

## Long-term creep-rate changes and their causes

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[1] We discuss measurements of the long-term creep rate from the San Juan Bautista section of the San Andreas fault. These observations of aseismic slip span a period of approximately half a century making it the longest creep record ever measured along an active fault zone. The records show systematic changes in creep rate over time scales of tens of years, and we address potential causes of these changes. In principle, they may be caused either by changes in resistive forces within the fault zone or by variations in plate-tectonic driving forces. Such driving force variations have recently been suggested to have occurred in the same area (Western North America) and within the same time window. These should give rise to observable effects in the creep data, but we find no evidence for temporal changes in plate-tectonic forcing at these time scales. A natural explanation for the observed long-term changes is elastic reloading of adjacent locked portions of the San Andreas fault after large earthquakes and possibly aseismic strain release that occurs before the earthquakes. *INDEX TERMS:* 1206 Geodesy and Gravity: Crustal movements—interplate (8155); 1208 Geodesy and Gravity: Crustal movements—intraplate (8110); 1213 Geodesy and Gravity: Earth's interior—dynamics (8115, 8120); 1236 Geodesy and Gravity: Rheology of the lithosphere and mantle (8160); 7230 Seismology: Seismicity and seismotectonics. **Citation:** Bokelmann, G. H. R., and R. L. Kovach, Long-term creep-rate changes and their causes, *Geophys. Res. Lett.*, 30(8), 1445, doi:10.1029/2003GL017012, 2003.

### 1. Introduction

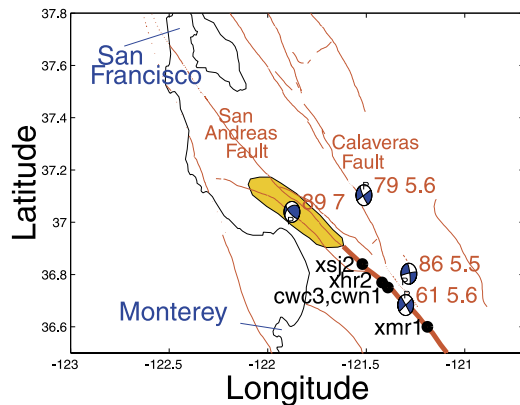
[2] Most active faults are thought to creep more or less freely at depth [Scholz, 2002], where it can usually not be observed directly. In a few places however, fault creep extends to the surface, and we can measure the creep there very easily using length changes of a rod, wire spanned across a fault or using optical techniques [Agnew, 1986]. The best-instrumented region among those is the creeping section of the San Andreas fault that stretches between San Juan Bautista (SJB) in the north to Parkfield in the south, a distance of about 140 km. In the central portion of the creeping section the creep rate is about 35 mm/yr, which is close to the relative plate velocity [Ben-Zion *et al.*, 1993]. Thus it appears that there is little strain built up there. The creep rate tapers off to zero toward the northern and southern ends of the creeping section. Near those transitions to the adjacent locked sections a number of creepmeters have been operating for considerable time. Several creep records in the San Juan Bautista are about 30 years long and one station has a remarkable record of about 50 years. This paper addresses

the characteristics of long-term creep and especially the causes of variations in the long-term creep rate that occur over time scales of years to tens of years. Changes over somewhat shorter time scales have been discussed previously by *Mavko et al.* [1985] and *Burford* [1988] who observed variations of fault creep rates associated with moderate-size earthquakes in Central California. These changes occurred during and after the earthquakes, but also in the months before earthquakes. We are studying the longer-term creep rate, over time-scales of years to tens of years. That long-term creep rate has to keep up with the plate tectonic rate [Mavko, 1982], and we assume that the long-term creep rate represents the rate at which the deeper creep occurs. This allows us to address the causes of long-term variations in the fault creep rate, especially whether they are due to elastic reloading of locked sections of the fault after earthquakes, or to changes in plate-tectonic driving forces.

### 2. Creep Data

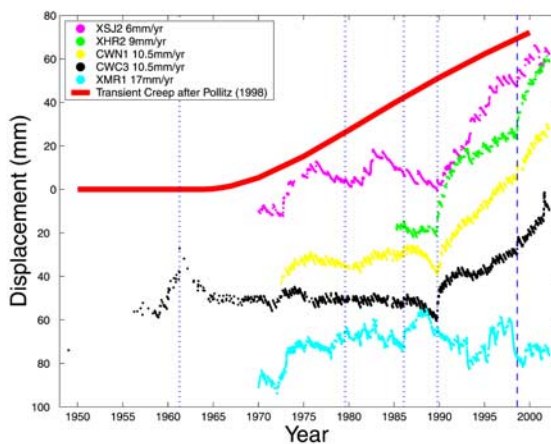
[3] Figure 1 shows the northern portion of the creeping section of the San Andreas fault, and the locations of five creepmeters that extend along about 30 km of the fault. Two of them (CWC3 and CWN1) are nearly co-located, separated by only 30 meters within a winery building. Figure 1 also shows the locations of a number of earthquakes that occurred on the San Andreas fault between 1960 and 2003. The largest of these events was the 1989 Loma Prieta earthquake immediate to the northwest of the creeping section. Three earthquakes with magnitude at least 5.5 occurred just a few kilometers from the creepmeters, two of which being collocated and occurring on the same day (April 29, 1961; magnitudes 5.6 and 5.5; joint magnitude 5.7).

