Kinematics of shallow backthrusts in the Seattle fault zone, Washington State

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ABSTRACT

Near-surface thrust fault splays and antithetic backthrusts at the tips of major thrust fault systems can distribute slip across multiple shallow fault strands, complicating earthquake hazard analyses based on studies of surface faulting. The shallow expression of the fault strands forming the Seattle fault zone of Washington State shows the structural relationships and interactions between such fault strands. Paleoseismic studies document an ~7000 yr history of earthquakes on multiple faults within the Seattle fault zone, with some backthrusts inferred to rupture in small (M ~5.5–6.0) earthquakes at times other than during earthquakes on the main thrust faults. We interpret seismic-reflection profiles to show three main thrust faults, one of which is a blind thrust fault directly beneath downtown Seattle, and four small backthrusts within the Seattle fault zone. We then model fault slip, constrained by shallow deformation, to show that the Seattle fault forms a fault propagation fold rather than the alternatively proposed roof thrust system. Fault slip modeling shows that back-thrust ruptures driven by moderate (M ~6.5–6.7) earthquakes on the main thrust faults are consistent with the paleoseismic data. The results indicate that paleoseismic data from the back-thrust ruptures reveal the times of moderate earthquakes on the main fault system, rather than indicating smaller (M ~5.5–6.0) earthquakes involving only the backthrusts. Estimates of cumulative shortening during known Seattle fault zone earthquakes support the inference that the Seattle fault has been the major seismic hazard in the northern Cascadia forearc in the late Holocene.

INTRODUCTION

Major thrust faults can form complex systems of splay faults in the upper few kilometers that include multiple thrust fault strands and antithetic backthrusts (e.g., Mitra, 2002; Henstock et al., 2006; Giambiagi et al., 2009; Higgins et al., 2009; Neely and Erslev, 2009). Shallow backthrusts can form by slip along forelimb bedding planes (Erslev and Mayborn, 1997) or in response to a steepening of the main fault at shallow depths (e.g., Mitra, 2002), especially in the presence of layered strata (Niño et al., 1998). Such complex shallow fault splays can confuse interpretations of overall fault structure and complicate earthquake hazard analyses by creating wide zones of potential surface deformation or rupture during large earthquakes (e.g., Kelson et al., 2001; Lee et al., 2002). Studies of the rupture history of back-thrust scarps commonly are interpreted as revealing past earthquakes on the underlying main fault system (e.g., Nelson et al., 2003a); however, some small backthrusts have been interpreted to rupture independent of the main thrust fault, perhaps by storing strain during aseismic creep on the deeper faults (Kelsey et al., 2008). Understanding the relationships between shallow fault splays and the main thrust faults is thus important for interpreting paleoseismic data and understanding the slip distribution for hazard analyses. In this paper, we examine the shallow structure, deformation, and earthquake history of a broad zone of shallow fault splays in the Seattle fault zone, at the tip of the Seattle thrust fault beneath the Puget Lowland of Washington State.

The Seattle fault is a west-trending thrust fault system within the forearc of the Cascadia subduction zone, and it lies beneath some of the largest cities in the U.S. Pacific Northwest (Fig. 1). The fault system is moving in response to subduction of the Juan de Fuca plate and northward motion of the Coast Range block south of the fault (Pratt et al., 1997; Wells et al., 1998; Blakely et al., 2002; Mazzotti et al., 2002; ten Brink et al., 2002; McCaffrey et al., 2013). Potential field and seismic data show that ~10 km of slip on the Seattle fault have lifted Eocene Crescent Formation basaltic bedrock 5–7 km in its hanging wall (Daneš et al., 1965; Johnson et al., 1994; Pratt et al., 1997; ten Brink et al., 2002; Blakely et al., 2002). A M 7–7.5 earthquake on a blind, south-dipping thrust strand in A.D. 900–930 lifted wave-cut terraces along the shorelines of Puget Sound more than 6 m (Bucknam et al., 1992; ten Brink et al., 2006; Muller and Harding, 2007; Kelsey et al., 2008), caused a tsunami in Puget Sound (Atwater and Moore, 1992; Atwater, 1999; Bourgeois and Johnson, 2001), triggered landslides (Karlin and Abella, 1992; Schuster et al., 1992), and likely ruptured multiple Seattle fault zone backthrusts (Nelson et al., 2003a, 2014; Kelsey et al., 2008). These Seattle fault zone faults provide one of the best records of Holocene ruptures of a major thrust fault system in North America (Nelson et al., 2014; Sherrard and Gomberg, 2014).

Debate continues about the structure of the shallow Seattle fault zone fault strands, specifically, the relationships between the main thrust faults and the backthrusts (Kelsey et al., 2008; Nelson et al., 2014). The Seattle fault zone is an ~80-km-long (east-west), ~7-km-wide (north-south) zone of north-dipping
Figure 1. (A) Map of the Seattle fault zone (SFZ) in the central Puget Sound region (water is gray) near the Cascadia subduction zone (CSZ on index map at top right) in the U.S. Pacific Northwest. Map shows the distribution of Tertiary bedrock (red), locations of the surface projections of mapped or inferred south-dipping main thrust faults (red lines), and north-dipping backthrusts (blue lines). Red line labeled “D.F.” is the location of the synclinal axial surface (deformation front) at the surface or seafloor above the blind frontal fault, with solid lines marking where it is well located and dashed lines showing inferred locations. Green dashed lines show the location of the deformation front interpreted in Blakely et al. (2002) and the approximate limit of Seattle fault zone faulting in the south. Industry and deep marine seismic profiles acquired in Puget Sound are shown by black lines with 1 km marks shown as white dots, with a label on 10 km marks. Black lines on land in West Seattle, south Seattle, Mercer Island, and Bellevue are locations of land seismic profiles. Black triangles mark surveyed terrace elevations, with the amount of uplift or subsidence labeled (ten Brink et al., 2006). Focal mechanism shows the location of the 1997 Point White (Bremerton) M 4.9 earthquake at a depth of ~9 km. EH—Eagle Harbor; RP—Restoration Point; AP—Alki Point; BI—Blake Island; MI—Mercer Island; VI—Vashon Island; BR—Bremerton; TJH—Toe Jam Hill; VP—Vasa Park; RP—Rich Passage; WS—West Seattle; WP—Waterman Point; PG—Point Glover; IW—Island Wood; BH—Beacon Hill. Bedrock geology is from Booth (2007, written commun.) and Haugerud (2008). (B-C) Proposed thrust-fault and roof-thrust (wedge) models for the Seattle fault zone structure. The thrust-fault model is a fault-propagation fold with trishear behavior and several shallow fault splays including backthrusts (Suppe and Medwedeff, 1990; Shaw et al., 2005). The roof-thrust model has a wedge being thrust into Seattle basin strata, and backthrusts formed by bedding-plane slip caused by folding above the wedge tip (Brocher et al., 2004; Kelsey et al., 2008).
forelimb strata, south-dipping thrust faults, and north-dipping backthrusts. The north edge of the Seattle fault zone is a synclinal axial surface, termed the "deformation front," separating north-dipping forelimb strata from subhorizontal Seattle basin strata to the north (Pratt et al., 1997; Johnson et al., 1999; Brocher et al., 2004). The original "thrust-fault" model for the Seattle fault zone has south-dipping thrust faults forming a fault-propagation fold (Fig. 1B; Johnson et al., 1994, 1999; Pratt et al., 1997; Blakely et al., 2002). The subsequent discovery of south-facing scarps within the Seattle fault zone (Haugerud et al., 2003; Nelson et al., 2003a) led to recognition of north-dipping backthrusts (Fig. 1B; ten Brink et al., 2002; Liberty and Pratt, 2008) and the development of an alternative "roof-thrust" or "wedge" model in which the main fault flattens at 4–6 km depth to form a wedge structure beneath a subhorizontal detachment, forming a passive-roof duplex (Fig. 1C; Brocher et al., 2004; Kelsey et al., 2008). In this roof-thrust model, back-thrust slip along forelimb bedding planes above the wedge is driven by bending at the synclinal axial surface at the deformation front. The two models have different implications for seismic hazard estimates, including different depths and locations of the south-dipping thrust faults, different potential for surface rupture by the main thrust faults, and the presence or absence of the shallow roof thrust.

In either kinematic model, the Seattle fault zone backthrusts pose a conundrum because back-thrust ruptures appear to have occurred at times other than during the large A.D. 900–930 earthquake (Nelson et al., 2003a, 2003b, 2014; Kelsey et al., 2008). These back-thrust ruptures form localized (~400–500-m-wide) uplifts of the wave-cut terraces along the shores of Puget Sound, without an obvious longer-wavelength terrace uplift caused by slip on the deeper, more-extensive main thrust faults (Kelsey et al., 2008). This terrace morphology led to the hypothesis that aseismic creep on the main thrust faults causes the small backthrusts to store enough strain to rupture alone in small to moderate earthquakes (M 5.6–6.0; Kelsey et al., 2008). In this model, the lack of a regional terrace indicates that erosion kept pace with the gradual uplift caused by the aseismic creep, and the model implies that back-thrust earthquakes do not always reveal times of larger earthquakes on the deeper, main fault system, as has been interpreted in the past (Nelson et al., 2003a). The aseismic creep inferred in this hypothesis also implies that earthquakes on the main thrust faults are less frequent than might be inferred by assuming that the accumulated displacement occurred seismically.

Here, we assess the proposed Seattle fault zone main thrust–backthrust geometries and interactions by examining and modeling the shallow deformation near the fault tip. We first identify and characterize the shallow faults and folds within the Seattle fault zone using seismic-reflection data. These data include new high-resolution land profiles and legacy high-resolution marine profiles that were last interpreted before the backthrusts were discovered. We then use slip modeling, constrained by the morphologies of the wave-cut terraces and the geometry of the fault splays, to test the plausibility of proposed fault geometries, to examine potential relationships between the thrusts and backthrusts, and to estimate the magnitudes of earthquakes identified in the paleoseismic record. The results show the interaction of shallow fault splays in a major thrust-fault system, with implications for earthquake hazards in the U.S. Pacific Northwest and for interpreting the earthquake record on shallow backthrusts in general.

