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Kinematics of Slow-Slip Events

Chi-Yu King

Abstract

Large earthquakes are often preceded or followed by slow-slip events, which when better understood may help better understand the mechanisms of the earthquakes and the possibility of their prediction. This chapter summarizes kinematic values of large slow-slip events observed in Circum-Pacific subduction zones and creep events observed along strike-slip faults in California. The kinematic parameters include maximum slip S , duration T , rupture length L , rupture width, magnitude M , slip velocity V_S , rupture velocity V_R , and maximum slip/rupture length ratio S/L . For a large surface and subsurface creep event in California: $S = 0.9\text{--}2.5$ cm, $T = 2\text{--}5$ day, $L = 6\text{--}8$ km, $W = 3\text{--}4$ km, $M = 4.7\text{--}4.8$, $V_S = 0.4\text{--}0.5$ cm/d, $V_R = 1.6\text{--}3.0$ km/d, and $S/L = 0.3\text{--}1.5 \times 10^{-6}$. For a large short-term slow-slip event in Circum-Pacific subduction zones: $S = 1\text{--}20$ cm, $T = 2\text{--}50$ day, $L = 20\text{--}260$ km, $W = 10\text{--}90$ km, $M = 5.6\text{--}7.0$, $V_S = 0.1\text{--}0.8$ cm/d, $V_R = 2\text{--}20$ km/d, and $S/L = 0.3\text{--}1.5 \times 10^{-6}$. The latter kind of events have larger sizes in slip, duration, rupture length/width, and magnitude than the former, but are comparable in slip velocity, rupture velocity, and S/L ratio. The kinematic behaviors of both are similar, despite their large difference in temperature, pressure, and composition of the fault-zone materials. The larger size of the latter is probably due to their larger inertia caused by their larger overburden. Compared with normal earthquakes, the slip and rupture velocities of both are smaller by many orders of magnitude. But their S/L values, and thus stress drops, are smaller by only one or two orders of magnitude. For a large long-term slow-slip event in the subduction zones: $S = 1\text{--}50$ cm, $T = 50\text{--}2500$ day, $L = 40\text{--}1000$ km, $W = 30\text{--}750$ km, $M = 6.0\text{--}7.7$, $V_S = 0.01\text{--}0.10$ cm/d, $V_R = 0.1\text{--}2$ km/d, and $S/L = 0.1\text{--}2 \times 10^{-6}$. The estimated slip, duration, rupture length, and magnitude values are larger than the short-term events, but the average slip and rupture velocities are much smaller. This difference suggests that the long-term events may have commonly encountered stronger asperities, which can slow down or even break them into smaller short-term events.

Keywords: slow-slip, events, earthquake, tremor, fault zone, strike-slip, downdip, updip, plate interface, seismic, geodetic, Circum-Pacific, subduction zone, asperity, fault gouge, friction

1. Introduction

Tectonic faults may rupture rapidly (seismically) to generate earthquakes or slowly (aseismically) without doing so. During the last two decades, many slow-slip events have been discovered, especially in the subduction zones around the Pacific Ocean [1–4]. Some of them preceded or followed major earthquakes [5–15]. This chapter summarizes kinematic parameters of these slow-slip events reported in the literature and compares them with creep events and shallow slow-slip events

observed along the strike-slip faults in California. By such comparison, I hope to better understand not only the physical mechanisms of the different kinds of fault-slip behaviors but also the mechanisms of the related hydrological, geochemical, and geophysical changes often accompanying them [16, 17]. Such understanding, in turn, may help us to explore the possibility of short-term prediction of some earthquakes.

2. Creep events in California

The fact that a tectonic fault may slip slowly was first found on a surface trace of a strike-slip fault in central California [18]. Early measurements by creepmeters showed that the slow-slip motion can occur not only steadily but also episodically in small steps of several millimeters in short durations of a few days with no slip in between [19]. Subsequent measurements by widely distributed networks of creepmeters showed that the occurrence of creep events was quite common along many fault segments in central California, and elsewhere [20–22].