[4] The surface creep recorded at the five stations is shown in Figure 2, reduced by different background creep rates. Note the clear changes in the long-term creep rate, e.g. the longest creep record, for CWC3, varies between 10 and 20 mm/yr. In detail, the creep records are rather complicated. Much of the creep occurs in the form of “creep events” which are typically 0.1 to 10 mm. The occurrence of these events can be somewhat influenced by environmental factors such as rainfall [Schulz *et al.*, 1982]. Figure 2 also shows the times of the four largest earthquakes along the San Andreas fault within the time interval of recorded creep. The earliest data point for CWC3 (in 1949) and subsequent values were extracted from the construction of the building and scratch marks [Nason, 1969, 1973; Nason and Tocher, 1971]. It appears that there are major coseismic changes in creep rate during the 1961 and 1989 earthquakes. Beside the large creep events during both earthquakes the continuous creep rate seems to deviate from the background creep rate up to 2 years before the earthquakes, and that creep rate anomaly seems to persist for at least a



**Figure 1.** Map of the San Juan Bautista area showing the locations of 5 creep meters (black symbols) along the Creeping Section of the San Andreas fault (thick red line). Also shown (thin red lines) are the known fault zones and major ( $M \geq 5.5$ ) earthquakes on/near the San Andreas fault since 1960 (given as year, magnitude). The yellow area indicates the approximate extent of the source volume of the Loma Prieta earthquake [after Goyer *et al.*, 1994].

few years. Before the 1989 quake this period of slower creep may perhaps have started at the time of the 1986 earthquake, but for the 1961 earthquake there is no significant quake that can be associated with the beginning of that interval. Interestingly, the creep excess before the 1961 earthquake is roughly matched by the creep deficiency after the event, and the fault offset for this moderate-size earthquake is about that expected from the background rate. This



**Figure 2.** Creep rates on the five creepmeters sorted by latitude (reduced with creep rates given in legend). Dotted lines show the times of major earthquakes in the area (from Figure 1). The broken line shows the time of a strain-transient (and magnitude 5.1 earthquake) Note the clear change in long-term creep rate with the Loma Prieta earthquake in 1989 on all stations except the southernmost one (XMR1). The red curve is the prediction for a transient change in plate-tectonic driving [after Pollitz *et al.*, 1998, see text]. The transient Pacific plate motion is predicted to be up to 2.6 mm/yr faster than usual within our measurement interval, but that is not observed in the creep records.

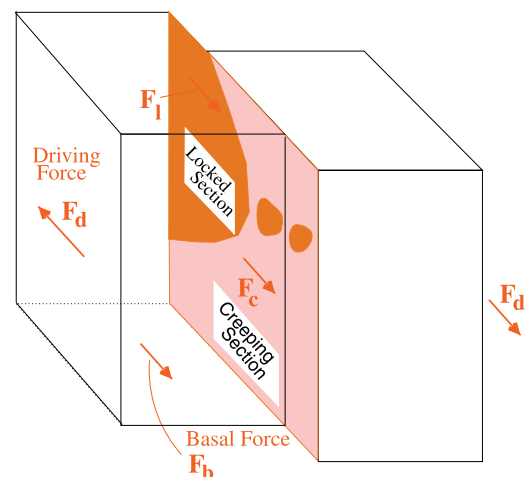
is different for the much larger 1989 earthquake which clearly produced an additional net fault offset of at least 4 cm due to the earthquake. Apart from the earthquakes on the San Andreas fault, three significant earthquakes took place on the Calaveras fault. They occurred in 1979, 1984 and 1986 (magnitudes 5.6, 6.2, and 5.7) at distances of 40, 63, and 82 km from CWC3, but there are no observable changes in creep rate associated with those events observed on the San Andreas fault.

### 3. What Causes the Long-Term Creep-Rate Changes?

[5] In principle the creep-rate changes can be caused by changes of resistive forces acting in the San Andreas fault system or by changes in the plate-tectonic driving forces (Figure 3). The different forces are constrained by the requirement of mechanical equilibrium. At all times except during large earthquakes the remotely applied driving force  $F_d$  must be balanced by the resistive forces acting on the faults,  $F_f$ , and at the base of the (upper) crust  $F_b$  [Lachenbruch and Sass, 1973]. Resistive forces  $F_f$  on the fault arise from the contributions of locked sections  $F_l$ , e.g. the section north of San Juan Bautista and of creeping sections  $F_c$ , so that we have in the static case

$$\sum_i F_i = F_d + F_b + F_l + F_c = 0. \quad (1)$$

[6] The magnitudes of these forces are difficult to measure. If we assume a rheology for fault creep, e.g., a viscous rheology [Savage, 1971] the forces  $F_c$  on the creeping sections are proportional to the creep-rate ( $F_c \propto \dot{\epsilon}$ ). Alternatively, we might assume a power-law rheology [e.g., Wesson, 1988] or an exponential dependence as arising from a rate-and-state-dependent friction formulation [e.g., Dieterich, 1978]. The choice of rheology has no effect on the qualitative conclusions of this paper. In any case, observations of creep-rate changes give us an opportunity to track changes in  $F_c$ , and due to the force-equilibrium equation, that constrains changes in at least one of the additional quantities  $F_d$ ,  $F_b$ , and  $F_l$ .



**Figure 3.** Configuration of forces acting on a fault (see text). The shaded area represents a creeping section, the stippled area a locked portion.

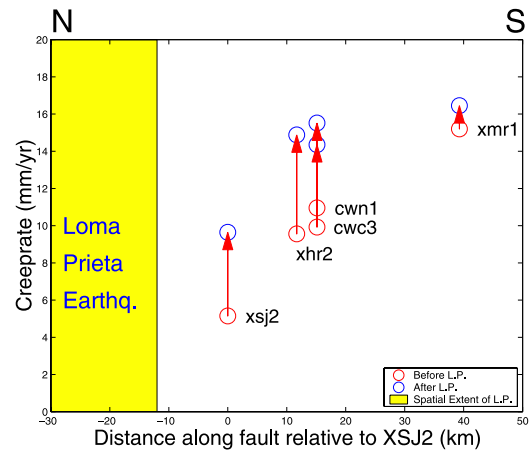
### 3.1. Changes in Resistive Forces $F_1$ From the Locked Portions of the San Andreas Fault

[7] The general coincidence of creep-rate changes (Figure 2) with the times of the largest earthquakes in the area suggests that changes in forces  $F_1$  from locked portions play an important role for creep-rate changes. In principle, these  $F_1$  are proportional to strain ( $F_1 \propto \epsilon$ ), rather than the strain rate  $\dot{\epsilon}$ . After an earthquake strain is built up gradually. Thus, the force contribution of the locked sections  $F_1$  increases with time after a large earthquake. Equation (1) shows that the force contribution from the creeping section must decrease simultaneously (and the creep rate decreases) unless either the driving forces  $F_d$  or the basal forces  $F_b$  decrease simultaneously. This creep-rate decrease is apparent after Loma Prieta (Figure 2). The behaviour after the 1961 earthquake is quite different. The two (magnitude 5.6 and 5.5) earthquakes ruptured an isolated patch on the San Andreas fault (as illustrated in Figure 3) that was sufficiently close to the creepmeter so that most of the strain was released there. The creep stopped for 18 months. Equation (1) would predict a small increase in creep rate elsewhere but there are no measurements to test this prediction.

[8] Note that the change in creep rate associated with the 1989 Loma Prieta earthquake occurs on all stations except the southernmost one, furthest away from the earthquake. The creep-rate change before-after Loma Prieta is shown in Figure 4 for all stations. The slopes are determined by fitting lines to the creep data (time windows 1982–1986 and 1994–2003). Note that the creep-rate changes increase toward the Loma Prieta quake location in the north, where the creep-rate is slower. The variation along the fault is even more dramatic for some other choices of time windows, e.g., those by *Behr et al.* [1997]. Such a decrease with distance from the locked segment is again what we expect for the release of forces on the locked section by the Loma Prieta earthquake. A nearly total stress relaxation had been suggested for the Loma Prieta earthquake [*Zoback and Beroza, 1993*], and if the creeping section were friction-free it would have to slip by 6–20 cm to accommodate that stress change [*Behr et al., 1997*]. If that range of values is correct, then there may be an increased creep-rate for another 1–2 decades. On the other hand, the leveling off after about 4 years after the Loma Prieta event that is apparent in Figure 2 (especially for CWC3, CWN1, and XHR2) might perhaps indicate that the San Andreas fault is accommodating only an excess slip of 2–4 cm, and other parts of the fault system accommodate the rest of the excess slip, e.g. the Calaveras fault [*Behr et al., 1997*]. In 1998 the creep-rates show an increase at all stations. This coincides with a strong strain transient and a magnitude 5.1 earthquake in the same area [*Johnston et al., 1998*]. Another strain transient occurred after the Loma Prieta earthquake [*Gladwin et al., 1991; Gwyther et al., 1992; Lienkaemper et al., 1991, 1997, 2001*]. These strain data may help to shed some light on the duration of excess slip. But nevertheless, it is quite apparent that the contribution from resistive forces  $F_1$  explains most of the observed long-term creep-rate changes well.

### 3.2. Changes in Plate-Tectonic Driving Forces $F_d$

[9] Changes in the plate-tectonic driving forces may in principle also cause creep-rate changes. We know little



**Figure 4.** Creep rates before and after the Loma Prieta earthquake for all stations from Figure 1, shown as a function of distance along the fault. Note the spatially systematic change of creep rate associated with the Loma Prieta earthquake. The time windows for the estimation are 1982–1986 and 1994–2003.

about such changes, but recently *Pollitz et al.* [1998] and previously *Romanowicz* [1993] suggested that the series of great earthquakes in Alaska between 1952 and 1965 produced transient deformation that propagates in the low-viscosity channel under the oceanic lithosphere, along the western margin of the North American continent. If it propagated within the oceanic lithosphere to the west of California, that would lead to temporarily increased strain across the plate boundary. The transient was predicted to arrive in Northern California in the 1970s, and somewhat later in Southern California, and the excess transient velocity predicted by that model is about 2.5 cm/yr for Northern California (and about 1.8 cm/yr for Southern California). It is known that the deformation in the area of the creeping section is strongly localized at the fault zone, and the two sides show simple “block motion” [*Brown, 1990*]. We should therefore anticipate that all or most of the excess deformation due to the transient motion would appear as a creep rate change on the San Andreas fault. Figure 2 also showed the predicted creep rate on the creeping section in comparison with observed creep rates. The predicted rate speeds up in the late 60s, but the observations don’t show a creep rate change at that time, but only 20 years later, associated with the Loma Prieta earthquake. Thus the observed creep-rate changes seem incompatible with the model of a transient change in plate-tectonic driving forces proposed by *Pollitz et al.* [1998].

### 3.3. Role of Basal Forces $F_b$

[10] During the earthquake cycle there is an exchange between resistive forces on the locked sections and viscous forces in the fault system that is governed by equation (1). After the strain-release associated with a large earthquake the locked section exerts less force, and equilibrium must be achieved by stronger viscous forces arising from the creeping section ( $F_c$ ) and the base ( $F_b$ ). This has been demonstrated for simple crustal models with a locked fault [*Huc et al., 1998; Bokelmann and Beroza, 2000*], and it is conceiv-

able that basal forces  $F_b$  and forces  $F_c$  on the creeping section play a very similar role.

#### 4. Conclusions

[11] We have presented a simple model for analyzing the role of individual forces in the long-term behaviour of a basic fault system that consists of a locked/creeping fault imbedded in an upper/lower crustal system, and we have shown that the long-term creep-rate changes are mostly due to elastic reloading of locked portions that have failed in an earthquake. An interesting exception to this is that creep-rate changes appear to start a considerable time before large and moderate-size earthquakes. This may be the case before the Loma Prieta event and especially for the 1961 quakes [see Burford, 1988; Wesson, 1988]. However, one can only speculate whether that is a general feature of larger earthquakes in the area. If this were the case this would have interesting implications for understanding earthquake nucleation [e.g., Beroza and Ellsworth, 1996] and it might even have some predictive value.

[12] While the relation of creep-rate changes with earthquakes is very clear, we could not confirm the variations in plate-tectonic driving forces that have been suggested to have occurred along the West Coast of North America during our observation interval [Pollitz *et al.*, 1998] as a result of the large earthquakes that occurred along the Aleutians in the 50s and 60s.

[13] In this paper we have focused on the first-order features of the fault system only, e.g., we ignored the complexity of the fault pattern in the area [Mavko, 1982] as well as the creep on the Hayward and the Calaveras faults [Lienkaemper *et al.*, 1991, 1997, 2001]. That detailed structure affects estimates for the total excess slip that the northern portion of the creeping section needs to undergo due to the Loma Prieta earthquakes. Estimates of the total excess slip range between 6 and 20 cm [Behr *et al.*, 1997] suggesting either that much of the excess slip has already been taken up or that it may take several decades before the creep-rate will approach the background rate again. The increased creep-rate loads smaller locked segments within the creeping section (as the one that ruptured in 1961) and that increases the risk of moderate-sized earthquakes in the San Juan Bautista area, as documented by the 1998 combination of a magnitude 5.1 earthquake [Uhrhammer *et al.*, 1999] and the aseismic transient.

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