Simple Elastic Dislocation Models for Interseismic Deformation in Subduction Zones

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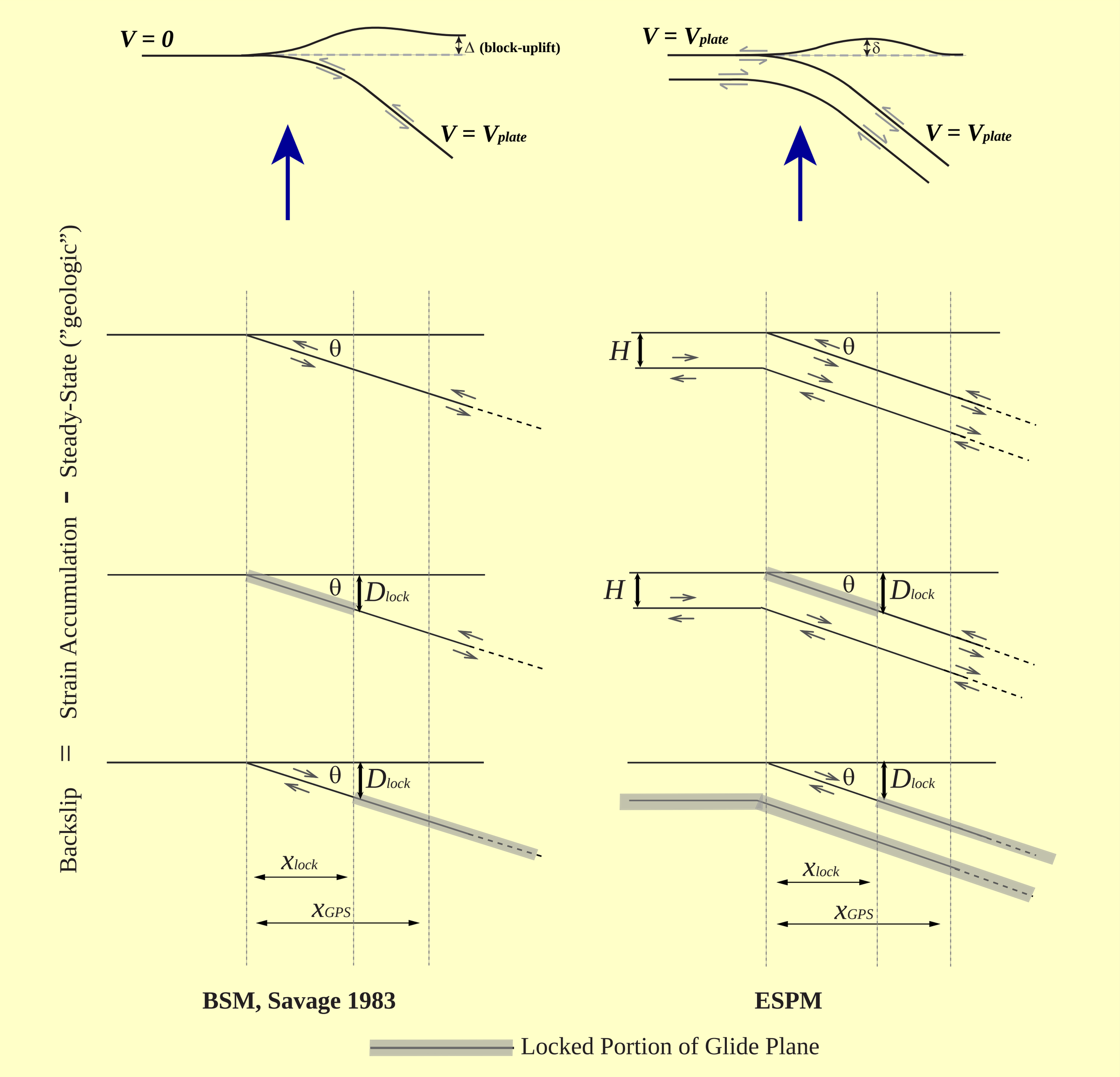
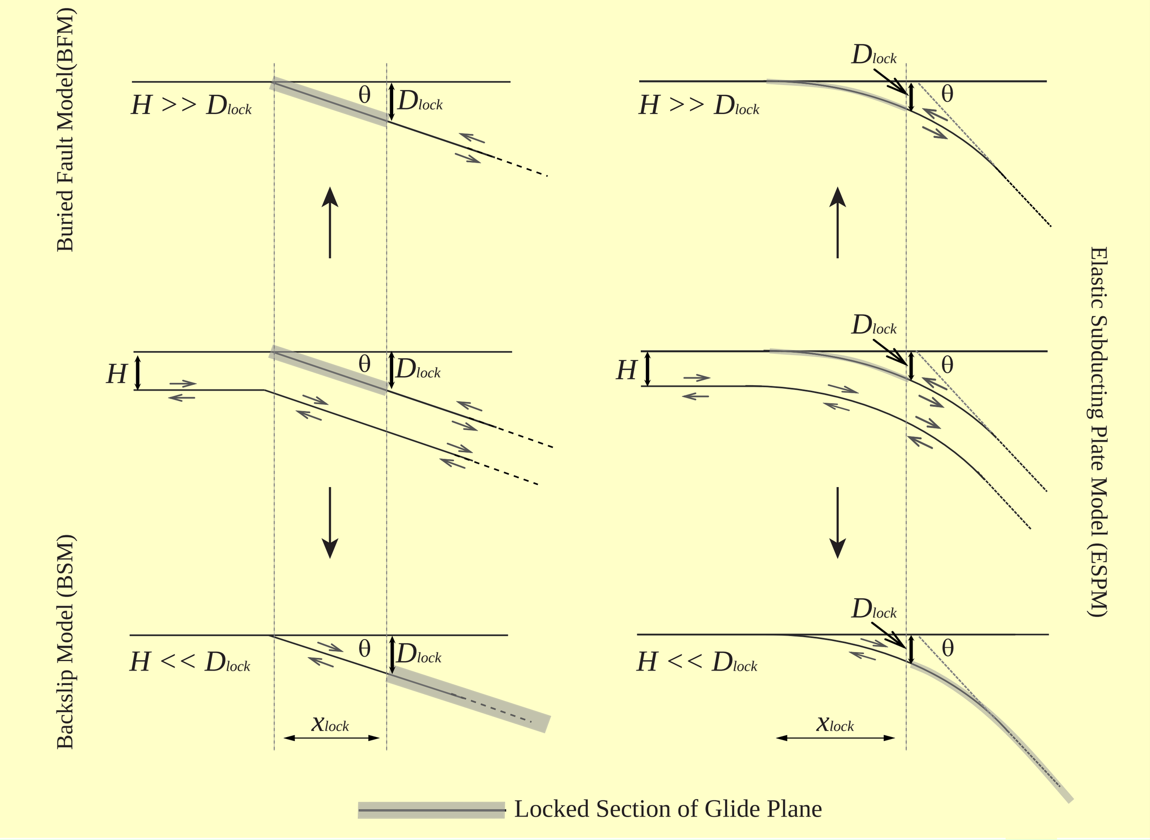
1. ESPM : An Alternate Motivation for the BSM

In this study, we aim to understand the physical rationale behind the success of the backslip model for interpreting subduction zone geodetic data (Savage, 1983) by studying a kinematically more consistent model for subduction. Specifically, we want to know under what conditions the backslip model is a good approximation for predicting surface deformation on the over-riding

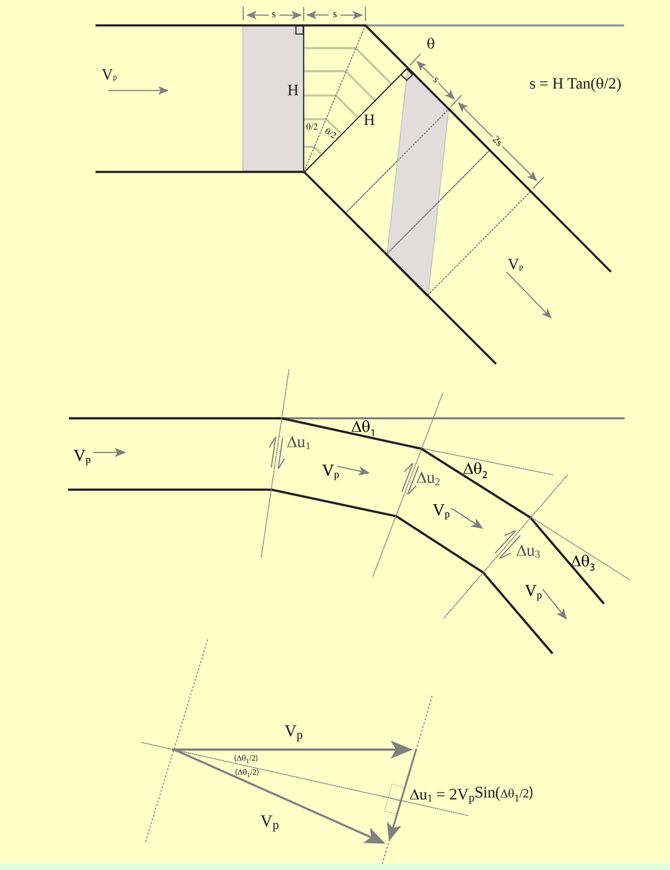


Comparison of the derivation of the BSM and the ESPM. The ESPM models a subducting plate using two edge-dislocation glide surfaces - representing the top and bottom surfaces of the subducting plate - that have slip vectors of the same magnitude but opposite sense. The bottom surface decouples the free surface from the rest of the half-space. The ESPM is kinematically more consistent in that over geologic time-scales, the subducting and over-riding plates have the correct sense of relative motion, and the ESPM does not predict net block-uplift of the over-riding plate. The right panels show that the BSM can also be derived from the ESPM by subtracting steady-state plate bending at the trench. The nearest geodetic observations from the trench, x_{GPS}, are typically located landward of x_{lock}, the surface projection of the downdip end of the locked zone.

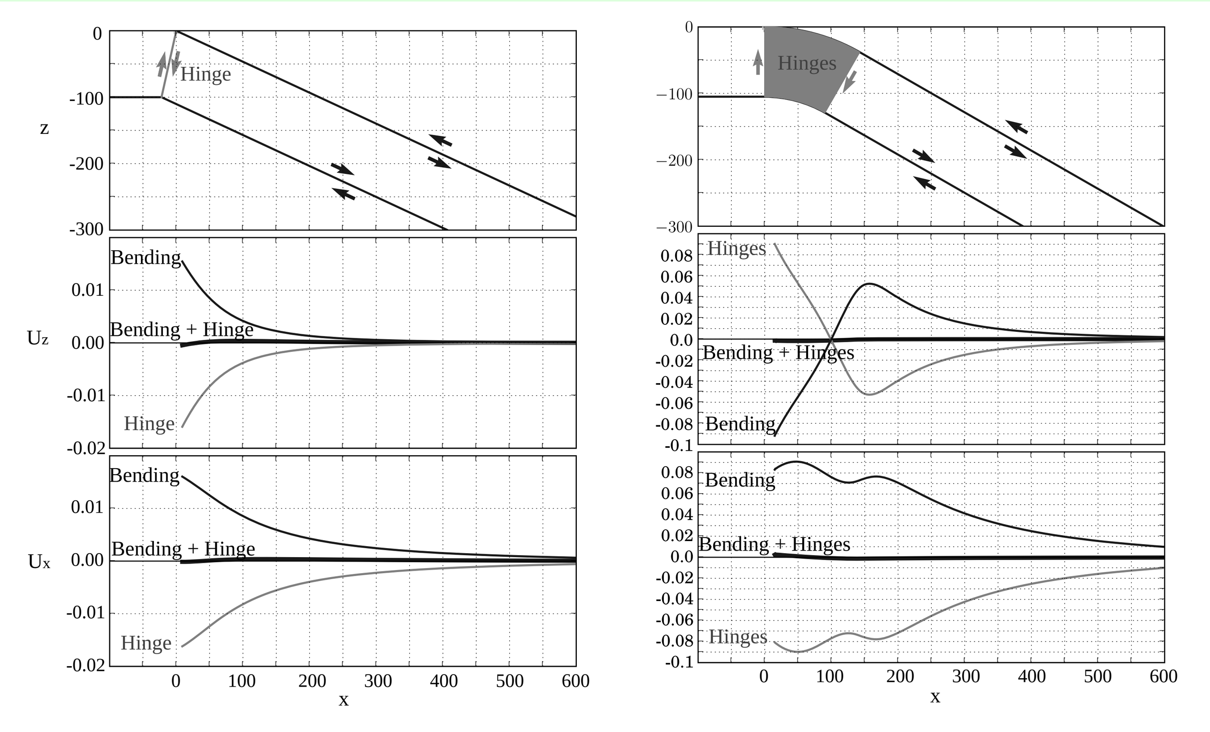


Both the BSM and the buried fault model (BFM) are end-member models of the ESPM. When plate thickness, $H = 0$, the slip vectors along the two glide surfaces representing the top and bottom of the plate cancel each other, resulting in normal-slip along the locked interface - thus retrieving the BSM. Therefore, when the BSM is used, backslip must be applied to the megathrust interface - whose shape is identical to that of the bottom plate surface directly beneath it - independent of the fault interface geometry. When H = *infinity*, the ESPM reduces to the BFM.

2. Principal Difference between the ESPM and the BSM: Plate Bending

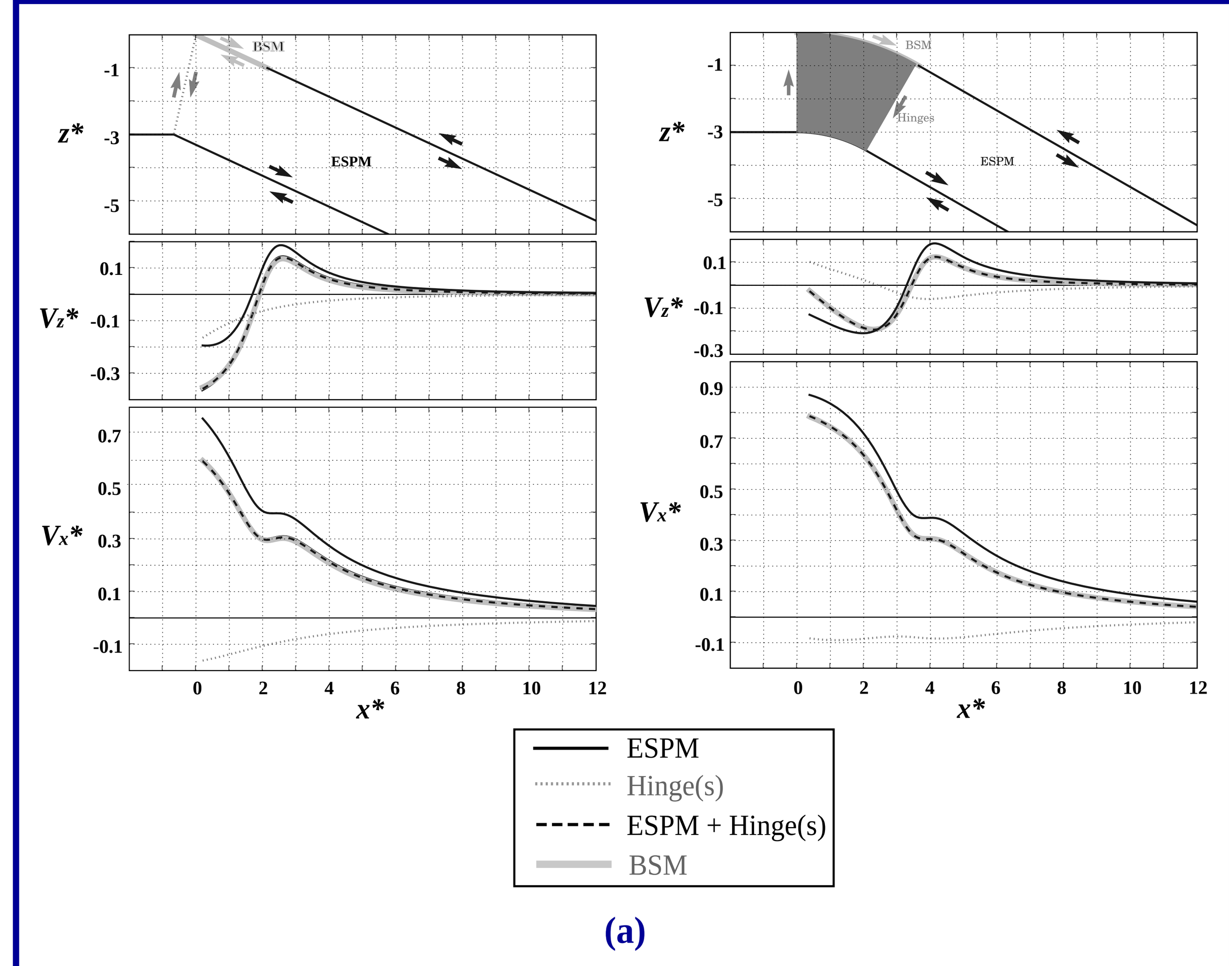


Slip along the axial hinge(s) (thin gray curves) EXACTLY compensate(s) for the effects of bending (thin black curves) - provided both processes operate over the same time period - leading to zero net deformation (thick black lines) in the frontal wedge of the over-riding plate.



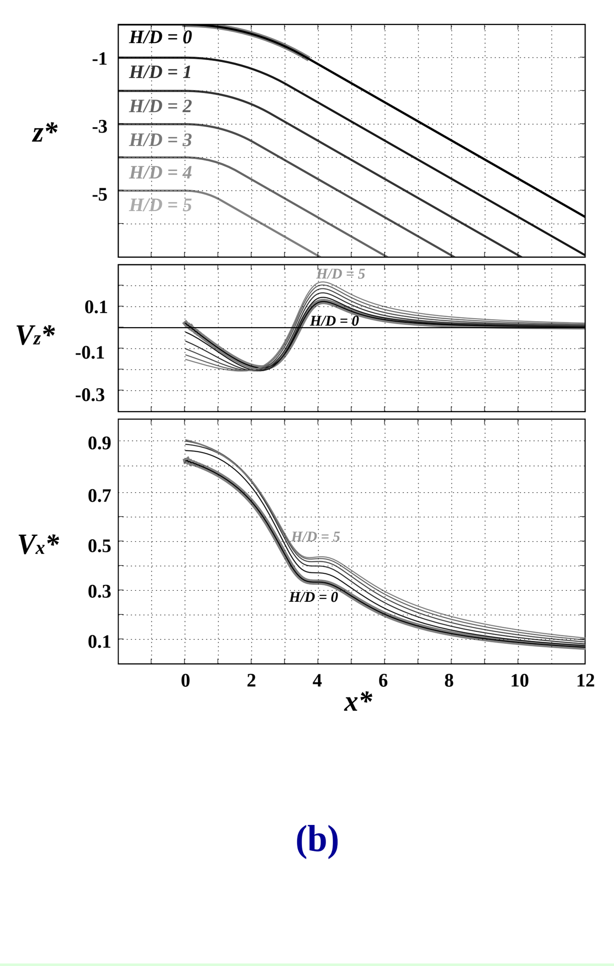
For the ESPM, our assumption of the same slip velocity for both surfaces of the plate is equivalent to flexural shear folding - plate thickness remains unchanged after bending at the trench. As subducting material passes through the trench, it undergoes simple shear (depicted by the gray areas in upper left panel), whose magnitude is proportional to the dip (linear fault geometry) or curvature (curved geometry) of the plate interface. Because of the change in direction of subducting material at the trench, the overriding plate in the ESPM experiences compression adjacent to the trench, leading to surface uplift (Box 1, and middle panels of above plots). For the subducting plate to not experience this permanent bending strain, a process equivalent to slip along axial hinge planes (equivalent to plastic flow) can be invoked, with a slip-rate, Δu_i, computed from vector addition (left-bottom panel). For a curved interface, slip is distributed over a series of such planes (gray band, right-upper panel), with slip rate proportional to local curvature. An equivalent ("viscous") method to compensate for bending strain is to introduce a uniform velocity gradient within the plate cross-section - speeding up the upper surface, while slowing down the lower surface. If all of the bending stresses associated with the above strains are released aseismically during each seismic cycle in the shallow portion of the subducion zone, or at large depths (> 100km), then no permanent bending deformation accrues in the frontal wedge - the deformation field due to the hinge(s) *EXACTLY* cancels that due to bending - and we recover the BSM from the ESPM having *ANY* plate thickness. However, if these bending stresses are even partly released episodically in the shallow portion of the subduction zone, the ESPM predicted field predicts permanent frontal-wedge deformation, with surface velocities significantly different from that of the BSM during the interseismic, within a few D_{lock} from the trench. Persistence of only a fraction of this permanent deformation in the frontal wedge can potentially explain coastal uplifts (e.g. S. Chile) or stability of islands in the forearc (e.g., Sumatra) - as can be inferred from the uplift velocities due to bending in the above plots (middle panels).

3. Comparison of the ESPM & the BSM Deformation Fields

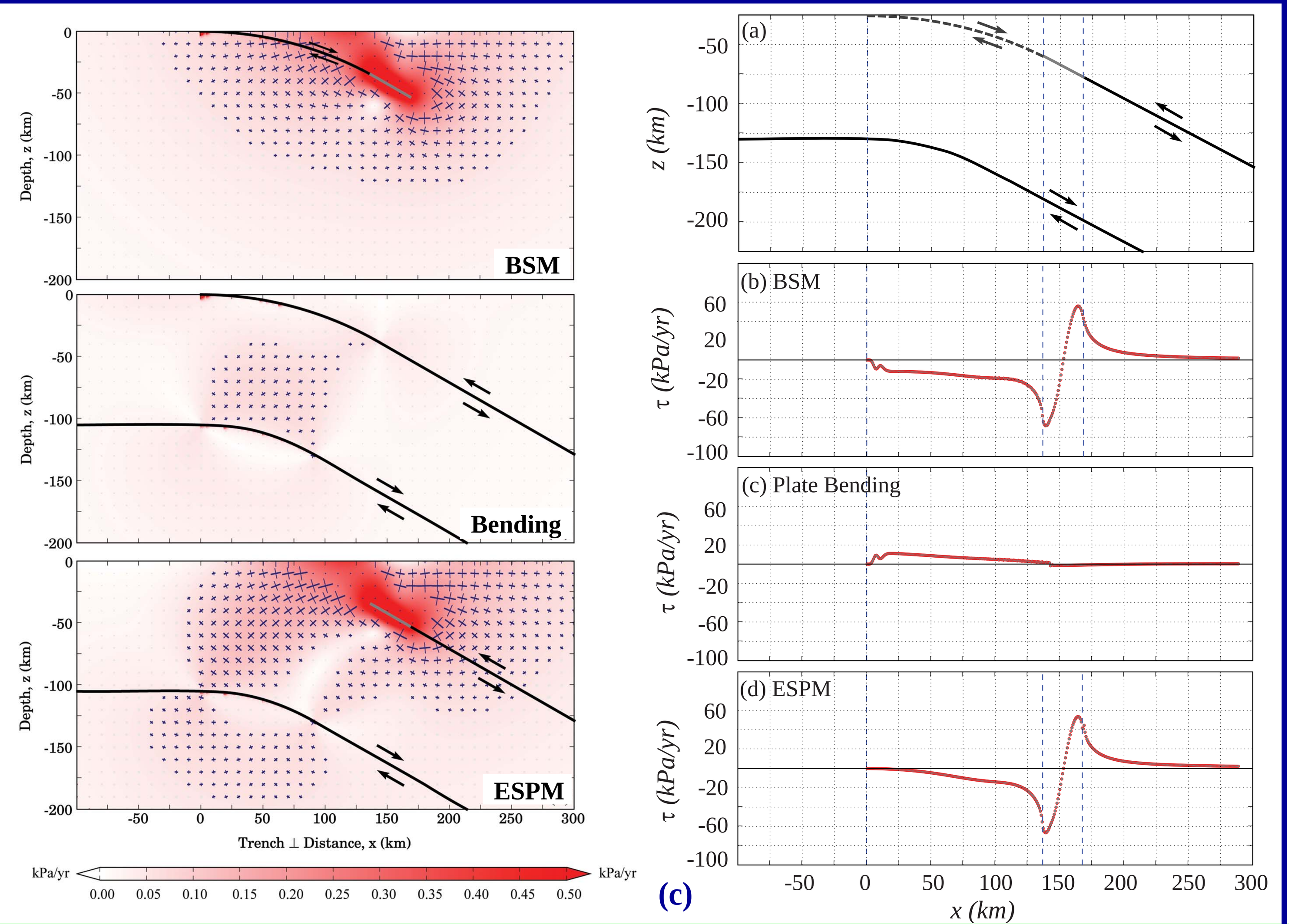


As illustrated in the top panel of Box 1, adding the deformation field due to the long-term steady state plate motion (i.e., contribution of bending, Box 2) to the BSM yields the deformation field of the ESPM. As expected, the difference between these two fields exactly cancels the field due to the hinge(s).

NOTE: There is a tradeoff between the plate thickness estimated for the ESPM from geodetic inversions and the fraction of bending stresses that are assumed to be released episodically at shallow depths (< 100 km, Box 2). Therefore, the effective elastic thickness estimated for the ESPM is a minimum thickness, contingent on the assumption that all bending stresses are released episodically at shallow depths.



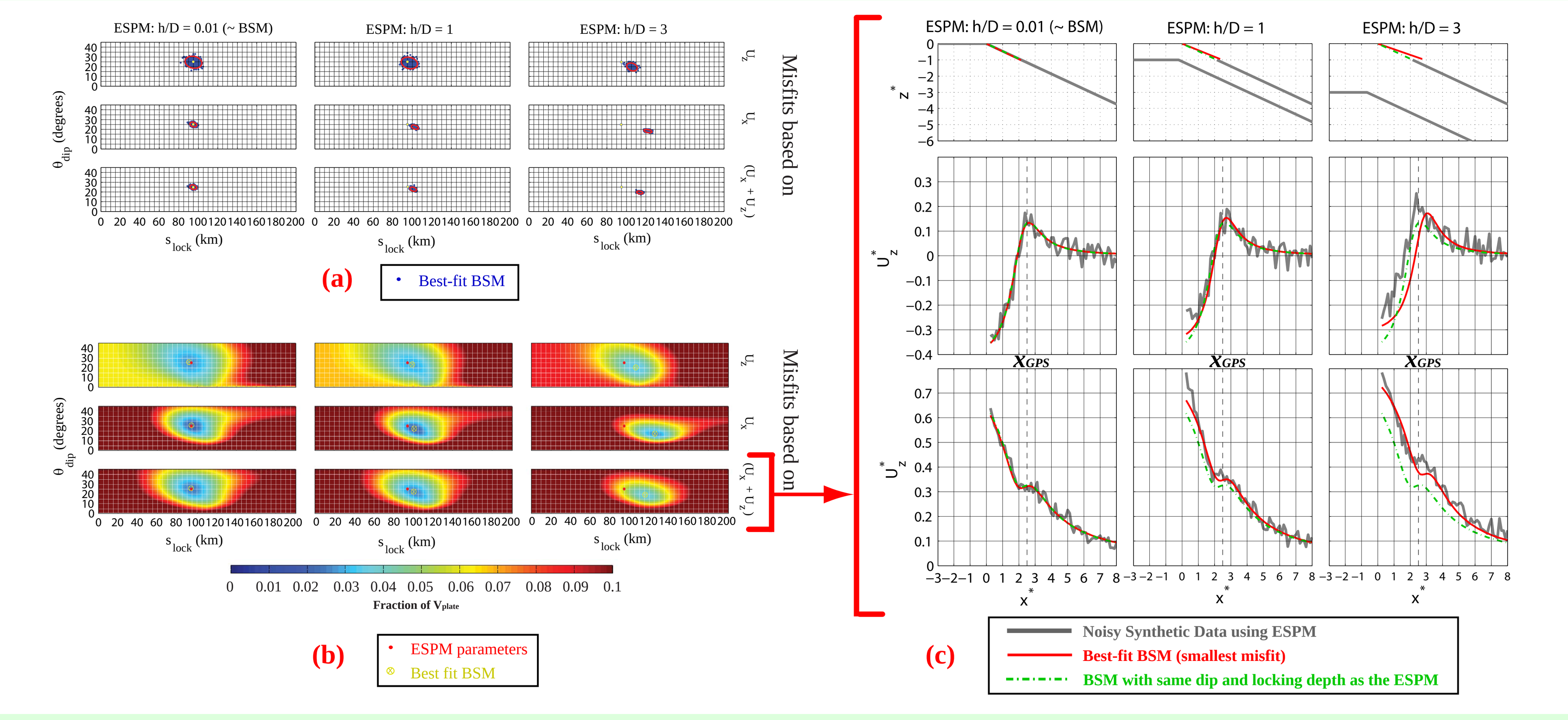
In the ESPM, a larger value for H results in larger magnitudes of vertical velocitity in the vicinity of the trench compared to the BSM. However, horizontal velocity profiles merely "pivot" upward about a point far from the trench on the over-riding plate - so horizontal strain profiles remain more or less unchanged. Therefore, vertical velocities are the key to discriminating between the ESPM and the BSM.



The left panels depict the principal deviatoric stresses superimposed on a color intensity plot of the second invariant of the deviatoric strain tensor. The principal stress directions during the bending phase are consistent with potential slip along local axial hinges of folding depicted in part (a) as well as in Box 2 above. The right panels present "deviatoric" shear tractions along the bottom surface of the over-riding plate for the two models, and for their difference - plate bending - (negative values => downdip tractions). Transition zones having a linear tapered slip distribution are used as a proxy for the effects of anelasticity near the downdip end of the locked zone.

4. Influence of Geodetic Observation Locations: Monte-Carlo Simulations

Even if bending stresses are released episodically in the shallow portions of the subduction zone, discriminating between the ESPM and the BSM would be limited by the location of the closest geodetic observations from the trench, x_{GPS}.

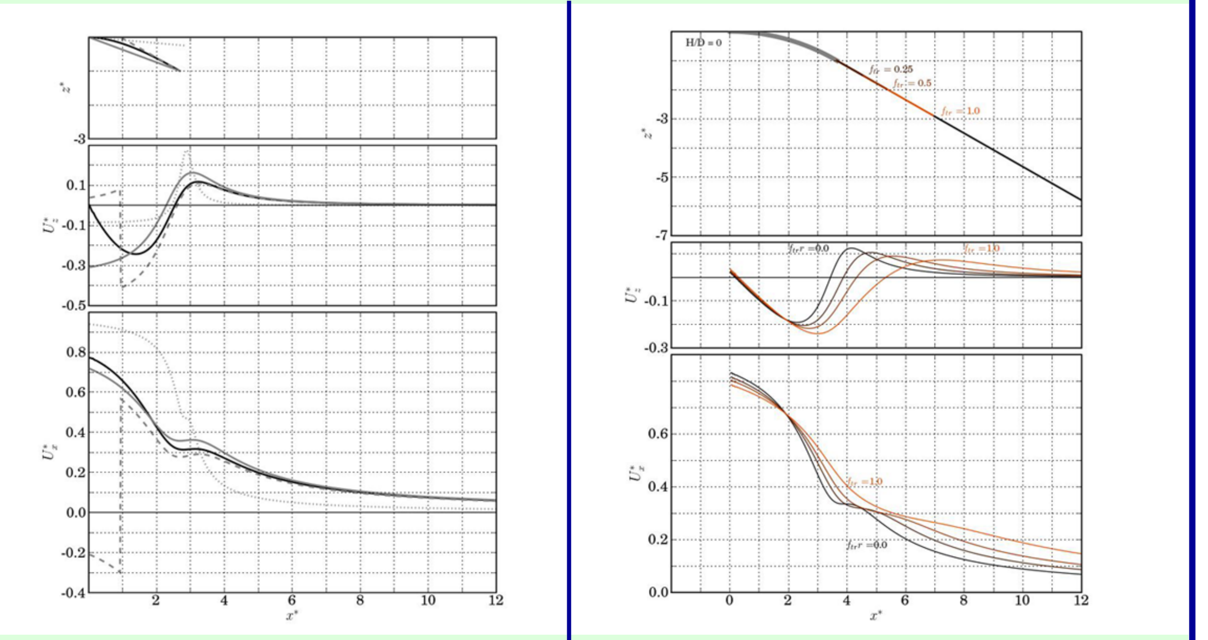


Assuming subduction zones can be represented by the ESPM, we first generate surface displacements for a specific geometry (θ = 25 deg, D_{lock} = 40 km, or s_{lock} = 95 km), for several plate thickness, h (= 0.01, 1, 3). We perturb this ESPM field to generate 1,000 noisy synthetic data-sets for each plate thickness, h - assuming Gaussian noise with zero mean and standard deviations of 2 mm/yr for the horizontals, and 3 mm/yr for the verticals. We then find the BSM with the smallest L_2-Norm misfit for each noisy data-set using grid-search. The resulting one thousand misfit-minima are presented in 9 panels of part (a) above. Color intensity plots in part (b) depict the misfit surface for the BSM, using just one of the data-sets from the corresponding panel in part (a). Each column in part (c) shows the geometry and best-fit model for the noisy data-set corresponding to the h/D_{lock} ratios presented in part (b), that fits both horizontal and vertical synthetic data (bottom row of part (b)).

As seen in Box 3 below, the ESPM predicts larger surface displacements closer to the trench than the BSM. Therefore, the BSM tends to underestimate fault dip, so that the downdip end of the locked zone is shallower, thereby increasing the magnitude of the surface displacements in the region, $x > x_{GPS}$. However, typically, fault dip is well constrained from teleseismic observations - in which case, part (b) implies that where thick slabs are subducting, the BSM will estimate a wider locked zone to account for the larger displacements for $x > x_{GPS}$.

5. Characterizing BSM Surface Observables

Given that the ESPM reduces to the BSM in all but one case, in this section, we try to build some intuitiion regarding the application of the latter model in geodetic inversions.



As illustrated schematically in the bottom panel of Box 1, when the BSM is used, backslip must be applied along the megathrust interface whose shape is identical to that of the bottom surface of the plate, directly beneath it - irrespective of the interface geometry. Some researchers have used a tangential approximation to the curved interface (dashed gray line above), which leads to significant error in surface velocity predictions right above the locked megathrust. The tangential approximation is reasonable only if we are interested in predictions beyond $3D_{lock}$ from the trench. So, for geodetic inversion of interseismic data using a kinematically consistent BSM, we should use the same fault interface as that which experiences coseismic rupture.

It is common to assume that the slip velocity changes abruptly from zero to creeping at the plate convergence rate near the downdip end of the locked megathrust interface. In reality, it is very unlikely that slip during the coseismic rupture of the megathrust (or interseismic creep below the locked megathrust) transitions so abruptly - anelastic processes would dominate near the bottom of the locked interface due to the large stresses characteristic of the region (see Box 3). Transition zones - over which slip gradually changes from zero to a finite value - provide a kinematic proxy for the integrated effects of such anelastic processses. The plausible length for such a zone could be estimated from a formal inversion using a BSM formulated to have a transition zone.

Location of surface observables like the hinge-line, X_{hinge} - where the vertical velocities change from negative to positive landward of the trench during the interseismic - and the location of the peak in the vertical velocity profile, X_{max}, provide good *apriori* constraints on the surface projection of the downdip end of the locked megathrust interface, X_{lock}. Tighter constraints on X_{lock} (from surface geodetic measurements) and fault dip (from teleseismic data) will allow us to constrainmore difficult to estimate parameters such as the width of the downdip transition zone better, especially using a formal Bayesian inversion technique. The dimensionless parameters plotted at the right are:

1) $\Delta x_h^* = (X_{hinge} - X_{lock})/X_{lock}$

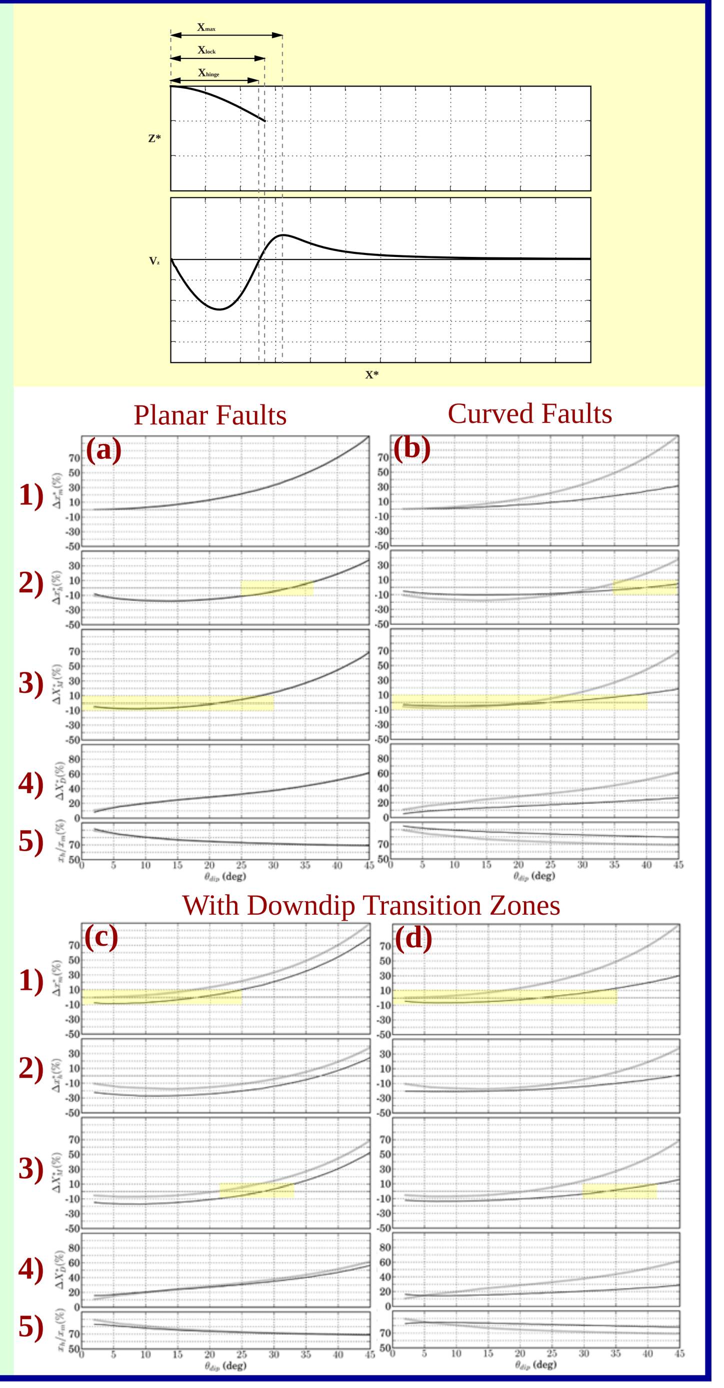
2) $\Delta x_m^* = (X_{max} - X_{lock})/X_{lock}$

3) $\Delta X_M^* = (\Delta x_m^* + \Delta x_h^*)/2$

4) $\Delta X_D^* = (\Delta x_m^* - \Delta x_h^*)$

5) x_h/x_m

When there is no downdip transition zone (a,b), the mean of X_{hinge} & X_{max} provides a very good constraint on X_{lock} (to within 10%), for most realistic plate interface dips. However, in the presence of a downdip transition zone (c,d), the effective width of locking increases, thereby making X_{max} a better estimator of X_{lock}, for dips up to 25 deg. These results can also be directly applied to the ESPM having shallow dips (< 30 deg).



6. Conclusions

(a) The ESPM provides a more intuitive rationale for the BSM. It has the correct long-term motion for the subducting plate. Unlike the BSM, it does not predict net block-uplift of the over-riding plate over geologic time scales.

(b) The ESPM is a more general model for convergent plate boundaries. The BSM (H = 0) and the BFM (H = *infinity*) are end-members of the ESPM having extreme plate thickness values.

(c) Even the ESPM with a finite plate thickness is equivalent to the BSM if all of the plate bending stresses are released continuously and aseismically in the shallow portion of the subduction zone or if they are released (episodically or continuously) at depths large enough to not influence surface deformation (> 100 km).

(d) If at least part of the bending stresses are released episodically at shallower depths, then the ESPM predicts permanent uplift in the frontal wedge (see Box 2) - which can potentially explain coastal uplift in some subduction zones, or stability of islands in the forearc.

(e) For the BSM with non-planar geometry, it is kinematically more appropriate to apply backslip on the same fault that experiences coseismic ruptures, rather than along its tangent approximation.

(f) For the ESPM or the BSM, the mean of X_{max} & X_{hinge} provide a good constraint on X_{lock} when there is no downdip transition zone. Otherwise, X_{max} is a better estimator of X_{lock}, especially for dips < 30 deg