By studying creep events recorded at neighboring sites, Nason [20] noticed that they often began at different times, suggesting that a creep event is a rupture propagation process with a velocity of about 1–10 km/day. King et al. [23] estimated the maximum slip velocity to be about 0.01–1 cm/day. By analyzing creep data recorded at many network sites (**Figure 1**) and by fitting the creep data to a faulting model [24], King et al. [25] found that a large creep event might have a rupture length of

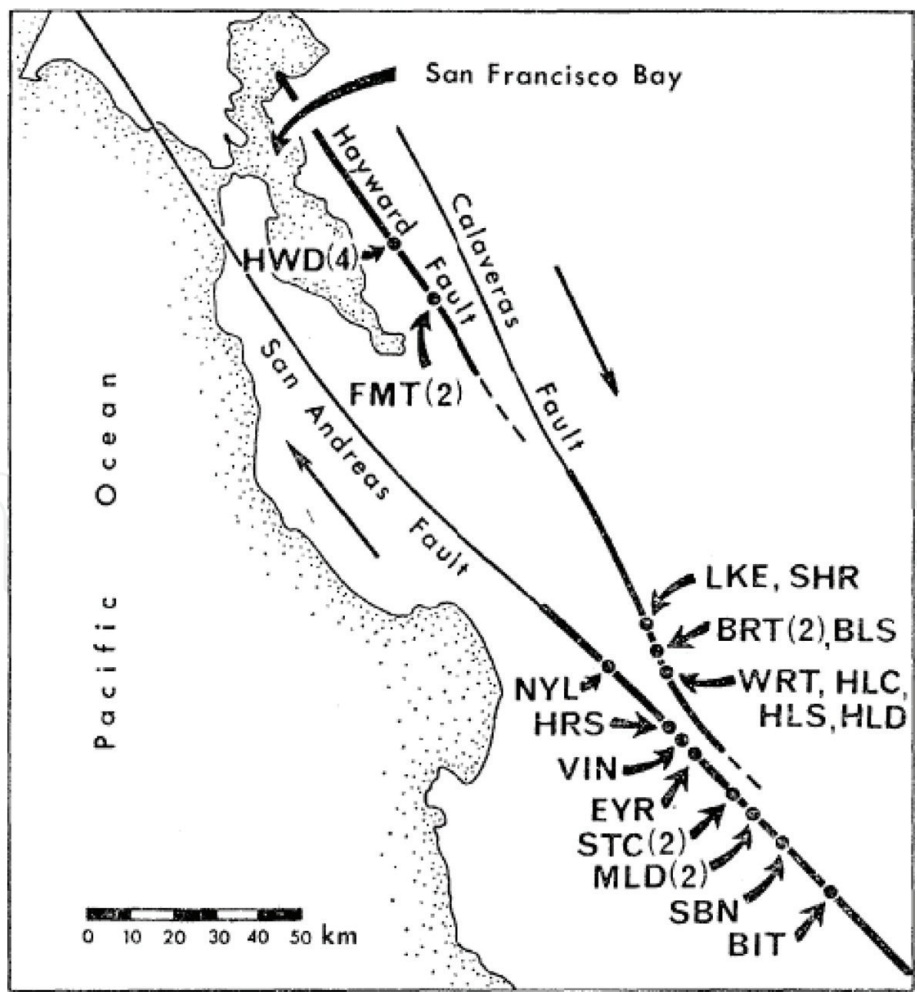


Figure 1. Distribution of creepmeters along San Andreas, Calaveras, and Hayward faults in central California. The long-term fault motion in this region is right-lateral strike slip, ranging from 0 to 3 cm/year (after King et al. [25]).