### DATA AND METHODS

We use seismic-reflection profiles to characterize shallow structures within the Seattle fault zone. The profiles include marine industry and deep profiles acquired in Puget Sound (Johnson et al., 1994; Pratt et al., 1997; ten Brink et al., 2002; Fisher et al., 2006), marine high-resolution profiles from Puget Sound and Lake Washington (Harding et al., 1988; Johnson et al., 1999), and land seismic profiles (Stephenson et al., 2007). The marine profiles were collected and originally interpreted before recognition of the backthrusts, and most of these profiles have not subsequently been examined to see whether they image the splay faults.

The seismic-reflection profiles are time sections, with the exception of the migrated and depth-converted profiles collected as part of the Seismic Hazards Investigations of Puget Sound (SHIPS) project (Fisher et al., 2006). The high-resolution marine profiles used an extremely short receiver array that provided little or no velocity information, and the original processing therefore used a single stacking velocity. For this paper, we frequency-wavenumber (f-k) migrated all of the time sections and then depth-converted them using a velocity function estimated from the industry stacking velocities. This velocity function ranged from 1480 m/s in the water column to 2200 m/s at 0.8 s travel-time and 4000 m/s at 3 s travelt ime.

To compute the shallow deformation during earthquakes for different fault geometries, we used an elastic boundary-element code (Gomberg and Ellis, 1994). We defined the fault planes as stress-free surfaces under the assumption that faults are inherently weak and therefore relieve long-term strain (e.g., Gomberg and Ellis, 1994). Fault planes were divided into multiple small fault segments to allow decreasing slip near fault tips. After the faults were defined, we simulated compression of the model space to force the faults to slip freely, without specifying the amount of slip. Allowing the faults to slip unconstrained during compression creates a more realistic partitioning of slip between faults than is achieved by explicitly assigning the amount of slip on each fault.

To constrain the kinematic models, we used the elevations of the wave-cut terraces along the shores of Puget Sound under the assumption that the terraces represent a former sea-level datum that has been uplifted tectonically. The amount of uplift can be determined by measuring the elevation difference between the uplifted notches or inflection points at the inland edges of...
the terraces and the equivalent modern notch near sea level. This measurement generally requires excavating the upper portions of the terraces from beneath debris flows and colluvium (ten Brink et al., 2006; Muller and Harding, 2007). Therefore, the hand-surveyed points given in ten Brink et al. (2006) are likely more accurate than elevations derived from light detection and ranging (LiDAR) imaging of the terraces (e.g., Muller and Harding, 2007), because the LiDAR picks may not be at the slope break at the upper edge of the terrace or may be on debris resting on the terrace. The lower points in the LiDAR-derived elevation profiles matched the hand-surveyed elevations, suggesting that these lower points coincide with the bedrock surface and that the higher points are caused by more resistant layers in the platforms or by debris on the inland part of the terrace.

It is important to note that the elevations of the wave-cut terraces represent the cumulative uplift from all earthquakes since the terrace formed minus any decrease in elevation caused by erosion. We cannot separate the uplift contributions from individual earthquakes, so we modeled the cumulative uplift from all earthquakes on both the main fault and the backthrusts. Although overwhelmingly dominated by the uplift in the A.D. 900–930 event, the terrace morphology also reflects elevation changes during the various backthrust earthquakes that occurred at other times (Nelson et al., 2003a; Kelsey et al., 2008).

### STRATIGRAPHY

The shallowest strata imaged by the seismic profiles are as much as 300 m of late Pleistocene to Holocene unconsolidated strata, which lie above a prominent unconformity at the top of as much as 700 m of hummocky Quaternary glacial deposits (Fig. 2; Easterbrook, 1986; Booth, 1994; Porter and Swanson, 1998; Johnson et al., 1999; Booth et al., 2003, 2005; Troost and Booth, 2008). The base of the Quaternary deposits, and top of consolidated Tertiary strata, is interpreted at the downward change to more reflective and more-continuous strata (Johnson et al., 1999). Shoreline exposures within the Seattle fault zone (Fuller, 1975; McLean, 1977; Haugerud, 2005; Kelsey et al., 2008) and borings north of the Seattle fault zone show the Tertiary strata to be Miocene Blakely Harbor Formation, Oligocene Blakeley Formation, and Eocene marine strata above Crescent Formation basaltic bedrock (Johnson et al., 1994, 1999; ten Brink et al., 2002; Haugerud and Troost, 2011).

It is tempting to interpret the unconformity at the base of the unconsolidated strata as being an erosional surface formed during the latest, Vashon glaciation.
17,600–16,500 cal. yr B.P. (Porter and Swanson, 1998; Booth et al., 2003, 2005), with the overlying deposits being postglacial. The 12–15 mm/yr modern sedimentation rate within central Puget Sound (Lavelle et al., 1985) is consistent with 200–250 m of sediment accumulating since glacial retreat ~16,500 yr ago. The rate was likely not constant, however, as a large influx of sediments is documented in some glacial fjords immediately following glacial retreat (e.g., Cowan et al., 2010), and indeed Troost (2011) determined a 16,500 and 7600 cal. yr B.P. sedimentation rate of 25 mm/yr in Lake Washington that slowed to 1.3 mm/yr in the late Holocene. The ~300 m of post-unconformity sediments beneath central Puget Sound therefore could be entirely Vashon recessional and later deposits. However, as described later in more detail, the amount of deformation of these post-unconformity strata appears to be too large to have occurred since the last glaciation, suggesting that the lower parts of the subhorizontal strata above the prominent unconformity are Olympia-aged interglacial strata (60–176 cal. k.y. B.P.; Johnson et al., 1999; Booth et al., 2003, 2005) or older strata. Exposures on land show that Vashon deposits can accumulate on Olympia interglacial beds without a prominent unconformity between them, as in exposures in northwest Seattle (e.g., Booth et al., 2003, 2005). The unconformity may thus predate the last, Vashon, glaciation.

### SHALLOW SEATTLE FAULT ZONE THRUSTS AND BACKTHRUSTS

The seismic profiles across the Seattle fault zone show disrupted, north-dipping forelimb strata between the subhorizontal strata of the Seattle basin to the north and the folded but continuous strata of the Seattle uplift to the south (Figs. 2 and 3). On Bainbridge Island, along the west shore of Puget Sound, Tertiary forelimb strata dip 35° to 90° northward and in some places are overturned (Fulmer, 1975; McLean, 1977; Haugerud, 2005). Beneath Puget Sound, the deformation front is a prominent synclinal axial surface that sharply folds strata at ~1 km depth. Forelimb strata south of the deformation front are broken into several coherent blocks separated by south-dipping faults that lifted Crescent Formation rocks now at ~7.5 depth beneath the Seattle basin to 1–2 km depth in the hanging wall south of the Seattle fault zone (Johnson et al., 1994; Pratt et al., 1997; Brocher et al., 2001; ten Brink et al., 2002). Potential field modeling and tomographic data are consistent with thrusting of Crescent Formation rocks over younger sediments on a fault or faults dipping 35° to 50° south (Pratt et al., 1997; Blakely et al., 2002; ten Brink et al., 2002). Miocene and younger subsidence in the south part of the Seattle basin causes a southward thickening of reflectors within the basin strata (Johnson et al., 1994; Pratt et al., 1997; ten Brink et al., 2002).

#### South-Dipping Thrust Faults

The presence or absence of south-dipping thrust faults in the shallow portions of the Seattle fault zone is a key difference between the thrust-fault and roof-thrust models, with the roof-thrust model not having any south-dipping faults within the upper 4–6 km depth range (Figs. 1B and 1C). There are four publicly available seismic profiles that image the Seattle fault zone in the upper ~5 km, and two proprietary industry profiles. These profiles have variable quality, and uncertain velocity models and three-dimensional effects complicate migration and depth conversion of the data. The profiles...
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possibility is that fault A ruptured to the surface in the Quaternary but was strata into a gentle synclinal axial surface with uplift to the south (Fig. 7). One displacement (Fig. 6). Slip on fault A folds the well-imaged post-unconformity strata immediately below the deformation front, with ~83 m of apparent strata above the unconformity show only subtle deformation. Previously un- ing ~83 m of fault motion aligns reflectors below the unconformity (arrows with the presence of fault A (between the red arrows on Fig. 5A), and revers- and Quaternary strata downdip of the deformation front that are consistent with the interpretation of strong, continuous reflectors apparent on profile PS-1 (Fig. 3) and Prat- strata at depths of ~1 km. The location of the deformation front was mapped at the surface above the north-most location where this folding is clearly evident (black triangles in Figs. 2, 3, 5, and 6), but projecting the axial surface upward along the same dip as in the deeper strata places the deformation front at the seafloor 1–1.5 km farther north (e.g., Figs. 1–7; interpreted locations are shown in Fig. 4 for comparison). This discrepancy is apparent in Fisher et al. (2006; their fig. 6), who showed shallow folding beneath central Puget Sound ~1 km north of where the deformation front had been interpreted.

