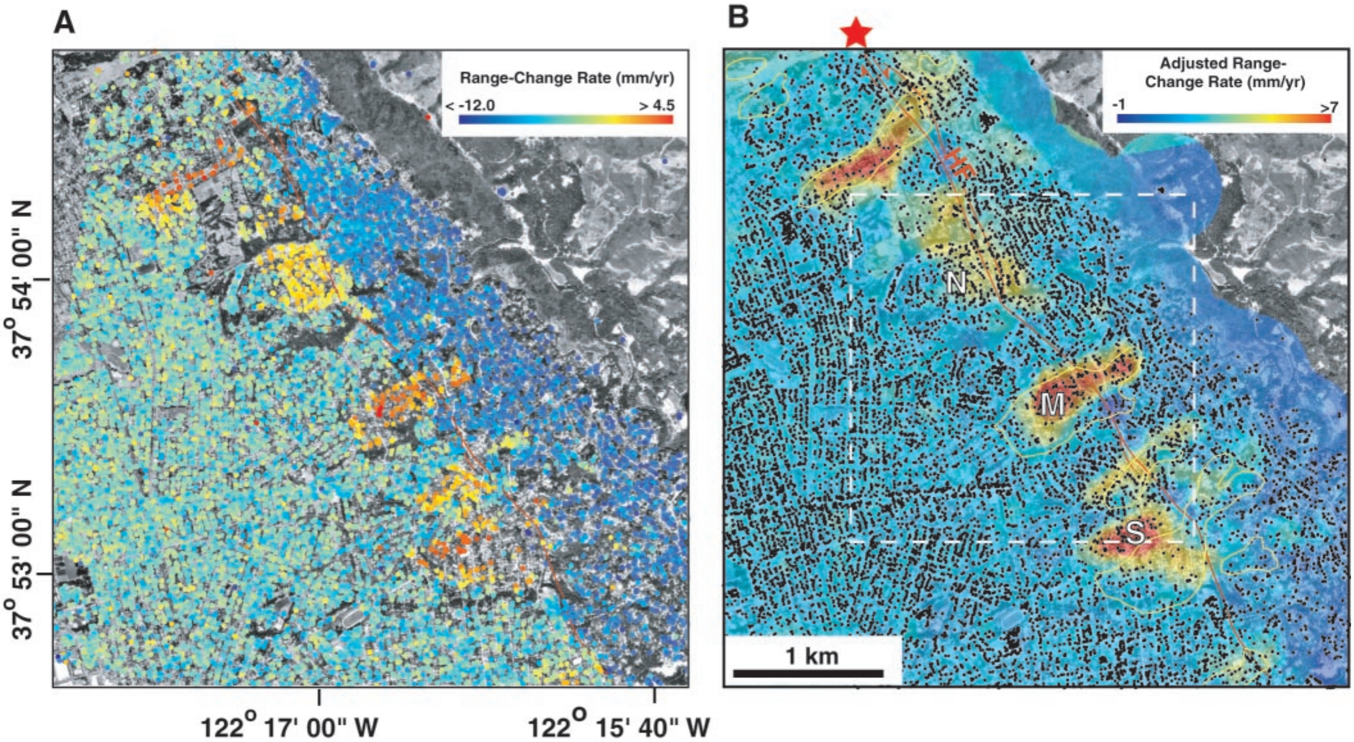

Fig. 1. (**A**) Map view of PS-InSAR range-change rate measurements for the study area. Underlying image is an orthorectified air photo of the area; HF trace is indicated by a red line (*28*). (**B**) Map view of interpolated range-change rates (colors) adjusted for shallow creep (4 to 5 mm/year) along the HF and uplift (0.4 mm/year) of the EBH (black dots show PS locations). Yellow outlines show the location of mapped active landslides (*14*). Dashed box indicates the extent of panels in Fig. 2. N, M, and S denote locations of the northern, middle, and southern landslides investigated. Red star shows location of $M_L = 4.1$ earthquake on 4 December 1998 (*8*, *18*).

this relationship is most pronounced. Landslides can be triggered by the weight of the overlying material and fluid pressure changes driven by groundwater flow. In soil or rock near the surface, where cohesion is negligible, failure ensues when the shear traction (τ) acting along a potential failure surface exceeds the normal traction (σ) scaled by the material friction coefficient (μ) (*21*). The presence of groundwater increases the pore fluid pressure (p) along this potential failure surface and reduces the normal traction that would otherwise discourage failure (*19*). To quantify these competing effects, we use the Coulomb failure function (CFF) (*21*) that gauges the propensity for failure in the near-surface layer:

$$CFF = |\tau| - \mu(\sigma + p) \qquad (1)$$

Here, compressive tractions (defined as negative) are generally opposed by positive pore pressures. When $CFF \geq 0$, the near-surface material will fail as a landslide.

We simulated the approximate surface geometry of the middle slide with a model (*8*) that couples lithostatic stresses (*22*, *23*) with pore pressures generated by steady groundwater flow (*24*–*26*). From these models, we calculated the stress tensor for every point near the surface. The shear and normal tractions in Eq. 1 vary with the orientation of the potential failure plane, and so we let this orientation vary at each point so as to maximize the value of CFF (CFF_{max}) (*21*) (Fig. 3). Values of CFF_{max} close to zero should outline the approximate extent and geometry of new landslide failure surfaces. The combined stresses due to gravity and pore pressures move the near-surface region along the entire hill slope close to or beyond their failure limit. However, the change in slope in the vicinity of the HF causes a deepening of the failure zone downslope of the head scarp of the slides. Finally, the failure envelope shallows both toward the top and bottom of the topography. Therefore, the model results indicate that failure is promoted within the mid-slopes of the topography and may be exacerbated by slope changes in these areas. These predictions agree with the observed location of the landslides (*27*), indicating that the lithostatic stresses and groundwater flow are important components of the failure and movement of these slides.

Our study demonstrates a method that can resolve detailed seasonal variations in the movement of slow-moving landslides. Our analysis of these features shows that their location may be influenced by modification of near-surface groundwater flow by tectonically induced slope changes. This flow is, in turn, modulated by seasonal increases in precipitation. The progressive saturation of this near-surface groundwater system during the wet season may induce a lag between the beginning of intense rains and acceleration of the landslides. The relationship between seasonal precipitation and slide movement is nonlinear, in that rainfall above certain amounts may not cause additional pore-pressure increases and slip acceleration of these types of landslides.

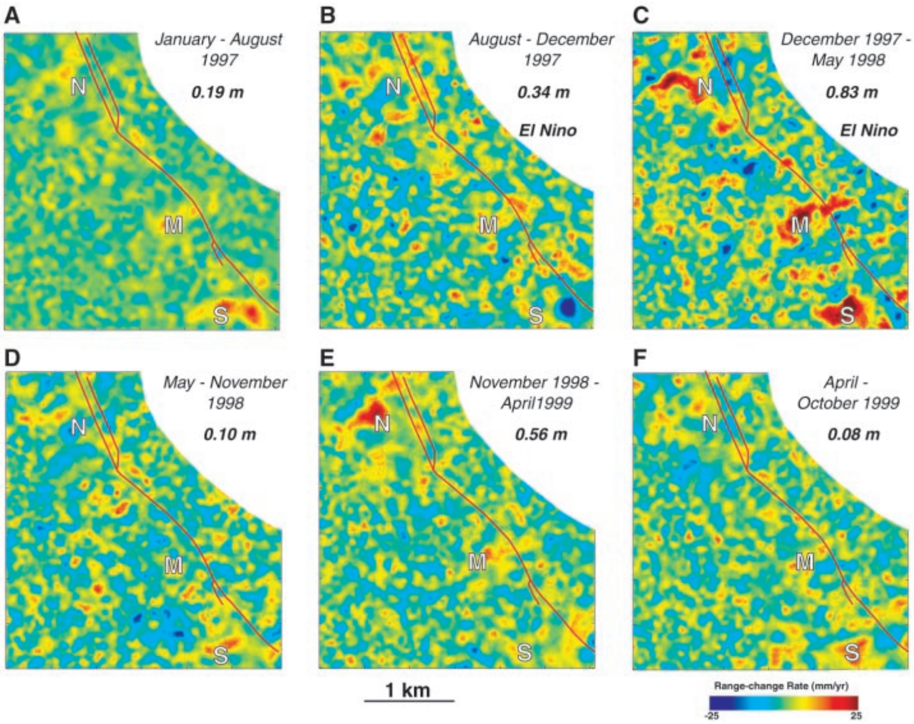

Fig. 2. Time-dependent range-change rates from 1997 to 2000 (panels span time indicated in upper right corner), including the 1997–1998 El Niño period. See Fig. 1 for location of panel. Bold italic numbers denote the amount of cumulative precipitation experienced by the area during the elapsed time shown. The HF is shown as a thin red line in each panel.

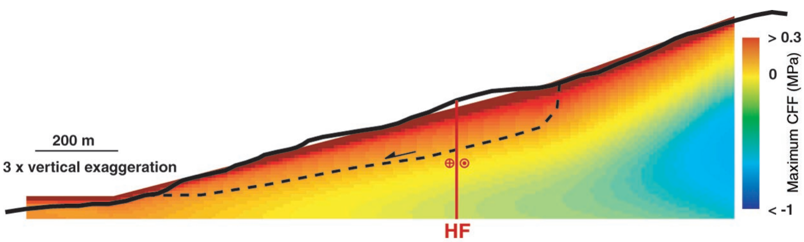

Fig. 3. Topographic profile of the middle slide (heavy black line), inferred location of slide base (heavy dashed line), and modeled value of CFF_{max} (color scale bar on right) (*8*, *21*). Here, slopes along the lower and upper portions of the hills increase from 4.2° to 6.3°. Orientation of optimal failure plane in the near-surface region varies slightly with downslope position but is close to ~15° downslope. Positive values of *CFF* indicate regions expected to undergo brittle failure; negative values of this parameter indicate mechanically stable regions.

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15. Sliding velocities were computed by projecting the unit vector of the downslope sliding direction (average slope angle 4°; downslope direction vector v = [−0.66 −0.75 −0.07], where vector is defined as [x y z], positive-up, right-handed coordinate system) onto the unit vector defining the look direction of the satellite (azimuth = 13.8°; off-vertical axis = 23.0°; satellite look direction vector u = [−0.23 0.06 −0.97]) using the equation $S = 1/(v^T u)$, where S is the amplification of the range-change rate when projected into the downslope direction, and v^T is the trans-

pose of ν. This computation yields a downslope projected velocity of 5.5 times the measured range-change rate.
16. Data obtained from the NOAA National Climatic Data Center (http://cdo.ncdc.noaa.gov/CDO/cdo).
17. Time intervals between each pair of SAR satellite scenes varied from year to year. To avoid bias in seasonal cumulative displacements, we manually inspected the range-change time series and found that displacement of the slides occurred almost exclusively during the wet season, so displacements from the wet season capture the total yearly change.
18. The north slide was closest to a M_L (catalog local magnitude) = 4.1 earthquake on 4 December 1998. Although the temporal resolution of the time series is insufficient to clearly resolve triggered sliding caused by this event, the seasonal displacement along the north slide during this period was anomalously high given the amount of precipitation during this interval (*8*), possibly indicating that the sliding may have been accelerated by seismic shaking.
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Supporting Online Material
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Materials and Methods
Fig. S1
References

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Abrupt Tropical Vegetation Response to Rapid Climate Changes

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Identifying leads and lags between high- and low-latitude abrupt climate shifts is needed to understand where and how such events were triggered. Vascular plant biomarkers preserved in Cariaco basin sediments reveal rapid vegetation changes in northern South America during the last deglaciation, 15,000 to 10,000 years ago. Comparing the biomarker records to climate proxies from the same sediment core provides a precise measure of the relative timing of changes in different regions. Abrupt deglacial climate shifts in tropical and high-latitude North Atlantic regions were synchronous, whereas changes in tropical vegetation consistently lagged climate shifts by several decades.

In order to evaluate the relative roles of high and low latitudes in initiating and propagating abrupt global climate changes, we need precise information regarding the relative timing of abrupt changes in different regions. Dating uncertainties, however, are typically too large to constrain the timing of the briefest decadal events in records from different sites (*1–4*). Another approach is to identify high- and low-latitude climate proxies in the same high-resolution record, and determine the relative timing of changes stratigraphically (*3–6*). For instance, increased methane concentrations attributed to the expansion of tropical wetlands (*7*) have been measured in air trapped in Greenland ice and used to infer shifts to warmer and/or wetter tropical climate during the abrupt Glacial/Bølling and Younger Dryas/Preboreal transitions (*3*, *4*). Temperature changes over Greenland were also reconstructed from the same samples using nitrogen and argon isotopes, allowing the precise identification of relative timing for rapid changes between the tropics and high latitudes. The tropics were found to lag Greenland by 20 to 30 and 0 to 30 years for the Bølling and Preboreal warmings, respectively (*3*, *4*), favoring a North Atlantic trigger at least for the Bølling event (*4*). In similar studies, radiocarbon measured in planktonic foraminifera from tropical Cariaco basin sediments was shown to have increased steeply during the onset of Younger Dryas cooling (*5*, *6*). Atmospheric concentration of cosmogenic beryllium-10 does not show a similar increase during the Younger Dryas onset (*8*), suggesting that the ^{14}C increase was not caused by changes in production rate, but that it instead reflects an abrupt decrease in high-latitude North Atlantic Deep Water (NADW) formation and export (*5*, *6*, *9*). In addition, relative reflectance (gray scale) and laminae thickness from the same sediments reveal rapid shifts in Cariaco upwelling and trade-wind intensity caused by shifts in the mean position of the Intertropical Convergence Zone (*2*, *5*, *6*). Direct comparison of Cariaco radiocarbon and gray-scale data showed that high- and low-latitude climate shifts during the onset of Younger Dryas cooling were synchronous within 10 years (*6*), and thus allows either a North Altantic or a tropical trigger for this rapid climate cooling. If abrupt deglacial warming and cooling events were manifestations of the same millennial-scale shifts in global climate, this subtle discrepancy in timing between Cariaco and Greenland data must be resolved before we can understand the mechanisms responsible for abrupt climate change.

The delayed increase in atmospheric methane following abrupt warmings may have been caused by the release of gas hydrates rather than expansion of tropical wetland vegetation (*10*). However, studies have shown fluctuations in deglacial tropical moisture balance similar in timing to the Bølling/Allerød and Younger Dryas oscillations in the North Atlantic region (*11–17*), supporting tropical wetlands as the source for the atmospheric methane signal. Specifically, detailed pollen records from Central and northern South America (*11*, *12*), including the Cariaco basin watershed (*13*, *14*), show that vegetation shifted between predominantly dry grasslands during the Glacial and Younger Dryas periods, and wet montane forest during the Bølling/Allerød and Preboreal periods. The lag of the methane increase behind Greenland warming may have occurred because the time scale necessary for tropical wetland expansion and development of anoxia following a climate shift may be longer than previously thought (*4*, *18*). However, the response time for changes in vegetation fol-

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