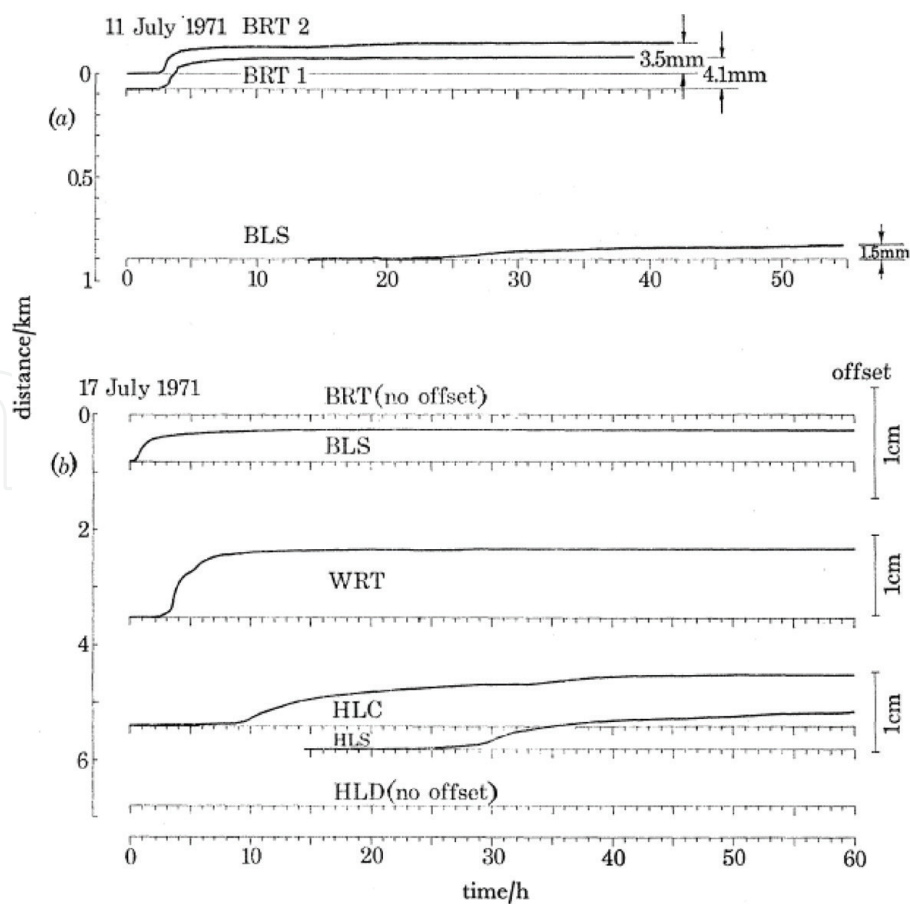


Figure 2.
Two creep events that occurred within a dense array of creepmeters on Hayward-Calaveras fault (a and b). The records are arranged properly in both time and space, except for BRT2 where the timer was out of order (after King et al. [25]).

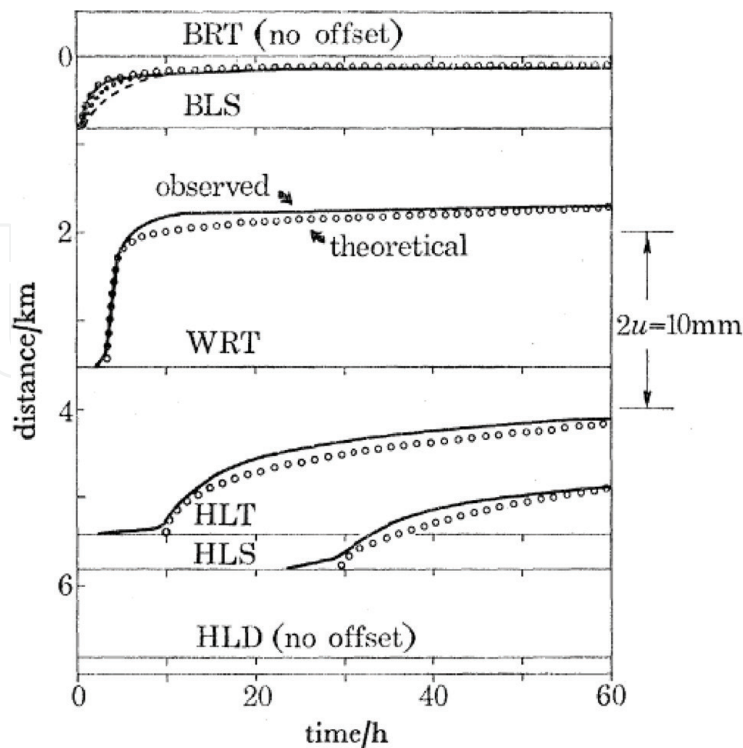


Figure 3.
Fit of the observed creep curves for the 17 July 1971 event with a theoretical model with three different guiding-center depths: 0 km (open circle), 0.5 km (dot), and 1.0 km (dashed line). The model curves are significantly different only at BLS. $2u$ is the amount of slip (after King et al. [25]).

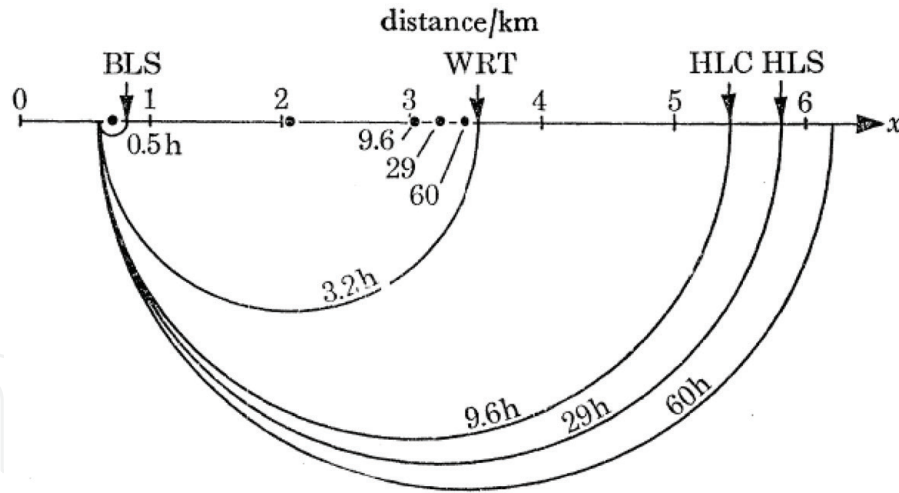


Figure 4.

Fault-plane view of a theoretical faulting model that can reasonably fit the creep data of the 17 July 1971 event (case of 0 km depth in Figure 3). The fault trace is along the x-axis, below which is the fault plane. The rupture expands with a circular boundary whose positions are shown for several specified times (after King et al. [25]).

several kilometers, an offset of about 1 cm. They also concluded that a creep event was kinematically similar to seismic faulting, but with rates that are five or more orders of magnitude smaller. **Figures 2–4** show the records of two creep events and the model fitting of the larger (**Figure 2b**) by King et al. [25], who found the following kinematic values: maximum slip $S = 0.9$ cm, duration $T = 2$ days, rupture length $L = 6$ km, rupture width $W = 3$ km, magnitude $M = 4.7$, average slip velocity $V_S = S/T = 0.45$ cm/day, average rupture velocity $V_R = L/T = 3$ km/d, and slip/rupture length ratio $S/L = 1.5 \times 10^{-6}$. The S/L ratio, which is a measure of stress drop, is an order of magnitude smaller than that of seismic faulting.

3. Subsurface slow-slip events in California

Subsurface slow-slip event was first observed at San Juan Bautista (SJB) in central California, by using two continuously recording borehole strainmeters [24]. The recorded slip curves showed some kinks, which are probably caused by some higher-resistance patches, for similar kinks were found in a theoretical faulting model by barriers [25]. This event reached an estimated depth of 4 km and lasted about 5 days with an estimated maximum slip of 2.5 cm, a rupture length of 8 km, and a magnitude of 4.8. The kinematic values are comparable to those of the surface creep event described above.

In Parkfield area of California, Guilhem and Nadeau [26] studied 52 tremor episodes, which may be related to slow-slip events about 25 km deep. They estimated that a typical event has a duration of about 10 days, maximum slip of about 7.8 mm, rupture length of about 25 km, width of 15 km, and equivalent magnitude of 5.0–5.4. Excepting the slip value, it is larger than the surface and subsurface events described above. Similar slow-slip events have been detected along other inland faults also [7, 13].

4. Slow-slip events in Circum-Pacific subduction zones

Since about the beginning of the twenty-first century, with the deployment of an increasing number of geodetic instruments, many below-surface slow-slip events have been detected, especially along plate boundaries in the subduction zones

around the Pacific Ocean, where megathrust earthquakes often occur [1, 4, 27–30]. Most slow-slip events occurred offshore and were relatively far away from onland instruments. In areas where such events were continuously detected by dense networks of geodetic and seismic instruments, the resultant data have been analyzed to estimate their kinematic values [2, 31–33].

In the Circum-Pacific subduction zones, most observed slow-slip events occur in transition areas down dip the seismogenic areas (asperities) and up dip the stable-sliding areas of the plate interfaces (e.g., **Figure 5**). Except in New Zealand and Costa Rica, they are mostly located quite far from onland instruments, including strainmeters, tiltmeters, and GPS stations [27–28, 34]. To delineate the kinematic parameters of a slip event, data have to be recorded continuously at multiple sites and inverted with the help of some elastic dislocation models, such as that by Okada [35]. Thus, the estimated kinematic values are rather uncertain. Some slow-slip events occurred in transition areas of plate interfaces up dip the seismic asperities also. They have been detected more frequently as more instruments are deployed further offshore [9, 36].

Some slow-slip events were found to be accompanied by seismic tremors and low-frequency/regular earthquakes in/near the same areas of the plate interface [29, 37–40]. This suggested that these tremors and earthquakes were generated at small asperities swept over by the rupture front of the slow-slip events. Since then, additional information about the slow-slip kinematics has been obtained from the distribution and migration of these events recorded by seismic instruments.

The depths of the observed slow-slip events usually range from 25 to 60 km; the durations in some zones show bimodal distribution of long term (of months to years) and short term (days to weeks). Short-term events in most subduction zones occurred closer to the deeper stable-sliding zone, while long-term events closer to the shallower seismogenic zone (**Figure 5**) [4]. The situation is somewhat different in New Zealand and Costa Rica, however [41].

The estimated kinematic parameters are given in the following order: maximum slip S , duration T , rupture length L , width W , magnitude M , average slip velocity $V_s = S/T$, average rupture velocity $V_R = L/T$, maximum slip/rupture length ratio $= S/L$.

In *Northeast Japan*, where the Pacific plate subducts west-northwestward beneath the North American or Okhotsk plate along the Northeast Japan Arc and the M 9.0 Tohoku-Oki megathrust earthquake occurred in 2011, a 9-year-long geodetic transient was detected that can be attributed to a very-long-term slow-slip event (possibly consisting of a series of short-term subevents) with $S = 40$ cm,

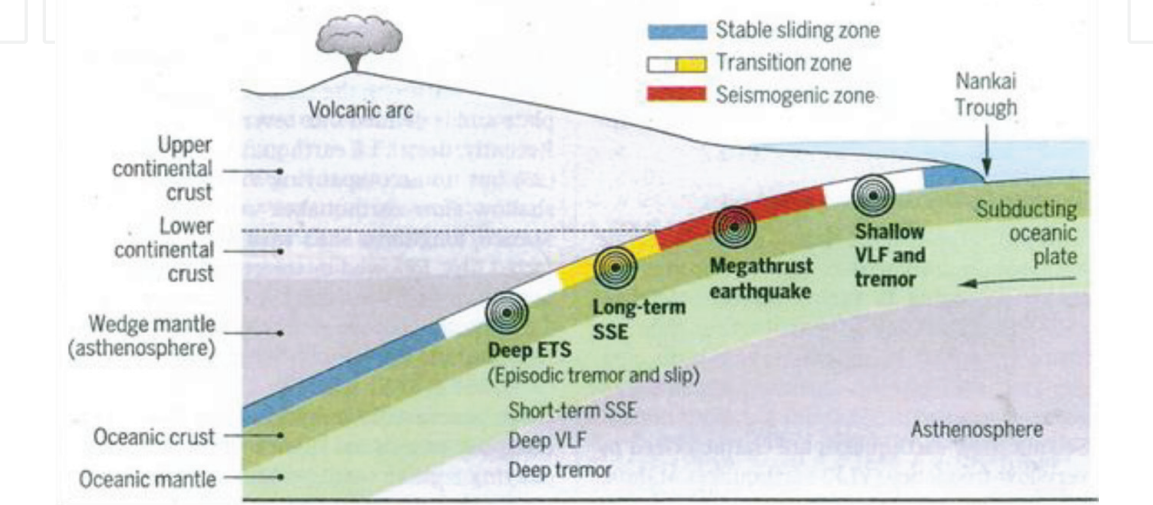


Figure 5.
A cross-sectional sketch of the Nankai subduction zone in Japan (after Obara and Kato [4]).

$T = 3000$ days, $L = 250$ km, $W = 150$ km, and $M = 7.7$. Also short-term slow-slip events here have the following kinematic values: $S = 2\text{--}20$ cm, $T = 7\text{--}35$ days, $L = 30\text{--}80$ km, $W = 30\text{--}50$ km, and $M = 6.8\text{--}7.0$ [8–9, 42–44].

Boso Peninsula is in a complicated tectonic setting, where the Philippine Sea plate subducts northwestward beneath the Okhotsk plate at the Sagami trough, and the Pacific plate subducting westward beneath the Philippine Sea plate. Slow-slip events occur here roughly every 5–7 years on the interface between the Philippine Sea plate and the Okhotsk plate at a shallow depth of about 13 km. Short-term large events have the following kinematic values: $S = 1.6\text{--}20$ cm, $T = 2\text{--}50$ days, $L = 40\text{--}100$ km, $W = 30\text{--}50$ km, $M = 6.0\text{--}6.7$ [11, 45–50].

In the tectonically more complicated Southwest Japan including Tokai region and Kii and Bungo Channels, where quite a few great earthquakes occurred in history, the Pacific plate in the east subducts westward beneath the Philippine Sea plate, which in turn subducts northwestward beneath the Eurasian plate along the Nankai trough and northeastward beneath the Okhotsk plate along the Sagami trough. Many “long-term” and “short-term” slow-slip events at depths of 30–40 km have been observed since 1997.

In *Tokai region* near the Suruga trough, where the Philippine Sea plate subducts northwestward beneath the Eurasian plate at an annual rate of 2–3 cm/year, large long-term events have the following kinematic values: $S = 7\text{--}30$ cm, $T = 2000\text{--}2500$ days, $L = 80\text{--}100$ km, $W = 60\text{--}70$ km, and $M = 6.6\text{--}7.1$. Large short-term events have the following kinematic values: $S = 0.7\text{--}1.8$ cm, $T = 2\text{--}5$ days, $L = 20\text{--}90$ km, $W = 20\text{--}40$ km, and $M = 5.6\text{--}6.2$ [38, 50–53].

In/near *Kii channel*, long-term slow-slip events have the following kinematic values: $T = 398$ days and $M = 6.7$ [54]. Short-term slow-slip events have the following kinematic values: $S = 1\text{--}2$ cm, $T = 2\text{--}5$ days, $L = 20\text{--}90$ km, $W = 20\text{--}30$ km, and $M = 5.3\text{--}6.1$ [36, 38, 55].

In/near *Bungo Channel*, large long-term events have the following kinematic values: $S = 1\text{--}50$ cm, $T = 90\text{--}700$ days, $L = 40\text{--}200$ km, $W = 40\text{--}100$ km, and $M = 6.0\text{--}7.3$. Some of them were found to possibly consist of multiple short-term events. Large short-term events have the following kinematic values: $S = 1\text{--}4$ cm, $T = 4\text{--}10$ days, $L = 20\text{--}100$ km, $W = 20\text{--}50$ km, and $M = 5.8\text{--}6.3$ [27, 38, 56–64].

In *Gisborne/Raukumara Peninsula, New Zealand*, where the Pacific plate subducts westward beneath the eastern North Island at the Hikurangi subduction zone, long-term events down dip the seismogenic interface area at the depth of 25–60 km in the southern margin have the following kinematic values: $S = 4\text{--}56$ cm, $T = 50\text{--}550$ days, $L = 60\text{--}200$ km, $W = 30\text{--}150$ km, and $M = 6.5\text{--}7.2$. Short-term events up dip the seismogenic area (5–15 km deep) along northern Hikurangi margin have the following kinematic values: $S = 1.2\text{--}24$ cm, $T = 5\text{--}36$ days, $L = 50\text{--}180$ km, $W = 50\text{--}90$ km, and $M = 5.8\text{--}7.0$ [41, 65–75].

In *Alaska* subduction zone, where the great $M_w = 9.21964$ Prince William Sound earthquake ruptured a large portion of the shallow plate interface above 30 km depth at the eastern end, several long-term slow-slip events were detected just below the seismogenic zone at depths between 25 and 45 km with the following kinematic values: $S = 2\text{--}40$ cm, $T = 620\text{--}1600$ days, $L = 150\text{--}1000$ km, $W = 140\text{--}750$ km, and $M = 6.9\text{--}7.5$ [76–78].

In *Cascadia* subduction zone, where the Juan de Fuca plate subducts beneath the North American plate, geodetic measurements show that the plate interface along an entire 1000 km segment between British Columbia and northern California is locked from near the surface to a depth of about 20 km [84]. No long-term slow-slip event has been detected here. Large short-term slow-slip events at depths of 30–55 km in this segment have the following kinematic values: $S = 1\text{--}8$ cm, $T = 7\text{--}50$ days, $L = 25\text{--}400$ km, $W = 25\text{--}70$ km, and $M = 6.1\text{--}6.9$ [2, 28, 79–86].

In *Mexico*, where the Cocos plate subducts beneath the North American plate along the Middle American Trench and great earthquakes occurred every 30–100 years, several long-term slow-slip events were detected below the seismogenic depth of 15–40 km with the following kinematic values: $S = 9\text{--}30\text{ cm}$, $T = 90\text{--}400\text{ days}$, $L = 200\text{--}500\text{ km}$, and $W = 150\text{--}230\text{ km}$. Short-term events have the following kinematic values: $S = 2\text{--}10\text{ cm}$, $T = 30\text{--}45\text{ days}$, $L = 200\text{--}260\text{ km}$, $W = 50\text{--}130\text{ km}$, and $M = 6.3\text{--}7.2$ [10, 87–96].

In northwestern *Costa Rica*, the Nicoya Peninsula is located along the Middle America Trench where the Cocos plate subducts beneath the Caribbean plate at about 8 cm/yr. The subduction segment has ruptured repeatedly in the past. The peninsula lies directly over the seismogenic zone, and several slow-slip events possibly updip the seismic interface were detected with the following kinematic values: $S = 1.5\text{--}15\text{ cm}$, $T = 20\text{--}180\text{ days}$, $L = 30\text{--}120\text{ km}$, $W = 20\text{--}40\text{ km}$, and $M = 6.6\text{--}7.2$ [97–103].

Events	S (cm)	T (day)	L (km)	W (km)	M	S/T (cm/d)	L/T (km/d)	S/L
California surface creep	0.9	2	6	3	4.7	0.45	3	1.5
SJB subsurface slow slip	2.5	5	8	4	4.8	0.5	1.6	0.31
Parkfield deep slow slip	0.8	10	25	15	5.0–5.4	0.08	2.5	0.03
Circum-Pacific short term								
Northeast Japan	2–20	7–35	30–80	30–50	6.8–7.0	0.29–0.57	2.3–4.3	0.67–2.50
Boso Peninsula	1.6–2.0	2–50	40–100	30–50	6.0–6.7	0.04–0.80	2–20	0.20–0.40
Tokai region	0.7–1.8	2–5	20–90	20–40	5.6–6.2	0.35–0.36	10–18	0.20–0.35
Kii Channel	1–2	2–5	20–90	20–30	5.3–6.1	0.40–0.50	10–18	0.22–0.50
Bungo Channel	1–4	4–10	20–100	20–50	5.8–6.3	0.25–0.40	5–10	0.40–0.50
Hikurangi, New Zealand	1.2–24	5–36	50–180	50–90	5.8–7.0	0.24–0.67	5–10	0.24–1.33
Cascadia	1–8	7–50	25–400	25–70	6.1–6.9	0.14–0.16	3.6–8.0	0.20–0.40
Mexico	2–10	30–45	200–260	50–130	6.3–7.2	0.07–0.22	5.8–6.7	0.10–0.38
Northwestern Costa Rica	1.5–15	20–180	30–120	20–40	6.6–6.7	0.07–0.08	0.7–1.5	0.50–1.25
Central Ecuador	8–40	4–40	30–80	10–60	6.0–6.8	1.00–2.00	2.0–7.5	2.67–5.00
Chile	1.3–8	2–15	20–60	20–30	6.5–6.7	0.53–0.65	4–10	0.65–1.33
Circum-Pacific long term								
Northeast Japan	40	3000	250	150	7.7	0.013	0.08	1.60
Tokai region	7–30	2000–2500	80–100	60–70	6.6–7.1	0.004–0.012	0.04	0.86–3.00
Bungo Channel	1–50	90–700	40–200	40–100	6.0–7.3	0.01–0.07	0.29–0.44	0.25–2.50
Hikurangi, New Zealand	4–56	50–550	60–200	30–150	6.5–7.2	0.08–0.10	0.36–1.20	0.67–2.80
Alaska	2–40	620–1600	150–1000	140–750	6.9–7.5	0.003–0.025	0.24–0.63	0.13–0.04
Mexico	9–30	90–400	200–500	150–230	6.5–7.6	0.08–0.10	1.25–2.22	0.45–0.60
Chile	50–80	240	70–150	20–30	6.5–6.9	0.21–0.33	0.29–0.63	5.33–7.14

Table 1.
Kinematic values.

In the Central *Ecuador* subduction zone, short-term slow-slip events along this segment of the North Andean subduction zone, where the Nazca plate subducts beneath South America plate, are estimated to have the following kinematic values: $S = 8\text{--}40\text{ cm}$, $T = 4\text{--}40\text{ days}$, $L = 30\text{--}80\text{ km}$, $W = 10\text{--}60\text{ km}$, and $M = 6.0\text{--}6.8$ [42, 104–105].

In *Chile*, along a megathrust fault off northernmost Chile, where the Nazca plate subducts beneath the South American plate, a long-term slow-slip event occurred in a transition zone at depth of 40–60 km with the following kinematic values: $S = 50\text{--}80\text{ cm}$, $T = 240$, $L = 70\text{--}150\text{ km}$, $W = 20\text{--}30\text{ km}$, and $M = 6.5\text{--}6.9$. Also a short-term event indicated by earthquake migration occurred with $S = 1.3\text{--}8\text{ cm}$, $T = 2\text{--}15\text{ days}$, $L = 20\text{--}60\text{ km}$, $W = 20\text{--}30\text{ km}$, and $M = 6.5\text{--}6.7$ [11, 106–108].

Table 1 gives a summary of the estimated kinematic values of slip S , duration T , rupture length L , rupture width W , and magnitude M , as well as calculated values of slip velocity $V_S = S/T$, rupture velocity $V_R = L/T$, and S/L , which is a measure of stress drop. In this table, a factor of 10^{-6} is omitted from the S/L values.

5. Discussion

Among the short-term slow-slip events included in **Table 1**, those in New Zealand and Costa Rica occurred updip the seismogenic area of the subduction interface. Yet the estimated kinematic values are comparable to those downdip the seismogenic interface areas in other subduction zones. On the other hand, the kinematic values for central Ecuador are quite different, due to the unusually large slips reported. The same is true with Chile in the case of long-term events, and the opposite is true for the Parkfield events, when compared with two other cases in California. In the following discussion, we shall exclude these three cases from further consideration.

It may be seen that most short-term slow-slip events in the various Circum-Pacific subduction zones have comparable kinematic values: a slip of about 1–20 cm, duration of 2–50 days, rupture length of 20–260 km, width of 10–90 km, magnitude of 5.6–7.0, slip velocity of 0.1–0.8 cm/d, rupture velocity of 2–20 km/d, and an S/L value of 0.1–1.3. Compared with creep and shallow slow-slip events in California, they have larger values in slip, duration, rupture length/width, and magnitude, but comparable values in slip velocity, rupture velocity, and S/L ratio. This result indicates that kinematics of slow-slip events are basically similar, independent of temperature, pressure, and composition of the fault-zone materials, as long as they are mostly velocity-strengthening fault-gouge type; the larger sizes of the subduction events are probably due to their larger inertia associated with larger overburden at greater depth. Compared with normal earthquakes, the slip and rupture velocities of the slow-slip events are all smaller by many orders of magnitude, and the estimated S/L values are smaller by one or two orders of magnitude. Since S/L is proportional to stress drop, this result shows that the stress drops for the slow-slip events are one or two orders of magnitude smaller than normal earthquakes.

The long-term events have large variations in their kinematic values: slip of about 1–50 cm, duration of 50–2500 days, rupture length of 40–1000 km, width of 30–750 km, magnitude of 6.0–7.7, slip velocity of 0.01–0.10 cm/d, rupture velocity of 0.1–2 km/d, and S/L of 0.2–3. Compared with the short-term events, they have larger values in slip, duration, rupture length, and magnitude, comparable values in slip/rupture length ratios (thus stress drops) and smaller values in average slip and rupture velocities. This feature arose may be because they occurred closer to the seismogenic area of the plate interface [4], which has more asperities to hinder the

rupture process or even to break them into smaller events, as shown in the cases of northeast Japan and Bungo Channel. Being more remote from the instruments, they are less distinguishable, thus giving an appearance of slower propagation. This possibility is further supported by some recent analysis of long-term events [109, 110].

The reason why slow-slip events and seismic tremors occur in the transition areas, both downdip and updip the seismogenic area, of a subduction interface (e.g., **Figure 5**) is not well known, but may possibly be understood by the following consideration of a five-stage evolution of a subducting seafloor, which consists of seamounts of different heights and strengths embedded in sediments at its surface. At the initial stage of subduction, while under the front edge of the accretionary wedge where the temperature and confining pressure are low, the effect of heterogeneity of sliding friction is small, and thus fault slip caused by crustal convergence proceeds in the form of aseismic stable sliding. When the same seafloor subducts further down and encounter larger confining pressure but still relatively low temperature, the friction at the stronger patches (seamounts) begins to show its stick-slip (velocity weakening) feature in sliding, while the sedimentary parts acts like (velocity strengthening) fault gouge. As a relatively strong asperity breaks under increasing shear stress, the rupture propagates slowly into a larger area of interface consists of weaker asperities embedded in compliant gouge materials, thus causing slow-slip events and small earthquakes in the transition zone updip the seismogenic area of the plate interface. When the same seafloor subducts further down to the seismogenic depth, where the confining pressure becomes sufficiently large while the temperature is not, the heterogeneity contrast becomes very sharp, and thus when a strong patch (asperity) breaks, it may cause a rupture to propagate rapidly into a large area of the interface, sweeping through smaller asperities embedded in compliant gouge materials and resulting in a large or even megathrust earthquake in the seismogenic area of the interface. When the seafloor subducts further down and encounter still larger confining pressure and higher temperature, the large asperities may have been worn down by now and become softened, while the surrounding fault-gouge materials further cumulated in volume and strength. When such a weaker asperity breaks, it encounters stronger resistance and thus may cause only a slowly rupturing event in the transition zone of the downdip area; as the rupture sweeps through some even smaller asperities, it may cause seismic tremors and perhaps small earthquakes. When the seafloor subducts further down and encounters still higher confining pressure and temperature at the deepest level of subduction, the asperities may have become sufficiently worn and softened and the gouge materials further cumulated; the frictional heterogeneity may finally become insignificant and thus the sliding becomes stable again.

What further directions should be pursued? Besides acquiring additional high-quality data through more continuous monitoring efforts closer to the events, it is important to analyze the data with appropriate faulting models to better understand the mechanics of slow-slip events and their role as a silent agent in stress adjustment along fault zones. Such knowledge should help us to better understand the occurrence pattern of earthquakes, such as foreshocks, main shocks, aftershocks, earthquake swarm, and earthquake migration [111], as well as crustal deformation without earthquakes [112]. It may also help us better understand the mechanisms of various earthquake-related hydrological, geochemical, and geophysical changes [16]. Together with better monitoring efforts of such changes, especially those that precede earthquakes [14, 15], it may finally be possible to predict some destructive earthquakes and aftershocks.

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