Recent Deformation and Backthrusts within the Seattle Fault Zone

Profile P42 images well the faults and folds deforming subhorizontal, post-unconformity strata beneath central Puget Sound (Fig. 7). Folding is gentle at the deformation front, with hanging-wall strata showing disruption of a prominent reflector ~100 m above the unconformity. This folding involves even the shallowest strata near the seafloor. The synclinal axial surface projects to the seafloor almost directly east of Eagle Harbor on Bainbridge Island (Fig. 4). During the A.D. 900–930 earthquake, there was uplift of the south shore and subsidence of the north shore of Eagle Harbor (Fig. 4; Bucknam et al., 1992; ten Brink et al., 2006), which confirmed that the blind thrust fault beneath the deformation front had slipped during that earthquake.
Figure 4. (A) Map showing the surface projections of interpreted faults and folds, and the locations of high-resolution marine reflection-seismic profiles. Red lines are the locations of south-dipping thrust faults at the surface or seafloor; blue lines are locations of north-dipping backthrusts; dashed green lines are the location of the deformation front and south edge of Seattle fault zone in Blakely et al. (2002). The black lines denote the seismic profiles, with dots showing the kilometer marks on the profiles displayed in this paper. Other symbols and annotations are the same as in Figure 1. (B) Magnetic anomaly map of the Seattle fault zone from Blakely et al. (1999) with the same interpretation and seismic profiles shown in part A of the figure.
Figure 5. (A) Profile P346–348 from central Puget Sound. Red arrows show zone of fault A reflector disruptions and terminations. Faults B and C are in an area with few reflectors and therefore are not evident. (B) Interpreted seismic profile. Solid red line marks fault A, and red dashed lines are axial surfaces of folds and the projected locations of faults B and C. Dark-blue lines are backthrusts interpreted from deformation of shallow strata or apparent displacements of the water bottom. Yellow lines are reflectors within the Pleistocene and Holocene strata. Green lines denote the unconformity at the base of Pleistocene deposits. Turquoise lines mark prominent reflectors in the Quaternary and Tertiary strata. Black dots show the estimated boundary between Tertiary and Quaternary strata (Johnson et al., 1999). Black triangle is the location of the deformation front in Blakely et al. (2002); m—multiples. (C) Seismic profile cut along fault A and ~83 m of slip removed to show the alignment of prominent reflectors (red arrows) after reversing the fault motion.

Figure 6. (A) Profile P42 from central Puget Sound. Red arrows mark the folded and truncated reflectors at fault A. (B) Interpreted seismic profile. Lines and symbols are the same as in Figure 5. (C) Seismic profile cut along fault A with ~83 m of slip removed, showing the alignment of prominent Quaternary and Tertiary reflectors (red arrows) below the unconformity.
Profile P42 shows a sharp synclinal axial surface with well-defined limbs in the shallow strata at the updip projection of fault B (Figs. 6 and 7). This folding appears to postdate deposition of the lower post-unconformity strata, which retain their thickness across the fold rather than showing onlap. Fault C cuts shallow strata at ~400 m depth with little vertical change, implying it had only minor late Pleistocene and Holocene dip slip.

Six backthrusts within the Seattle fault zone form south-facing fault scarps (Fig. 4; Haugerud et al., 2003; Nelson et al., 2003a, 2003b, 2014; Kelsey et al., 2008; Sherrod and Gomberg, 2014). The mapped backthrusts are the Island-Wood (“Macs Pond fault” in Haugerud, 2005), Toe Jam Hill/Point Glover, Waterman Point, and West Seattle. There also is evidence from a change in elevation of a wave-cut terrace for an unnamed scarp south of the West Seattle scarp (backthrust “e” in Fig. 4), and another backthrust not evident on land can be identified from the high-resolution seismic profiles (backthrust “c” in Fig. 4; also see Sherrod and Gomberg, 2014). The displacements of late Pleistocene and Holocene strata are near or below the resolution of the industry and deep data, so the backthrusts in our interpretations of these data (Figs. 2 and 3) are based on their seafloor locations and the assumption that they sole into forelimb bedding planes.

The high-resolution seismic profiles (Johnson et al., 1999), which were collected and interpreted before the backthrusts were recognized (Haugerud et al., 2003; Nelson et al., 2003a), show that some of the north-dipping backthrusts disrupt reflectors and cause changes in reflector dip. The Toe Jam Hill backthrust (Nelson et al., 2003a) forms the southern edge of the disrupted hanging-wall strata above fault A (backthrust “b” in Figs. 4, 6, and 7). Backthrust “d” forms a scarp on the uplifted terrace in West Seattle (Kelsey et al., 2008), cuts a prominent reflector at ~400 m depth (Figs. 5 and 7), may disrupt strata near the seafloor (Fig. 5), and clearly displaces strata up to the north at ~300–350 m depth (Fig. 7). Backthrust “e” (Fig. 4) coincides with an elevation change in the uplifted terrace in West Seattle and deforms strata and/or the water bottom on several profiles (e.g., Fig. 5). Another, previously unmapped, backthrust that extends beneath Puget Sound from the south edge of Alki Point cuts a prominent reflector at ~420 m depth and appears to displace the water bottom (“c” in Fig. 7; Sherrod and Gomberg, 2014). This scarp is not identified on land, perhaps because the scarp was eroded in the wave zone before uplift during the A.D. 900–930 earthquake, but it extends toward the south edge of Tertiary Blakeley Formation strata at Alki Point (Troost et al., 2005). The Island-
Wood fault (‘a’ in Fig. 4; Kelsey et al., 2008) is not obvious on the seismic profiles, probably because it is too shallow to extend beneath the deeper parts of Puget Sound.

To summarize, the seismic-reflection profiles and geologic data show the Seattle fault zone near Puget Sound to be an ~7-km-wide (north-south) zone with three south-dipping thrust faults and at least six backthrusts deforming the shallow strata beneath and adjacent to Puget Sound. All of these faults deform the strata above a prominent unconformity or reach the surface, consistent with late Pleistocene to Holocene movement.

**Amount of Post-Unconformity Fault Slip**

Profile P42 shows that the synclinal axial surface at the deformation front lifts a prominent, likely late Pleistocene to Holocene reflector ~24 m in a fold extending ~650 m south of the deformation front (Fig. 7). The prominent reflector likely was deposited in a nearly horizontal attitude, as the lower strata above the unconformity appear to be nearly parallel across axial surface B on profile P42, without onlap or lateral thinning indicative of deposition on a slope (Fig. 7). Strata in front of both advancing and retreating glaciers are known to profile P42, without onlap or lateral thinning indicative of deposition on a slope (Fig. 7). The prominent reflector is consistently within a nearly horizontal attitude, as the lower strata extending ~650 m south of the deformation front (Fig. 7). The prominent reflector is consistent with late Pleistocene to Holocene movement.

The post-unconformity deformation imaged on profile P42 is consistent with folding above shallow thrust faults. Modeling of ~37 m of slip on a 40°S-dipping thrust fault (fault A) terminating ~400 m below the unconformity (~850 m depth) reproduces the shape and the ~24 m of uplift of the prominent post-unconformity reflector at the deformation front (Fig. 7C). This suggests that the 6–8.5 m of terrace uplift in the A.D. 900–930 earthquake (ten Brink et al., 2006) represents ~25% of the total motion along fault A since deposition of the prominent reflector.

There is greater post-unconformity uplift above thrust fault B, suggesting it has been more active than fault A in the late Pleistocene and Holocene. Fault B forms a well-defined synclinal axial surface with ~48 m of structural relief on the prominent reflector distributed across a forelimb ~1.25 km wide. Modeling slip on a 40°S-dipping thrust fault with a tip ~650 m below the prominent reflector (~1.1 km depth) reproduces the shape and amplitude of the uplift above fault B (Fig. 7C). This similarity in shape suggests that the post-unconformity uplift is due to ~75 m of slip on fault B. Fault C shows almost no post-unconformity uplift, suggesting that Pleistocene and younger dip slip has been concentrated on faults A and B or that fault C recently has predominantly strike slip.

The deformation of the prominent reflector at the base of the post-unconformity strata is too large to plausibly occur since glacial retreat, implying the strata above the unconformity predate the last glaciation. If the post-unconformity strata are recessional and postglacial in age (younger than 16.5 ka, Porter and Swanson, 1998), the deformation of the prominent post-unconformity reflector requires a rate of horizontal shortening of ~5.2 mm/yr. However, the documented ~5 mm/yr total rate of modern contraction for the entire Washington and Oregon forearc (Wells et al., 1998; Mazzotti et al., 2002; McCaffrey et al., 2013) makes ~5.2 mm/yr of shortening on the Seattle fault zone implausible. Similarly, a total of ~12 large earthquakes, each causing ~6 m of uplift in the Seattle fault zone, would be required to reproduce the apparent folding above faults A and B, requiring a recurrence interval of only ~1375 yr. Such a short interval contradicts paleoseismic evidence (Sherrod et al., 2000; Nelson et al., 2014) and is more than the total postglacial uplift estimated across the Seattle fault zone (Thorson, 1993).

The most straightforward explanation for the large amount of deformation of the post-unconformity strata is that the prominent reflector imaged on profile P42 is older than 16.5 ka. If we infer that the reflector has an age of ca. 60 ka, the rate of horizontal contraction across the Seattle fault zone would be ~2 mm/yr, and the recurrence interval for large earthquakes would be ~5000 yr, both of which can be reconciled with paleoseismic and geodetic evidence. This age for the lower post-unconformity strata would require that the latest, Vashon glacial advance did not leave an obvious unconformity, and indeed at some locations, only a disconformity is observed between the pre-Vashon, Olympia-aged beds and the overlying strata (e.g., base of the Lawton Clay in Seattle; Booth et al., 2003, 2005; Troost and Booth, 2008). Given this uncertainty in the age of the deformed strata, we cannot yet determine a slip rate based on the apparent uplifts above the Seattle fault strands.

**Location of the Deformation Front beneath Seattle**

High-resolution seismic profiles indicate that the deformation front lies beneath the downtown area of Seattle, north of where previous interpretations have placed it (Fig. 4 shows both interpretations). The synclinal axial surface can be traced unambiguously beneath Puget Sound with a linear trend to a point ~4 km north of Alki Point at the entrance to Elliott Bay (Fig. 4). No north-south horizontal offset of the deformation front is apparent due to north-south strike-slip faults beneath Puget Sound, as previously hypothesized (Johnson et al., 1999), and potential field anomalies extend across Puget Sound without any abrupt offsets (Fig. 4; fig. 2A in Blakely et al., 2002). At the mouth of Elliott Bay, we interpret the synclinal axial surface on profiles P53, P31, and P55 as a transition from predominantly north-dipping, discontinuous reflector segments on the south parts of the profiles to predominantly subhorizontal reflector segments to the north (Fig. 8).

East of Seattle, seismic profiles from Lake Washington show the synclinal axial surface projecting to a location just north of Mercer Island (Figs. 4 and 9). Composite marine profile P6–8 south of the Interstate 90 bridge shows strata dipping ~20° northward near its north end and steepening down section to ~50° beneath kilometer 8 (Fig. 9B). The synclinal axial surface at the deformation front must lie north of these dipping strata, and thrust fault A therefore must lie below the seismic profile. The lack of folding on profile P14 north of the bridge requires the deformation front to be near the south end of that profile. These observations constrain the deformation front to project to the lake bottom near the north edge of Mercer Island. Recently acquired land seismic-reflection profiles near the north end of Mercer Island likewise show...
Tertiary forelimb strata dipping ~45° north to depths of 1.5 km or more (e.g., Fig. 9C; Stephenson et al., 2007), confirming that the synclinal axial surface projects to the lake bottom near the north end of Mercer island.

Recently acquired high-resolution land seismic-reflection profiles south-east of downtown Seattle show that the deformation front must lie at or north of the location interpreted in Blakely et al. (2002). The north-dipping Tertiary strata imaged on the northernmost profile collected on Beacon Hill (Fig. 10) require the axial surface, if it dips ~40° south, to project to the land surface at least 700 m north of the profile. This places its southernmost possible location to be slightly north of the location interpreted in Blakely et al. (2002), but we do not know how much farther north it lies.

The deformation front has a nearly linear trend beneath and west of Puget Sound, and for ~18 km beneath and east of Lake Washington, it follows a similar trend (Fig. 4; Liberty and Pratt, 2008); extrapolation of the deformation front...
Figure 9. Seismic-reflection profiles from the Lake Washington area. (A) Seismic-reflection profile PG3 from Lake Washington (Harding et al., 1988) showing the synclinal axial surface (red dashed line). Black dots indicate estimated boundary between Tertiary (T) and Quaternary (Q) strata based on Johnson et al. (1999). Note that the dip of the strata steepens with increasing depth (turquoise lines), with modestly dipping Quaternary strata (Q) onlapping the more steeply dipping Tertiary strata (T). (B) High-resolution seismic profiles P6–8 (left) and P14 (right) from Lake Washington, which form a continuous profile except for a gap at the Interstate 90 bridge. The red dashed lines and red arrow near the gap show the potential range of locations for the axial surface at the deformation front assuming an ~40° dip. The appearance of slight flattening of the reflectors at the north end of profile B is most likely an artifact caused by the migration process creating “smiles” on the edge of the seismic profile. (C) Land seismic profile collected along 78th street near the north end of Mercer Island (see Fig. 4 for location) showing that the synclinal axial surface (red dashed line) must project to the surface near the north end of the island or farther north. Note that the dip of the Quaternary strata steepens with increasing depth, indicating fold growth during deposition.
between these locations along the same trend places it directly beneath downtown Seattle (Fig. 4). The profiles within Elliott Bay just west of downtown Seattle do not show a consistent pattern of deformation (Fig. 11), in part because a northwest-trending glacial channel as much as 500 m deep contains late Pleistocene deposits (Troost et al., 2005) that have only subtle deformation. The profiles do not extend far enough south to confidently interpret the axial surface above fault A within the underlying Tertiary reflectors.

Several lines of geomorphic and geologic evidence are consistent with the deformation front lying beneath downtown Seattle. First, the deformation front at the inferred location beneath Elliott Bay projects to the seafloor near a submarine channel (Fig. 12A), perhaps caused by erosion of weakened strata within the synclinal axial surface. Such west-trending channels are anomalous beneath Puget Sound, occurring primarily in fault zones.

Second, a borehole cross section along the transit (light rail) tunnel beneath downtown Seattle (Shannon and Wilson, 1986) shows two features consistent with folding and uplift of hanging-wall strata near the projected location of the deformation front (Fig. 12B). The units in the boreholes are identified by engineering properties rather than stratigraphy. The first evidence consistent with the deformation front is that the oldest strata in the boreholes, probably interglacial (“Duwamish”) strata older than 125 ka, are only present in the boreholes south of the projected location of the deformation front (i.e., in the hanging wall), and they presumably lie below the boreholes if these strata are present beneath the north part of the cross section. The second piece of evidence in the borehole profile is that the unit identified as “glaciomarine drift” (GMD) has a continuous upper surface that is nearly flat north of the inferred location of the deformation front, but south of the front, it dips 3° to 9° north. Similar strata on the south flank of the “Duwamish” strata may define a hanging-wall anticline, although they could also be deposited on the side of an erosional valley.

Third, a 1969 geologic map of the Interstate 5 right-of-way (Wegner, 1969) shows an unconformable contact between Esperance Sand (ca. 17.4 ka) to the north and older, likely Possession-aged (ca. 70,000 yr B.P.) deposits identified as “Duwamish clay” (but unrelated to the “Duwamish” unit in the borehole profile) to the south (Fig. 12C). The trend and location of this geologic contact on the Interstate 5 map are almost identical to those of the synclinal axial surface as extrapolated from the seismic data. The contact is consistent with onlap of the Esperance Sand onto a sloping surface that could be a fold limb, or with folding of both units at the deformation front.
Figure 12. (A) Air photo of Seattle and bathymetric map of Elliott Bay showing the locations of the deformation front (red dashed line; line shifts slightly north as elevations increase because of its south dip), the transit tunnel (turquoise line), and the seawall profile (yellow line) relative to major roads and landmarks. Red rectangle shows location of Interstate 5 geologic map in part C. PPM—Pike Place Market. (B) Cross section drawn along the downtown transit tunnel (Shannon and Wilson, 1986), with the predicted location of the deformation front based on the marine seismic profiles shown as red dashed line. Note that south of the interpreted deformation front, the prominent red unit (glacio-marine drift [GMD]) dips to the north, whereas it is flat north of the deformation front. The older (“Duwamish”) strata present in the hanging wall south of the deformation front presumably lie below the drill holes north of the front. The features are consistent with uplift of material south of the deformation front; v.e.—vertical exaggeration. (C) Geologic map along Interstate 5 (Wegner, 1969) with red dashed line showing the location of the deformation front interpreted from the marine seismic profiles. The contact between the Esperance Sand and older “Duwamish clay” almost precisely aligns with the predicted location of the deformation front. (D) Cross section along the waterfront Seawall constructed from bore holes (Shannon and Wilson, 2012). Note the prominent valley at the location of the deformation front, the rising top surface of the pre-glacial strata to the south, and the upward step (possible fold scarp?) in the Estuarine and beach deposits.
Fourth, geologic cross sections from recent work for the Alaska Way Viaduct replacement project (Fig. 12D) show an asymmetric valley in the vicinity of the inferred deformation front, and a potentially uplifted Holocene-Pleistocene contact south of the front. The strata on this cross section are younger than on the other cross section, with postglacial strata above a glacial erosional surface likely ~16,500 yr old. The valley may be an extension of the valley seen in the bathymetry in Elliott Bay, and again it could be the result of preferential erosion within a fault or shear zone. Southward, the Holocene-Pleistocene erosional surface and overlying estuarine and beach deposits rise in elevation without the latter showing a thinning due to onlap onto the higher surface. This southward rise may be caused by folding or faulting near the deformation front, similar to the morphology of the wave-cut terraces south of Eagle Harbor on Bainbridge Island, and the distinct step could be a fold scarp.

If the geologic evidence reflects a fault or fold directly beneath downtown Seattle, it raises the possibility of fault-parallel zones of surface rupture or fold scarps beneath the tallest buildings and greatest concentration of infrastructure in the region. The A.D. 900–930 earthquake caused slight northward tilting (~0.2°) of land south of the deformation front above the main thrust fault, and fold or fault scarps above the backthrusts (Fig. 7C; ten Brink et al., 2006). If the deformation front extends east-west as we hypothesize, the shallow depth of the tips of the thrust faults on the seismic data (~0.6–1 km), the exposed Vasa Park thrust fault, and the back-thrust scarps indicate that surface rupture could occur beneath downtown Seattle. Ground shaking and damage to buildings would be greatest in a narrow band near the tips of thrust faults (e.g., Shi and Brune, 2005; Pai et al., 2007), and a fault beneath downtown Seattle places this area of maximum damage beneath the tallest buildings in the region, rather than being 1–1.5 km to the south. For these reasons, the inferred location of the deformation front beneath Seattle merits further investigation.

**KINEMATIC MODELING OF SEATTLE FAULT ZONE STRUCTURES**

**Testing of Kinematic Models**

Among the remarkable features of the Seattle fault zone are the wave-cut terraces along the shores of Puget Sound that were lifted during the A.D. 900–930 earthquake (Bucknam et al., 1992), and we can test the plausibility of Seattle fault zone fault geometries by modeling how well they reproduce the morphology of these terraces (e.g., ten Brink et al., 2006). Several features of the terrace morphology are keys to constraining the kinematic models. The terrace that extends farthest north on the east shore of Bainbridge Island has a relatively steep north side, and the terraces on both sides of Puget Sound show south-facing scarps and a gentle southern edge. Perhaps the strongest constraint for the models, however, is the subsidence of the footwall just north of the deformation front that has been documented during the A.D. 900–930 earthquake at both Eagle Harbor on Bainbridge Island and at West Point on the east shore of Puget Sound (Fig. 4; ten Brink et al., 2006).

For our preferred thrust-fault (fault propagation fold) model (Fig. 1B), frontal fault A extends from 0.6 to 11 km depth with a 40° south dip (18 km downdip width), and the three backthrusts that form scarps in the wave-cut terrace (Toe Jam Hill/Point Glover, West Seattle, and backthrust “e” in Fig. 4) dip 40° north (Fig. 13A). Where exposed in trenches, the Toe Jam Hill backthrust ranges from near-vertical to dipping 30° north (Nelson et al., 2003a), and the Waterman Point backthrust dips northward at 40° to 45° (Nelson et al., 2003b). Reflector terminations on the seismic profiles are consistent with back-thrust dips of ~40° north beneath Puget Sound (Figs. 5–7).

Compressing the thrust-fault model produced an uplift with similar shape and magnitude to the observed terrace with 12.7 m of average slip on fault A, and the three north-dipping backthrusts produced south-facing scarps with similar amplitudes (~1–3 m) to those on the adjacent land areas (Fig. 13A). The predicted terrace uplift fits nearly all of the surveyed terrace elevations and mimics the overall shape of the LiDAR profiles. The three surveyed points south of the main terrace uplift appear contradictory (south of kilometer 4 in the model), but they are on different sides of Puget Sound and could be due to growth of additional structures such as the anticline beneath Blake Island just south of the Seattle fault zone (Figs. 2–4) or an as-yet-unnamedified backthrust. Changing the downdip length of the main thrust fault allows us to fit any of these points (Fig. 13A), but not all three simultaneously. The model results clearly show that a thrust-fault model is consistent with the overall terrace morphology, even if the slip accumulated during multiple earthquakes. Furthermore, the thrust-fault model can rotate the forelimb strata to any angle (Ersliev, 1991; Allmendinger, 1998; Shaw et al., 2005), providing an explanation for the steep forelimb dips exposed on Bainbridge Island (Fulmer, 1975; McLean, 1977; Kelsey et al., 2008) and the progressive rotation of strata evident on the seismic profiles from Lake Washington (Fig. 9).

The total of 12.7 m of average slip required to create the terrace morphology is similar to estimates of fault slip from previous modeling studies of the Seattle fault zone (Mueller and Harding, 2007; ten Brink et al., 2006). Regressions of average slip versus moment magnitude on reverse faults indicate that 12.7 m of average slip is consistent with a M 7.0–7.5 earthquake (Wells and Coppersmith, 1994). This may be a maximum magnitude for the A.D. 900–930 event, however, because some of the terrace elevation is likely due to the earthquakes that ruptured the backthrusts at other times. As analogs, the 1999 Mw 7.6 Chi-Chi, Taiwan, earthquake had as much as 20 m of slip on an ~40 km by ~20 km asperity near the north part of the fault (Chi et al., 2001; Ma et al., 2001), and slip during the 2001 Mw 7.6 Bhuj, India, earthquake was concentrated in a 20 km by 15 km asperity that had as much as 12 m of slip (Antolik and Dreger, 2003).

Roof-thrust models do not reproduce the relatively steep north edge of the uplifted terrace or the subsidence just north of the deformation front (Figs. 13B and 13C). The original roof-thrust model has a shallowly dipping thrust fault that creates only modest uplift above the fault tip, resulting in the north slope of the terrace being too gentle and the highest terrace elevations being too far south (Fig. 13B). More importantly, the model does not create any subsidence north of the fault tip. An alternative roof-thrust model in which the thrust fault...
Figure 13. Uplifts of wave-cut terraces above different fault models. Slip on the modeled faults (black lines) is in response to compression of the model space. Red denotes areas of uplift, and blue denotes subsidence, with the density of color proportional to the elevation change. Black dots are surveyed terrace elevations with 1 m error bars (ten Brink et al., 2006), and gray lines show light detection and ranging (LiDAR) elevations of the landward terrace edges along southern Bainbridge Island and West Seattle. (A) Thrust-fault model. The top of the main fault is at 0.8 km depth, and the three backthrusts are included (Toe Jam Hill/Point Glover, West Seattle, backthrust “e”). Different downdip widths (solid = 18 km; dashed = 12 km) to the main thrust fault match the conflicting elevations on opposite sides of Puget Sound south of the main terrace uplift. (B) A roof-thrust model as originally proposed (Brocher et al., 2004) cannot reproduce the steep north side of the terrace and does not show subsidence north of the deformation front. The model also concentrates the uplift toward the south end of the terrace, in contrast to the observed terrace morphology, which has the greatest uplift nearer the deformation front. (C) An alternative roof-thrust model with a steeper top to the wedge and a flat base (Kelsey et al., 2008) also does not match the steep north side of the terrace or cause subsidence north of the deformation front. This model also has the greatest uplift near the south part of the terrace rather than nearer the deformation front. (D) An alternative roof-thrust model with the wedge top at 3 km depth and the lower thrust fault dipping 50° south (ten Brink et al., 2006). This model fits the terrace morphology reasonably well, but the wedge tip is shallower than is plausible, and a blunt wedge is not viable with continued slip on the lower thrust fault (see text).
flattens to form a wedge with a horizontal base (Fig. 13C) also results in the highest terrace elevations being too far south and lacks subsidence north of the deformation front. Our conclusion from the modeling is that the relatively steep north slope of the terrace and the subsidence north of the deformation front require a moderately steep, south-dipping thrust fault reaching shallow depths beneath the deformation front, as also concluded by Muller and Harding (2007) and ten Brink et al. (2006).

A hybrid roof-thrust model that moves the tip of a south-dipping thrust fault to ~3 km depth beneath the deformation front creates an uplift pattern that matches most of the terrace morphology, provided the thrust fault has a relatively steep dip of ~50° (Fig. 13D; ten Brink et al., 2006). This fault geometry, however, is kinematically unrealistic and violates stratigraphic evidence. A blunt wedge is not a viable long-term kinematic model because continued slip on the main thrust fault would force it to either cut upward toward the surface, causing it to evolve into the thrust-fault model that we prefer, or to bend to a horizontal position as in the previous roof-thrust model. Also, the ~2400 m thickness of Blakeley Formation exposed on southern Bainbridge Island (Fulmer, 1975; McLean, 1977; Kelsey et al., 2008) requires the wedge tip to lie substantially below the top of Blakeley Formation rocks to bring them into the hanging wall above the wedge tip. The top of the Blakeley Formation is at a depth of nearly 4 km, based on seismic-reflection and drill-hole data (Johnson et al., 1994; ten Brink et al., 2002), which means the wedge tip would have to lie substantially below 4 km depth (e.g., Brocher et al., 2004; Kelsey et al., 2008).

The roof-thrust models also contradict other observations. First, compelling evidence for the thrust-fault model is provided by the south-dipping thrust faults at shallow depths, as interpreted here and by Johnson et al. (1999) and Stephenson et al. (2006), and by the exposed thrust faults in the south Bellevue trench (Sherrod, 2002), in Green Mountain west of Puget Sound (Haeussler and Clark, 2000), and on southern Bainbridge Island (Hauerud, 2005). The severe disruption of Seattle fault zone forelimb strata evident on the seismic profiles is not explained by a roof-thrust model, which should produce a simple fold to create continuous north-dipping forelimb strata.

Second, roof-thrust models are not consistent with the steep dip of the forelimb strata. The strata above a thrust wedge should dip at an angle slightly steeper than, but opposite from, the ~40° dip of the master thrust fault (e.g., Fig. 1C). In contrast, exposed Miocene and Oligocene strata on Bainbridge Island dip northward at angles as great as 70°–90° or are overturned beneath southern Bainbridge Island and Alki Point (Fulmer, 1975; McLean, 1977; Kelsey et al., 2008). Steepening by imbricate thrusting (Dimieri, 1997; Brocher et al., 2004) is not a plausible explanation because repetitions of the stratigraphic section within duplexes are not observed on Bainbridge Island (Fulmer, 1975; McLean, 1977).

Third, there is no evidence for active thrust faults above a roof thrust (shallow detachment) between the Seattle and Tacoma faults despite Quaternary and Holocene motion on both of these fault systems. Detailed LiDAR imagery and examination of the shorelines between the Seattle and Tacoma faults (Booth et al., 2004) have not identified any fault ruptures at the surface (e.g., Sherrod et al., 2004; Barnett et al., 2010) despite the roof-thrust model predicting ~10 additional backthrusts between these active fault zones.

Fourth, the amount of uplift of the horizontal décollement forming the hypothesized roof thrust does not match the total uplift above the Seattle fault. The detachment would presumably have formed near the 4–6 km depth of the wedge tips, so the current ~2 km depth of the décollement implies that there has been ~2–4 km of uplift since the roof thrust formed. In contrast, uplift of ~7 km of Crescent Formation rocks has occurred above the Seattle fault zone (Johnson et al., 1994; Pratt et al., 1997; Blakely et al., 2002; ten Brink et al., 2002), which would have brought the subhorizontal detachment to an elevation above sea level.

Finally, a roof-thrust or wedge model does not predict progressive rotation of forelimb strata and the steepening of forelimb strata with age, as observed near the deformation front on the Lake Washington profiles (Fig. 9). A roof thrust should produce a growth triangle as younger strata move through the axial surface and up onto the wedge (Shaw et al., 2005; Kelsey et al., 2008) rather than the progressive rotation of older strata that characterizes the forelimbs of fault-propagation folds, especially those exhibiting trishear behavior (Erslev, 1991; Allmendinger, 1998; Shaw et al., 2005).

**Models of Back-Thrust Earthquakes**

Two potential mechanisms for forming backthrusts within forelimb strata (Fig. 14) are by development of an uplifted triangular zone between the main thrust fault and a backthrust, with both faults cutting across bedding, and by bedding-plane slip within forelimb strata during growth of the fault-propagation fold. The first style of backthrust is prevalent above thrust faults that form in the presence of layered strata (e.g., Niño et al., 1998; Schultz, 2000), particularly if there is a steepening of the shallow portions of the fault (e.g., Mitra, 1997).

**Figure 14.** Styles of backthrusts. (A) Backthrusts cutting across strata near the tip of a thrust fault (from Neely and Erslev, 2009). (B) Diagram of a fault-propagation fold showing the bedding-plane slip and location of potential backthrusts caused by the folding near and above the fault tip (modified from Erslev and Mayborn, 1997).
2002). Examples are numerous, commonly on a scale similar to the Seattle fault zone backthrusts (e.g., Lee et al., 2002, 2005; Davis et al., 2005; McLaren et al., 2008; Giambagi et al., 2009; Higgins et al., 2009; Neely and Erslav, 2009), and within the Seattle fault zone itself (Haeussler and Clark, 2000; Liberty and Pratt, 2008). A second style of forelimb backthrust can form because the bedding-plane slip accommodating folding within a fault-propagation fold can move shallower layers toward the anticlinal axial surface (e.g., Erslav and Mayborn, 1997). Continued motion can eventually cause these backthrusts to cut across overlying strata within the anticlinal fold. Examples of these bedding-plane backthrusts are less common in the literature (e.g., Erslav and Mayborn, 1997; Chester and Chester, 1990; fig. 6 in Rowan, 1997), possibly because their position along bedding planes can make them difficult to identify.

Regarding the precise mechanism for forming the Seattle fault zone backthrusts, the timing of slip on them relative to the A.D. 900–930 earthquake is one of the most perplexing aspects of the Seattle fault zone. Trenches across the backthrust scarps show 1–2 m of fault slip in multiple earthquakes before, during, and after the A.D. 900–930 earthquake (Nelson et al., 2003a, 2003b, 2014; Kelsey et al., 2008). If the backthrusts ruptured in concert with the south-dipping thrust faults, the back-thrust scarps should be superimposed on a broader uplift formed by motion on the larger south-dipping thrust fault. However, localized wave-cut terraces extending ~400 m behind some of the back-thrust scarps are not contemporaneous with a recognized broader, widely distributed terrace, which suggests that the backthrusts could rupture independently of motion on a deeper, main thrust fault. These terrace relationships and trenching results led to the suggestion that the backthrusts can store enough strain during aseismic creep on the main thrusts to later rupture by themselves in small to moderate earthquakes (M 6.0–6.5; Kelsey et al., 2008).

It is difficult to reconcile the amount of back-thrust slip during individual earthquakes with the dimensions of the back-thrust faults if the backthrusts rupture independently from the main thrust faults. The backthrusts have maximum downdip widths of only 3–6 km before they intersect either a south-dipping thrust fault or the axial surface (Figs. 1B and 1C), which limits the maximum magnitudes of earthquakes that can occur on the backthrusts alone to be small or moderate (M 6.0–6.5; Kelsey et al., 2008). Earthquakes of M 5.5–6.0 on such small faults rarely have slip approaching 1 m (Bonilla, 1988), yet greater than a meter of slip has occurred on at least five of the backthrusts (Nelson et al., 2003a, 2003b). One meter of slip is more characteristic of moderate earthquakes of M 6.5–7.0 (Wells and Coppersmith, 1994; Biasi et al., 2011). This problem is especially acute for the IslandWood scarp near the deformation front, which may have had ~1.6 m of slip during a single earthquake (Nelson et al., 2014) despite having an estimated downdip extent of only ~3 km (Kelsey et al., 2008). It is also difficult to understand how sufficient strain to produce a M 5.5 earthquake can be stored on such small faults within weak sedimentary strata, especially since the slip could occur along the weakest bedding planes. The amount of slip on the backthrusts led Nelson et al. (2003a) to conclude that the back-thrust ruptures must be coincident with earthquakes on the main thrust faults, and we agree with this conclusion.

The small terraces on the hanging walls of the backthrusts suggest several Seattle fault zone earthquake scenarios that can be tested with modeling. Scenario 1 is that a large (M 7–7.5) earthquake on the main fault and simultaneous rupture of one or more backthrusts causes a broad, high (~6–7 m) terrace with smaller scarps superimposed, as we modeled earlier (Fig. 13A). This is the most likely rupture scenario during the A.D. 900–930 earthquake. Scenario 2 is that a moderate (M ~6.5) earthquake on the main fault produces a broad, low terrace without any back-thrust ruptures (Fig. 15A). There is no firm evidence that this has occurred in the Seattle fault zone near Puget Sound, although a low terrace of 1 m or less formed during a moderate earthquake would be difficult to recognize and map, and rapid erosion of low terraces is demonstrated by the West Seattle scarp, which is ~1.5 m high on the terrace uplifted during the A.D. 900–930 earthquake (Kelsey et al., 2008) but has been removed by erosion within the adjacent wave zone. Scenario 3 is that a backthrust ruptures by itself in a small to moderate earthquake (M 5.6–6.0), causing uplift of a narrow terrace behind the backthrust without uplift of an accompanying regional terrace (Fig. 15B). Presumably, the shortening needed for such an earthquake is accomplished by aseismic creep on the deeper thrust faults, with erosion keeping pace with the gradual uplift of the regional terrace caused by the fault creep.

To reconcile the large amount of back-thrust slip (1–2 m) with the small fault areas of the backthrusts (~4–6 km downdip width), we favor a fourth scenario in which slip on the backthrusts (Toe Jam Hill in our model) is driven by moderate earthquakes on an underlying main thrust fault (fault A in the model). In our model, a moderate (~6.5–6.7) earthquake with ~2 m average slip on the 40°S-dipping main thrust fault produces a low (~0.6-m-high) terrace that extends ~5 km south of the back-thrust scarp. The ~2.6 m amount of maximum slip in the model is consistent with the ~3 m of slip on the south-dipping Seattle fault zone thrust fault at Vasa Park during an earthquake 5000–11,000 yr ago (Sherrod, 2002), and it is only slightly larger than the slip in the 1994 M 6.7 Northridge, California, earthquake (e.g., Wald et al., 1996). The low terrace produced by this earthquake would be difficult to recognize and could be quickly eroded in the wave zone (Fig. 16A). Note that the back-thrust model that invokes creep on the main fault also assumes erosion of a low terrace formed by aseismic slip on the main fault, but in that case, the erosion is assumed to be concurrent with the gradual uplift. The motion on the main thrust fault in our model transfers ~1.6 m of slip onto a small backthrust to produce an ~1.4-m-high scarp. This back-thrust scarp is superimposed on the broader, low terrace from the main thrust fault, raising it to an elevation ~2.0 m above sea level (Fig. 15C). After erosion of the low regional terrace, the back-thrust scarp and localized terrace behind it would look like the backthrust ruptured by itself. The rupture of the backthrust in concert with motion on the main thrust fault provides a plausible mechanism for causing the large amounts of slip evident on the backthrusts.

Our modeling implies that the back-thrust ruptures identified in scarp studies were caused by moderate (~M 6.5 to M 6.7) Seattle fault zone earthquakes, as originally interpreted by Nelson et al. (2003a), rather than small to moderate
Moderate thrust-fault earthquakes can cause strong ground motions and extensive damage, as, for example, the three deaths and $358$ million of damage in the Los Angeles, California, area during the 1987 M 5.9 Whittier Narrows earthquake (Hauksson et al., 1988), or the $60$ deaths and $>$$13$ billion damage caused by the 1994 M 6.7 Northridge earthquake (USGS and SCEC, 1994). Modeling of ground motions for the Seattle earthquake scenario indicates ground accelerations as great as $0.7g$ during a M 6.7 earthquake, with large areas experiencing accelerations of $0.3g$ or more (Weaver et al., 2005). An unusually extreme case is the 22 February 2011 Mw ~6.3 Christchurch, New Zealand, earthquake, which had 2–4 m of slip on a blind reverse fault, and which caused ~0.3 m of broad surface uplift, accelerations as great as $2.2g$, and about $30$ to $40$ billion in damages (Barnhart et al., 2011; Beavan et al., 2011; Bradley and Cubrinovski, 2011; Holden, 2011).

**DISCUSSION**

**Comparing Seattle Fault Zone Earthquake History with Amount of Shortening**

Our modeling yields estimates of the amount of shortening during past Seattle fault zone earthquakes, and we compared this shortening to that estimated from geodetic and geologic data. The results provide an estimate of the percent of the total shortening in the forearc that has been accommodated by motion on faults within the Seattle fault zone. Paleoseismic evidence for earthquakes on the Seattle fault zone backthrusts primarily comes from scarp studies (Nelson et al., 2003a, 2003b, 2014) and from uplifted terraces (Kelsey et al., 2008). These studies only include earthquakes evident in the paleoseismic record as surface ruptures or uplifted marine terraces. The record does not include earthquakes on blind thrust faults that did not cause surface rupture and/or were not large enough to cause strong ground shaking to trigger secondary events.

In addition to earthquakes known from surface ruptures, an independent record of episodes of strong ground shaking in the Seattle fault zone area comes from earthquake-triggered turbidite deposits in Lake Washington (Karlin et al., 2004); these earthquakes potentially include both Seattle fault zone earthquakes as well as those on other, nearby fault systems. The history of great Cascadia plate-boundary earthquakes, which may have triggered some of the Lake Washington turbidites, is given in Nelson et al. (2006). Not all of the plate-boundary earthquakes correspond with turbidites, however, so the correlation of turbidites with the time of a plate-boundary earthquake does not necessarily show causality. In contrast, the Seattle fault zone cuts directly beneath Lake Washington, and any large Seattle fault zone earthquake therefore likely triggered turbidites within the lake. We compiled a history of Seattle fault zone earthquakes over the past ~7 k.y. in Figure 17A (Nelson et al., 2014; Sherrod and Gomberg, 2014).
Although scarp growth at inland sites can be identified in trenches, coseismically uplifted wave-cut terraces are more complex because of the interplay between terrace uplift and erosion (Fig. 16). The high wave energy in Puget Sound appears capable of eroding terraces ≤2 m in height in about a thousand years or less, such as, for example, the ~1.5-m-high West Seattle scarp that crosses the regional A.D. 900–930 terrace but has been removed within the wave zone (Fig. 16A; Kelsey et al., 2008). Also, higher back-thrust terraces such as the 3.5- to 4-m-high terrace behind the Point Glover fault (Kelsey et al., 2008) could form during two earthquakes, with the older terrace being preserved by uplift during the second earthquake and erosion of the younger terrace resulting in a single high terrace that would appear to have formed in a single earthquake (e.g., Fig. 16B). The higher (3–5 m) localized terraces like those behind the Toe Jam Hill and Point Glover scarps therefore could be the result of multiple earthquakes rather than requiring 3–5 m of back-thrust slip in a single earthquake.

Figure 16. (A) Formation and removal of a low (<1 m) terrace during a small regional earthquake. The terrace forms outboard of the pre-earthquake shoreline, but it is barely lifted above the waterline. Erosion rapidly removes the small terrace until the shoreline reaches the older, higher shoreline, at which point the erosion slows because of the higher elevation of the land. (B) Formation of a high terrace from two smaller terraces. An initial terrace forms in the first earthquake with about a 2 m height (top). A second earthquake occurs before the first terrace is removed, forming a small (~2 m height) terrace outboard of the first terrace; erosion begins removing the lower terrace toward the shore (middle diagram). Eventually, the outboard terrace is completely removed, and erosion slows because of the higher shoreline; the remaining terrace appears to be a single terrace despite being lifted during two separate earthquakes.
Figure 17. (A) Paleoseismic record of Seattle fault zone (SFZ) earthquakes over the past 3500 yr, with large Cascadia subduction zone earthquakes also shown. The horizontal black bars show the range of time for documented earthquakes at each of the paleoseismic sites and inferred from turbidite deposits in Lake Washington. Vertical bars show inferred earthquakes on the Seattle fault (dark gray; dashed where uncertain), and on the Cascadia subduction zone (SZ; light gray). Inferred magnitudes of the earthquakes are listed at the bottom of the graph. See text for explanation. (B) Total north-south shortening across the Cascadia forearc based on geodetic data compared to accumulated shortening from known Seattle fault zone (SFZ) earthquakes. The earthquakes correspond with those in part A. Gray dashed, sloping lines show 75%, 50%, and 25% of the total forearc shortening inferred from geodetic data.
The paleoseismic record in the Seattle fault zone is consistent with a moderately large earthquake ca. 6.9 ka, followed by ~3500 yr of apparent quiescence before a cluster of earthquakes that included the large A.D. 900–930 event (Sherrod et al., 2000; Nelson et al., 2014). The following earthquakes are recognized in the paleoseismic record over the past ~75 k.y., during which time a marsh at Restoration Point has chronicled a nearly complete record of sea-level history (Sherrod et al., 2000), within which sudden uplift should be distinct. Sequences of wetland soils adjacent to back-thrust scarp should also record all major surface ruptures (Nelson et al., 2003a, 2014). We estimate the magnitudes of these earthquakes from the lower parts of the probability functions of Biasi et al. (2011) and the regressions in Wells and Coppersmith (1994); the shortening estimates come from our modeling (Figs. 13A and 15C).

**Earthquake 11.5–4.5 ka, M ~6.7.** During this broad time frame, a south-dipping thrust fault at Vasa Park, 16 km southeast of Seattle, broke the surface with ~3 m of slip (Sherrod, 2002). The amount of slip indicates a M ~6.7 earthquake that likely ruptured a fault with a length of 30–50 km, as described in the earthquake scenario of Weaver et al. (2005). A total of ~2.4 m of N-S shortening is indicated by the 3 m of slip in the trench (Sherrod, 2002; Weaver et al., 2005).

**Thrust fault earthquake ca. 6.9 ka, M ~6.7.** A change in salinity and biota in a small marsh at Restoration Point is consistent with coseismic uplift of ~1.5 m, although it also could alternatively have been caused by accretion of a beach berm (Sherrod et al., 2000). It is plausible from the magnitude and timing that this was the earthquake recorded in the Vasa Park trench, and we assume so here. Regardless of this correlation, the ~1.5 m of uplift recorded in the salt marsh at Restoration Point indicates ~1.8 m of shortening in the Puget Sound region. After this disturbance, the Restoration Point marsh apparently remained near sea level for nearly 3800 yr.

**Disturbance ca. 3.5–2.7 ka.** Sherrod et al. (2000) interpreted a change in salinity at the Restoration Point marsh as indicating uplift of ~4.8 m, suggesting coseismic formation of a regional terrace about two thirds the height of the A.D. 900–930 terrace. There is also a set of turbidites in Lake Washington that spans this time range. However, such substantial coseismic uplift would require subsequent rapid sea-level rise to maintain the site’s position near sea level, and the apparent uplift could be due to environmental changes (e.g., berm accretion), and the turbidites could be due to a plate-boundary earthquake. This disturbance is near the start of a series of back-thrust events, so it is possible that this is the first of a sequence of moderate events that ruptured the various backthrusts. An earthquake in the M 6.7 to M 7.0 range on a south-dipping thrust fault could cause ~2.4 m of shortening, 2.8 m of slip, and 1.5 m of uplift. The latter is a small enough uplift to cause a disturbance in the marsh but also plausibly create a terrace uplift that was low enough for it to be subsequently eroded or not be recognized.

**Toe Jam Hill back-thrust earthquake (3–2 ka), M ~6.5.** Event A either is an earthquake or is a disruption due to a falling tree, the roots of which spanned the Toe Jam Hill scarp (Nelson et al., 2003a). There is no other evidence for an earthquake at this time, and there are no associated turbidites in Lake Washington. Given the small amount of scarp growth (~1 m) and the lack of a turbidite, this earthquake, if indeed there was an earthquake, was probably moderate in size (M ~6.5), with ~1.2 m of shortening causing ~1 m of uplift. This possible earthquake is the first known on any Seattle fault zone backthrust, indicating a long period with no back-thrust events before event A.

**Toe Jam Hill earthquake B (2.5–1.9 ka), M ~6.5.** This earthquake ruptured the Toe Jam Hill scarp between 2500 and 1900 yr ago, and a turbidite occurred in Lake Washington 2300–1900 yr B.P. This earthquake produced 0.9 m of scarp uplift that indicates ~1.2 m of shortening, but uplift could be as much as 1.9 m if earthquake A did not occur. We use 0.9 m here because we also include earthquake A in our summary; the total shortening would be similar if we do not include earthquake A but use 1.9 m in earthquake B. Presumably, the earthquake also caused rupture of the Point Glover scarp that is in line with the Toe Jam Hill fault (Kelsey et al., 2008), creating localized terraces behind these backthrust scarps. The lack of a regional terrace and the presence of turbidites in Lake Washington about this time suggest that this was a moderate earthquake (M ~6.5) rupturing a single backthrust.

**Toe Jam Hill earthquake C (1.3–1.2 ka), two events?** Event C is perplexing because of the conflicting evidence regarding its magnitude. The IslandWood scarp had ~1.6 m of uplift during earthquake C, and there is a higher terrace behind the West Seattle scarp that predates the more obvious scarp cutting across the A.D. 900–930 terrace (Kelsey et al., 2008). The Toe Jam Hill scarp had 0.7–1.3 m of slip, and the Point Glover scarp has a backthrust terrace that is consistent with an earthquake a few centuries before the A.D. 900–930 earthquake. The simultaneous rupture of all of these backthrusts would suggest a large earthquake (M ~7+), but an earthquake of this size is implausible because there are no corresponding turbidites in Lake Washington or uplift of either a large regional terrace or the salt marsh at Restoration Point. An alternative is that the various backthrusts ruptured during two moderate events in the 1.9–1.2 ka time frame as depicted in Figure 17A (events C1 and C2). An initial earthquake (C1) corresponding to the turbidite sequence ca. 1.9–1.4 ka could have caused rupture of the IslandWood and West Seattle backthrusts with ~1.5 m of uplift (M ~6.7), and later earthquake C2 ruptured only the Toe Jam Hill and Point Glover backthrusts with ~1 m of slip (M ~6.5). Shortening would be ~1.6 m and 1.0 m, respectively.

**Regional earthquake (Toe Jam Hill earthquake D), M ~7–7.5.** This is the A.D. 900–930, M 7.0–7.5 earthquake that uplifted the regional terrace ~6.5 m (ten Brink et al., 2002; Kelsey et al., 2008). It also appears to have ruptured the Toe Jam Hill (~1 m scarp growth), West Seattle (~1.5 m scarp growth), Waterman Point, and IslandWood (~0.6 m scarp growth) backthrusts, and likely caused turbidites in Lake Washington (Karlin et al., 2004). Slip models (Fig. 13A; ten Brink et al., 2006; Muller and Harding, 2007) indicate 10–13 m of slip and 8–10 m of north-south shortening. The amount of shortening can be reduced slightly (~15%) if we assume that some of the regional terrace uplift occurred during back-thrust events shortly before or after, leaving ~6 m of uplift, ~9.3 m of slip on thrust fault A, and a total shortening of ~73 m. This is ~20% of the ~35 m of shortening in the Cascadia forearc over the past ~7 k.y. (Fig. 17B).
Earthquake E (1–0.2 ka), $M \sim 6.5$. The most recent Seattle fault zone back rupture is probably due to a moderate ($M \sim 6.5$–$6.7$) earthquake that ruptured the Waterman Point backthrust at least several decades after the A.D. 900–930 earthquake (Nelson et al., 2014). The earthquake predates European settlement in 1851. This earthquake caused $\sim 1$ m of scarp growth at Waterman Point and formed low terraces outboard of the higher A.D. 900–930 terrace. The small amount of slip suggests it was at the low end of the moderate range that we modeled ($M \sim 6.5$), with shortening of $\sim 1.2$ m.

Reconciliation of Back-Thrust Earthquakes with Total Shortening

We can compare the approximate shortening across the Seattle fault zone, based on the shortening in our models during individual earthquakes, to estimate what percent of the forearc deformation is taken up by slip on faults within the Seattle fault zone. The north-south, permanent shortening rate across the outer forearc of Washington State, caused by northward motion of the Coast Range block, is estimated from geodetic data to be $\sim 5$ mm/yr (Mazzotti et al., 2002; McCaffrey et al., 2013). This rate implies a total shortening of $\sim 35$ m over the past 7000 yr (Fig. 17B), which is the approximate length of time for which we have a paleoseismic record.

We estimate that the documented earthquakes on the Seattle fault zone in the past $\sim 7$ k.y. have accommodated $\sim 173$ m of horizontal shortening (Fig. 17B). Note that the shortening estimates in our models are not dependent on our conclusion that the backthrusts rupture in moderate earthquakes: If the backthrusts are loaded by aseismic creep on the main faults, the amount of creep would have to be similar to what we hypothesize occurs seismically. If we place a $\pm 30\%$ uncertainty on our estimate, it yields a range of $12.1$–$22.5$ m of shortening. This is $\sim 50\%$ (35% to 65%) of the total shortening in the forearc, confirming inferences that the Seattle fault zone has been the primary seismic hazard to the region in the late Holocene. The apparent lack of Seattle fault zone surface ruptures in the 3500 yr preceding the cluster of back-thrust earthquakes suggests that slip may have been concentrated on other faults during that time frame. Another intriguing possibility is that Seattle fault zone earthquakes were predominantly blind before $\sim 3500$ yr ago, but more recently have become primarily surface rupturing.

An independent check on the percentage of shortening absorbed by the Seattle fault is provided by the $\sim 24$ m and $\sim 48$ m of uplift above faults A and B, respectively, on profile P42 (Fig. 7C). This total $\sim 72$ m of uplift requires $\sim 86$ m of shortening, which is consistent with our earlier inference that the shortening is greater than the $\sim 82$ m that geodetic data indicate in the $\sim 16.5$ k.y. of postglacial time, and the prominent reflector therefore must be older. Spread over $\sim 60$ k.y. since the previous glacial retreat, the $\sim 86$ m of shortening deduced from profile P42 represents $\sim 29\%$ of the total $\sim 300$ m of shortening expected in the forearc. Converted to a recurrence interval with a Gutenberg-Richter b-value of 1 (10 moderate earthquakes for each large earthquake), the data are consistent with having 4–5 large and 40–50 moderate earthquakes on the Puget Sound portion of the Seattle fault zone in the past $\sim 60$ k.y., which implies average recurrence intervals of $\sim 12,000$–$15,000$ yr for large events ($M \sim 7$–$7.5$) and $\sim 1150$–$1500$ yr between moderate ($M \sim 6.5$) events. These recurrence intervals are broadly consistent with the paleoseismic record along the Puget Sound corridor over the past $\sim 7$ k.y. if we average out the apparent clustering of the back-thrust events (i.e., few back-thrust earthquakes from ca. 7 to 3.5 ka, but 4–5 since ca. 3.5 ka).

Note that the shortening estimates here are for the Puget Sound region of the Seattle fault zone, and other portions of the fault system may have a different history of moderate earthquakes. The back-thrust scarps appear to have short fault lengths of 6–12 km (e.g., Kelsey et al., 2008), so presumably there are other backthrusts on adjacent portions of the Seattle fault zone that are now obscured by development. Moderate earthquakes might rupture only about one third of the total fault length, as in the scenario $M \sim 6.7$ earthquake in Weaver et al. (2005). For moderate earthquakes, therefore, paleoseismic studies along the Puget Sound corridor may be recording only about one third of the total number of moderate events distributed along the length of the Seattle fault zone. We also point out that not all of the moderate earthquakes need to rupture the surface and therefore be expressed in the paleoseismic record, as some fraction of them could be blind ruptures, like, for example, the $M \sim 5.9$ Whittier Narrows and $M \sim 6.7$ Northridge earthquakes in southern California. The recurrence interval therefore could be substantially shorter than $1150$–$1500$ yr for moderate earthquakes along the entire length of the Seattle fault zone.

Relationship between Faults and Seismicity

There is a long-standing enigma regarding the lack of a direct association between instrumentally recorded earthquakes and known faults beneath the Puget Lowland (e.g., Pratt et al., 1997; Brocher et al., 2001, 2004; Blakely et al., 2002; Van Wagoner et al., 2002), with the exception of the 1995 $M \sim 5.0$ Point Robinson earthquake, which may have occurred on the lower part of the Seattle fault (Dewberry and Crosson, 1996). The most significant earthquake near the shallow part of the Seattle fault zone in the past two decades was the $M \sim 4.9$ Point White earthquake (also called the Bremerton earthquake) that occurred at $\sim 8.8$ km depth beneath southern Bainbridge Island in 1997 (Fig. 4). The Point White earthquake and its aftershocks appear to define a steeply ($\sim 65^\circ$ to $85^\circ$) north-dipping, north-side-up fault plane that projects to the surface near the Toe Jam Hill back-thrust scarp within the Seattle fault zone (Blakely et al., 2002). The alternative plane is gently south-dipping, perhaps denoting northward thrusting along a low-angle fault. The hypocentral depth, originally placed at 4–8 km depth, locates at $\sim 8.8$ km depth with the current network velocity model, although the error estimate remains large. This depth places the main shock well below the Seattle fault zone thrust faults and any proposed thrust wedge (Blakely et al., 2002; Brocher et al., 2004). Blakely et al. (2002) suggested that the north-dipping fault plane could be a deep-seated reverse
fault that projects to the surface near some of the back-thrust scarps, but a deep-seated fault would seem inconsistent with the small lateral extent of the backthrusts and reverse motion on the north-dipping plane, which is opposite to the long-term subsidence of the Seattle basin. Kelsey et al. (2008) suggested that the earthquake actually was located at ~6 km depth, and that the main shock was a bedding-plane back-thrust event with aftershocks concentrated along the synclinal axial surface at the deformation front in a roof-thrust (wedge) model. The 65°N to 85°N dip of the fault plane approximately aligns with the dip of bedding at the surface, but as we discussed already, it is not clear how beds can reach that steepness in the proposed roof-thrust model. Also, the earthquake epicenter was located beneath the surface location of the Toe Jam Hill fault (Fig. 4), which is several kilometers south of the synclinal axial surface that lies at the downdip end of the north-dipping Toe Jam Hill fault plane. The Point White earthquake thus does not appear to fit either of the kinematic models discussed here (Figs. 1B and 1C).

Like previous workers (e.g., Blakely et al., 2002, we do not understand the relationship between most of the seismicity and fault planes within the Seattle fault zone, nor have we identified the fault on which the Point White earthquake occurred.

Main-Thrust and Back-Thrust Relationships

The shallow backthrusts discussed here are different from deep-seated backthrusts that diverge from the main thrust fault at midcrustal depths. The Tacoma fault, which reaches the surface 20–30 km south of the Seattle fault zone, appears to be a deeply rooted backthrust that merges with the Seattle fault at midcrustal depths (e.g., Brocher et al., 2004). The Tacoma fault likely ruptured in concert with the Seattle fault during the A.D. 900–930 earthquake (Sherrod et al., 2004; Nelson et al., 2014), but the fault is large enough and extends deep enough that it clearly could pose an independent seismic hazard. An example of a backthrust on the same scale as the Tacoma fault rupturing independently of the main fault is the Pico thrust, which is a backthrust to the Elysian Park fault system that ruptured in the 1994 Northridge earthquake (Davis and Namson, 1994).

The rupture lengths of the shallow Seattle fault zone backthrusts discussed in this paper are unknown, but the amount of slip in individual earthquakes suggests rupture lengths of 20–40 km based on regressions of empirical data (Wells and Coppersmith, 1994) and on model probabilities of rupture length (Blasi et al., 2011). There is not a direct correlation of most of the backthrusts across Puget Sound, as would be expected if they had lengths of 20 km or more, but the urban areas on both sides of Puget Sound may obscure backthrust scarps. Backthrusts of a similar scale have been mapped in south Bellevue from seismic-reflection profiles (Liberty and Pratt, 2008) and are mapped in Green Mountain to the west (Haeussler and Clark, 2000). One possibility is that the backthrusts observed on Bainbridge Island and south Bellevue are part of a set of short, en echelon backthrusts that form the wide Seattle fault zone.

Similar, complex sets of thrust and back-thrust faults form near the tips of major fault systems elsewhere (e.g., Neely and Erslev, 1999; Giambiagi et al., 2009). The Seattle fault zone fault geometries presented here provide insights into the interpretation of earthquakes on shallow main-thrust and back-thrust faults near the tips of major thrust systems. The shallow Seattle fault zone backthrusts formed in weak forelimb strata at low confining pressures, making it unlikely that they can store large amounts of strain to rupture independently of the main thrust faults. The Seattle fault zone earthquake history and structural relations demonstrate that not all backthrusts rupture in every large earthquake, but we argue that none of the shallow backthrusts will rupture without a moderate to large main-fault earthquake to drive the slip. The obvious implication is that paleoseismic studies of any shallow fault within a multistrand fault system will provide information about some, but not all, of the moderate to large earthquakes on the main faults. It is therefore only through study of all of the backthrusts that a paleoseismic record of earthquakes on the main fault system can be established (e.g., