NPTEL ONLINE CERTIFICATION COURSES

EARTHQUAKE SEISMOLOGY

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Module 10 : Brief on Earthquake geodesy

Lecture 04: Interseismic deformation and the seismic cycle

CONCEPTS COVERED

- Interseismic deformation and the seismic cycle
  - Interseismic shear strain rate
  - Advantages Of Geodesy
- Summary

Recap

- **Geodetic methods using signals from space permits all three components of position to be measured to sub-centimeter precision and now give coseismic motion to high precision much more easily than was previously possible.**

- **Very Long Baseline Interferometry (VLBI), Satellite Laser Ranging (SLR), and Global positioning system (GPS) are popular geodetic techniques to measure ground deformation.**

- **GPS uses a constellation of satellites transmit coded timing signals on a pair of microwave carrier frequencies synchronized to very precise on-board atomic clocks.**

- **GPS data can be obtained for continuous period of time or short period of time. The later one is cheaper.**

- **The biggest limitation of geodetic data for earthquake studies is that the positions of geodetic markers before the earthquake are needed.**

Recap

- For radar, d is the antenna length, so a radar a distance r above the earth’s surface could resolve objects of size x, where $$\theta_d = \lambda / d = x / r$$

- The phase difference between radar signals with wavelength λ reflected from the Earth’s surface and recorded by antennas at position A_1 and A_2 is $$\phi = (4\pi/\lambda)(r_2 - r_1)$$

  r_i is the range from the antenna at A_i to the reflection point.

- If differences in satellite positions between the measurements are removed, a vector surface displacement D causes a phase change

$$\phi \approx (4\pi/\lambda)\delta r, \quad \delta r = (D.\hat{r}),$$

where δr is the projection of the vector displacement along, the look direction connecting the satellite and reflection point.

- The results are shown as a phase difference map, called a differential interferogram.

Recap

- **Static coseismic displacement contain $1/r^2$ terms, compared to $1/r$ terms for the propagating waves. Thus, it decay more rapidly with distance from the earthquake.**

- **For infinite length fault, the fault-parallel displacement in the x direction, u(y), varies with distance from the fault y as**

$$u(y) = \pm D/2 - (D/\pi)\tan^{-1}(y/W)$$

- **For finite length faults the displacement tapers off rapidly past the fault ends. If a fault is buried and extends from depth w to depth W**

$$u(y) = (D/\pi)\left[\tan^{-1}(y/w) - \tan^{-1}(y/W)\right]$$

- **A fault that does not reach the surface, the displacement is both reduced in amplitude and varies more smoothly with distance than it would for a fault extending to the surface.**
- **Estimation of fault parameter using geodetic data is an inverse problem and has highly non unique solutions.**

Recap

- Seismic waves have an ambiguity in distinguishing between the fault plane and the auxiliary plane, the geodetic data do not as static displacements models do not have nodal plane perpendicular to fault plane.

- We use strong motion data, Teleseismic data and geodetic data that provides good constraints on fault geometry and slip on it.

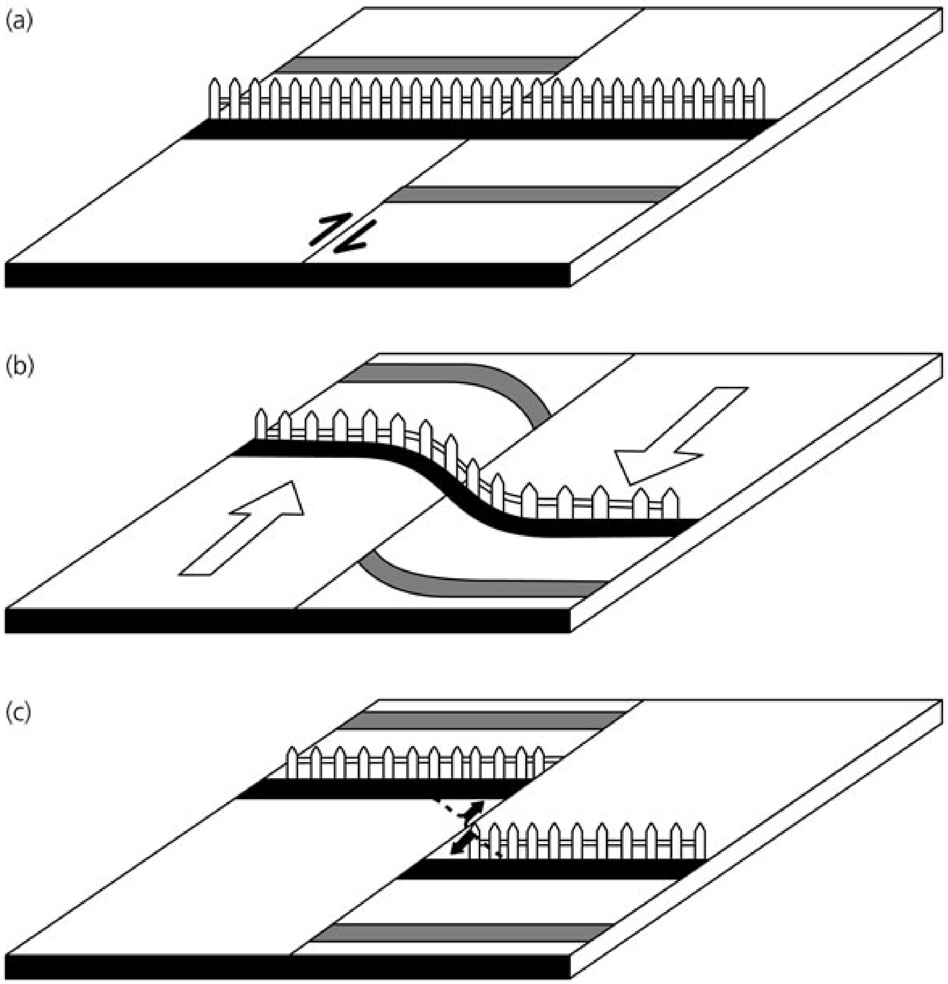
- Geodetic data that depend on the difference in position before and after an earthquake provide no information about what happened during the earthquake, whereas seismological data can sometimes show how the rupture evolved

- The results for the different data types differ because each is sensitive to different features of the slip

- In post-seismic slip deformation goes on “silently” (without a seismic signal) for some time after an earthquake and its seismologically observed aftershocks

Interseismic deformation and the seismic cycle

Figure 4.1-3: Cartoon of the elastic rebound model for a strike-slip earthquake.



- ➔ Geodesy gives insight into the seismic cycle before, after, and between earthquakes, whereas we can only study the seismic waves once an earthquake occurs.

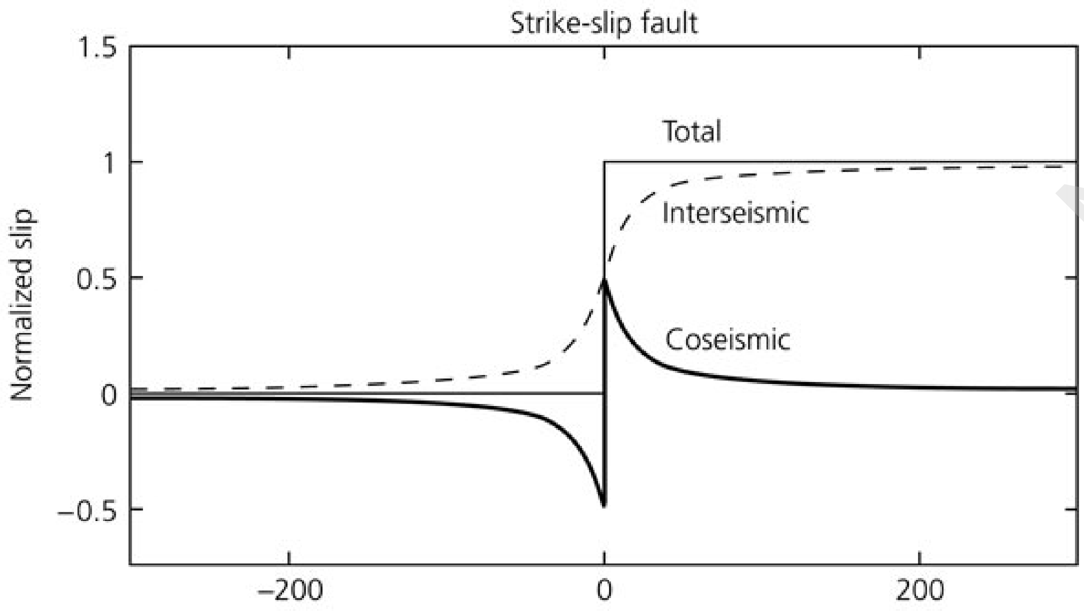
- ➔ Consider a simple elastic rebound model of an infinite strike-slip fault at a plate boundary, assuming that large earthquakes release all the strain which accumulates between earthquakes.

Interseismic deformation and the seismic cycle

- After an earthquake, material on the right (+y) side far from the fault moves at the far-field rate v relative to the left (−y) side of the fault, and so has moved a distance vt by time t.

- However, between earthquakes the fault is locked down to depth W, although it slips freely below, so material at the fault does not move between earthquakes.

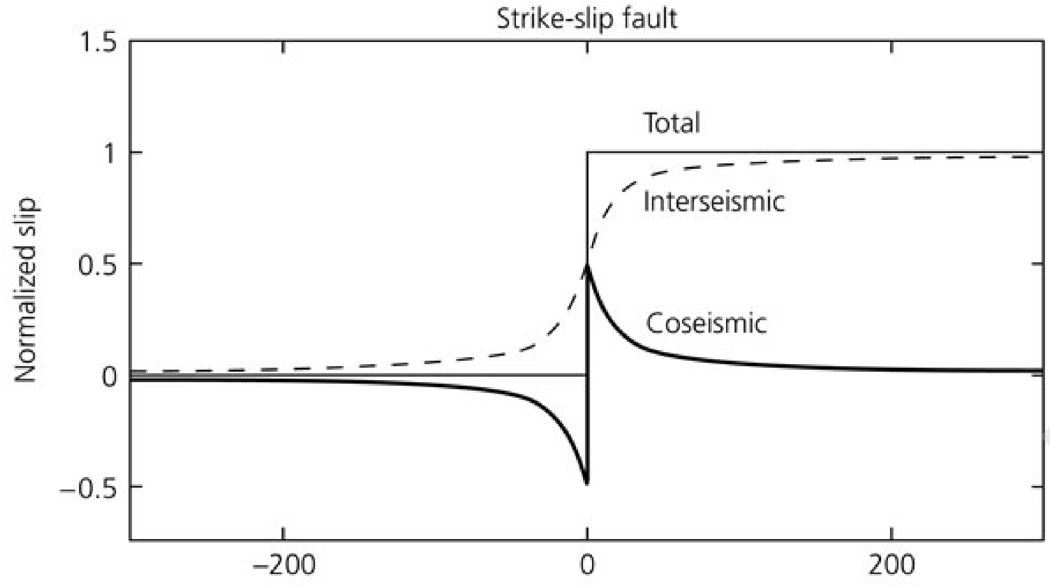
Figure 4.5-12: Coseismic and interseismic slips and strains.



- When the next large earthquake occurs, completing the seismic cycle, everything to the right of the fault must have moved a distance vt.

Interseismic deformation and the seismic cycle

Figure 4.5-12: Coseismic and interseismic slips and strains.



- **The earthquake's coseismic displacement will be given by with D = vt.**

$$u(y) = \pm D/2 - (D/\pi)\tan^{-1}(y/W)$$

- **The coseismic slip u(y) is less than D except at the fault.**
- **Points away from the fault already have moved part of the distance D before the earthquake.**
- **Everything on the left side must have had no net motion from the seismic cycle, even though material near the fault moved "backward" (in the −x direction) during the earthquake.**

Interseismic deformation and the seismic cycle

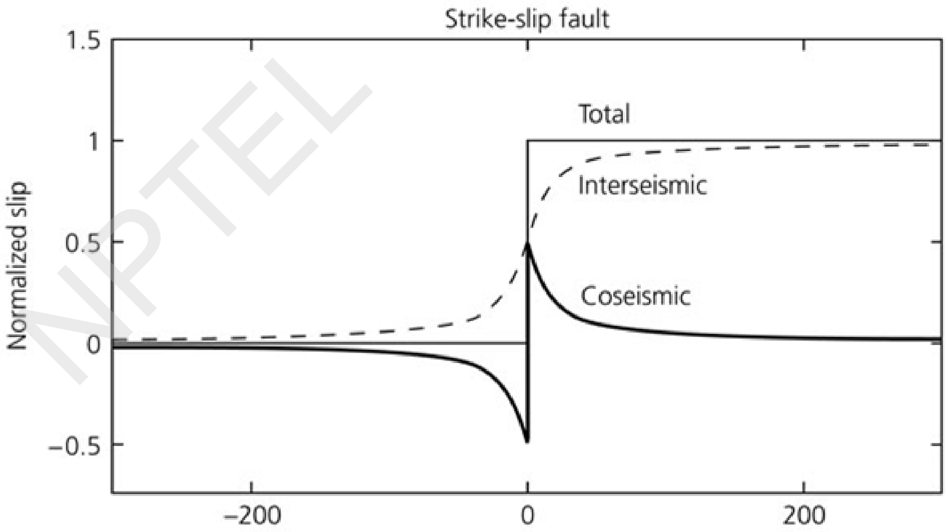
- **Fault-parallel interseismic motion s(y) is found by subtracting the coseismic slip from the far-field (or net) motion, giving**

$$s(y) = D/2 + D/\pi \tan^{-1}(y/W) \qquad \text{.... equ. 1}$$

- **Material on the left side near the locked fault is "dragged along" during the interseismic period, and then rebounds during the earthquake.**

- **Material on the right side near the fault is retarded during the interseismic period, and then "catches up" to the far-field motion due to the coseismic deformation.**

Figure 4.5-12: Coseismic and interseismic slips and strains.



Interseismic deformation and the seismic cycle

- If the fault is a plate boundary, the interseismic deformation occurs over a finite plate boundary zone within which sites on either side of the boundary move relative to the interior of the plate they are on. In this case, the boundary zone is relatively narrow, comparable to the depth to which the fault is locked.

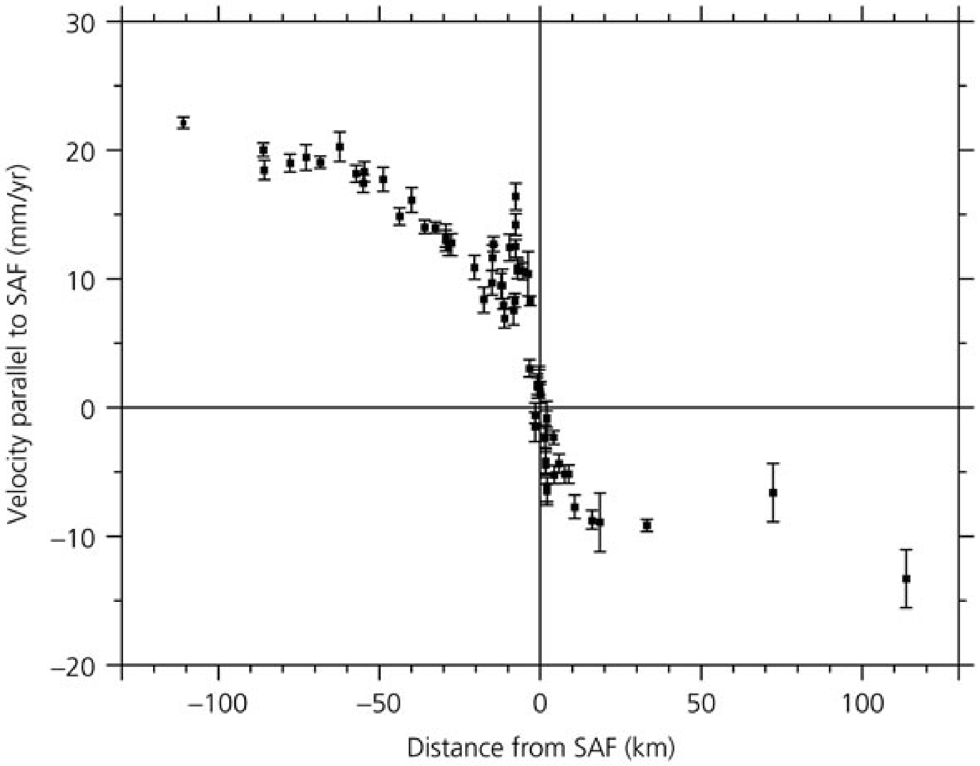
- However, many plate boundary zones are broader because additional faults take up some of the plate motion.

- Because the interseismic motion is the difference between the far-field motion and coseismic deformation, its variation with distance from the fault depends on the locking depth and farfield rate.

Interseismic deformation and the seismic cycle

- Shallow locking concentrates interseismic slip near the fault, whereas deeper locking spreads it out into a broad shear zone

Figure 4.5-13: Fault-parallel horizontal interseismic motion across the San Andreas fault.



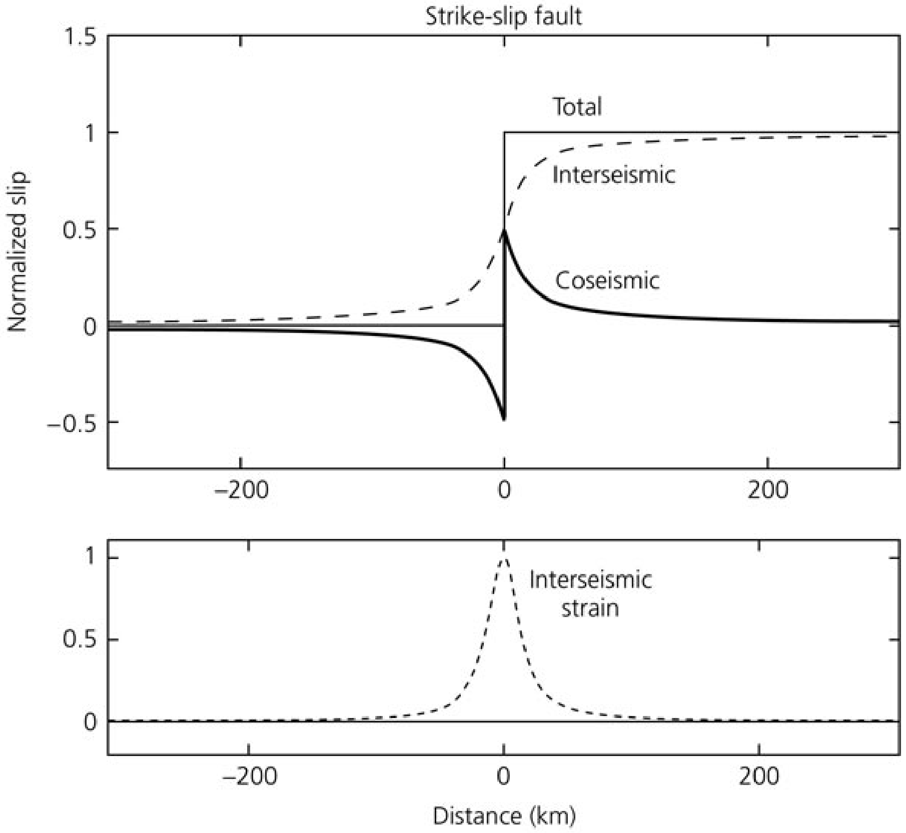
- Comparison with the coseismic slip shows that the width of the zone across which the motion changes rapidly depends on the locking depth.

Interseismic shear strain rate

We can use Eqn 1 to find the interseismic shear strain rate

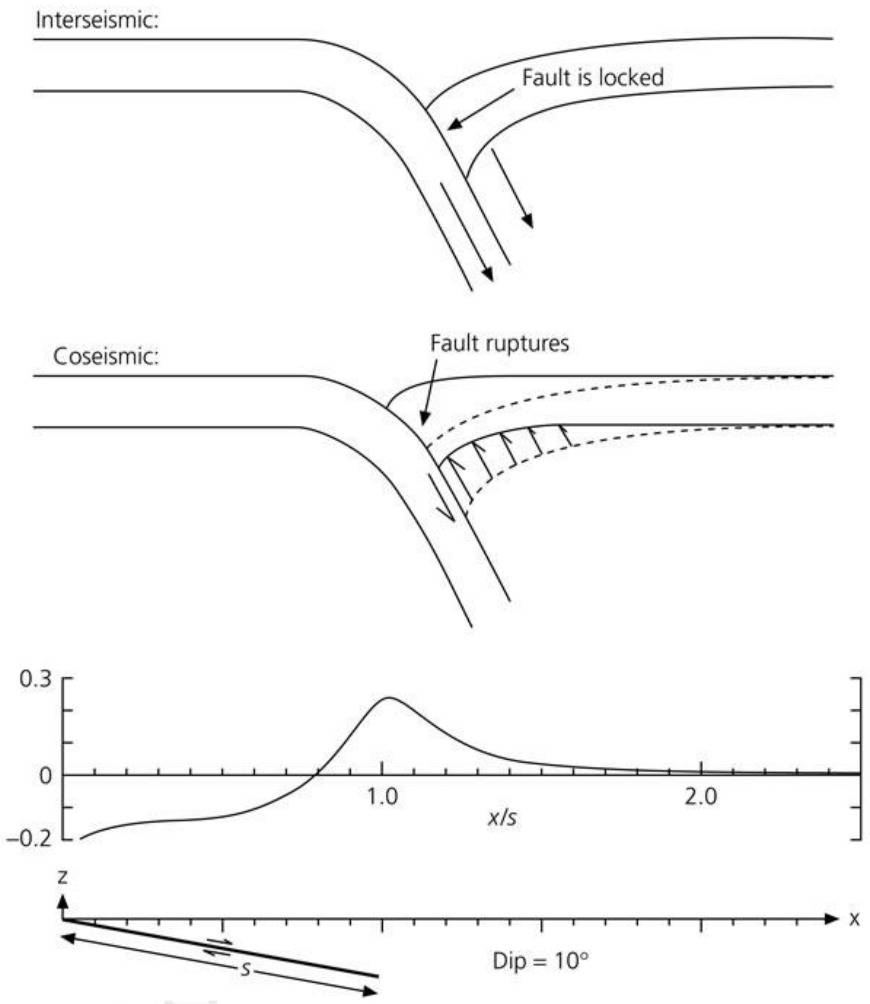
$$\dot{e}_{xy} = \frac{1}{2}\frac{ds(y)}{dy} = \frac{v}{2\pi W}\frac{1}{\left[1+(y/W)^2\right]}$$

Figure 4.5-12: Coseismic and interseismic slips and strains.



- **Strain accumulates near the fault during the interseismic period and is released in large earthquakes.**
- **Like the displacement, the variation of strain with distance from the fault depends on the locking depth and far-field rate.**
- **The strain rate can be inferred from changes in the angles between geodetic markers.**

Figure 4.5-14: Predicted interseismic vertical motions due to a locked fault at a subduction zone.

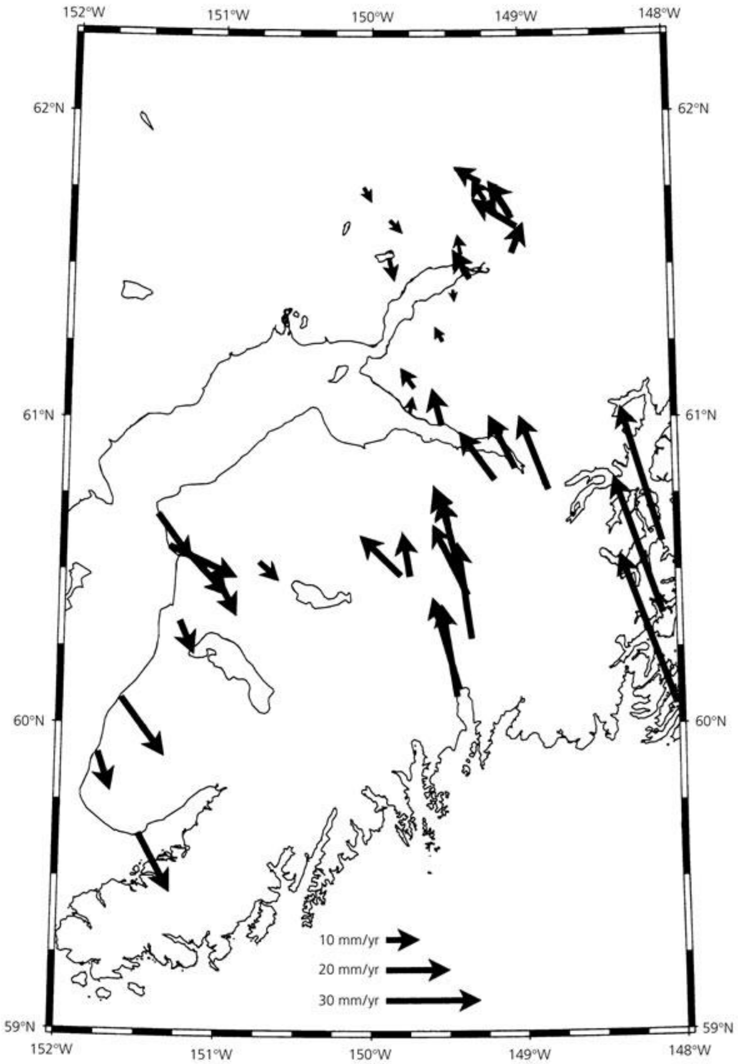


- ➔ **So far we have seen example of strike-slip fault. But the approach is similar for thrust fault at subduction zones.**

- ➔ **For the figure on the left, modeling predicts interseismic subsidence and landward motion for most sites above the locked fault, and uplift further inland.**

- ➔ **Thus geodetic data near trenches can identify the interseismic deformation and provide insight into the mechanics of the subduction interface and future large earthquakes on it.**

Figure 4.5-15: GPS velocities relative to North America for sites near the 1964 Alaska earthquake rupture zone.



- ➔ **Figure shows GPS velocities relative to the stable interior of North America for some sites near the rupture zone of the great 1964 Alaska earthquake.**
- ➔ **Sites to the east of the area shown move northwest, in the direction of Pacific plate subduction beneath North America, as we would expect for the interseismic motion of sites on the overriding plate above a locked fault.**

- ➔ **The motion decays rapidly landward with distance from the trench.**
- ➔ **These observations together with the observed uplift, are reasonably consistent with the interseismic motion.**

Advantage of geodesy

- In general, geodetic data from the interseismic period give insight into the mechanics of a fault and future earthquakes on it, even before they occur.

- This is gratifying because the seismic cycle is so long, typically hundreds of years, that we generally have to wait a long time to study a major earthquake on a given fault segment.

- Consider measuring the rate v of motion of a monument that started at position x_1 and reaches x_2 in time T. If the position uncertainty is given by its standard deviation σ, then the propagation of errors relation shows that

$$v = (x_1 - x_2)/T \quad \Longrightarrow \quad \sigma_v = \sqrt{2}\sigma/T$$

where σ_v is the uncertainty of the inferred rate. Thus the longer we wait, the smaller the velocity uncertainty becomes, even if the data do not become more precise.

Advantage of geodesy

- The geodetic data let us see the rate at which locked slip is accumulating, and hence infer the maximum possible slip in a future earthquake, depending on when it occurs.

- We can also estimate the time until a future earthquake from records of past earthquakes, by assuming what the coseismic slip will be.

Advantage of geodesy

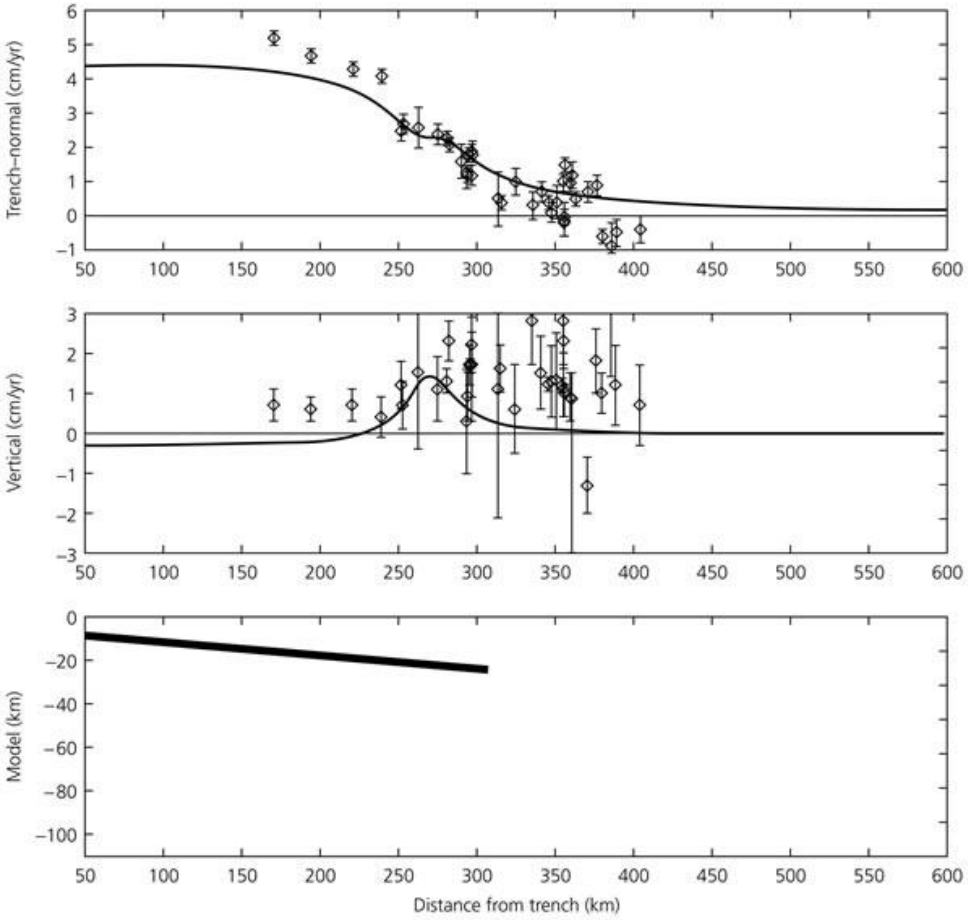
In some places, geodetic data imply that slip is accumulating on the locked fault at a rate less than the far-field motion.

This difference seems to be due to plate motion taken up elsewhere.

The difference is thought to indicate that some of the plate boundary slip occurs by aseismic slip or sliding (perhaps as "silent earthquakes") on the fault, and hence will not appear in future earthquakes.

Advantage of geodesy

Figure 4.5-16: Horizontal and vertical GPS velocities relative to North America at the sites in Fig. 4.5-15.



- **Profiles of horizontal (top) and vertical (center) GPS velocities relative to North America for eastern sites.**

- **The data are reasonably similar to predictions (solid line) for a locked fault model (bottom).**

- **Note that uncertainties for the vertical GPS data are larger than for the horizontal data.**

- **Plate interface has a shallow dip so there is a large fault area at depths shallow enough to accumulate strain and then rupture.**

Summary

- **Fault-parallel interseismic motion s(y) is given as:** $s(y) = D/2 + D/\pi \tan^{-1}(y/W)$

- **The coseismic slip u(y) is less than D = vt except at the fault.**

- **Interseismic motion is the difference between the far-field motion and coseismic deformation, its variation with distance from the fault depends on the locking depth and farfield rate.**

- **Interseismic shear strain rate** $\dot{e}_{xy} = \dfrac{1}{2}\dfrac{ds(y)}{dy} = \dfrac{v}{2\pi W}\dfrac{1}{\left[1 + (y/W)^2\right]}$

- **Consider measuring the rate v of motion of a monument that started at position x_1 and reaches x_2 in time T. If the position uncertainty is given by its standard deviation σ, then the propagation of errors relation shows that**

$$v = (x_1 - x_2)/T \implies \sigma_v = \sqrt{2}\sigma/T$$

where σ_v is the uncertainty of the inferred rate.

REFERENCES

- Stein, Seth, and Michael Wysession. An introduction to seismology, earthquakes, and earth structure. John Wiley & Sons, 2009.
- Lowrie, William, and Andreas Fichtner. Fundamentals of geophysics. Cambridge university press, 2020.
- Kearey, Philip, Michael Brooks, and Ian Hill. An introduction to geophysical exploration. Vol. 4. John Wiley & Sons, 2002.
- https://geologyscience.com/geology-branches/structural-geology/stress-and-strain/
- Seismology course, Professor Derek Schutt, Colorado State Univ., USA.

THANK YOU !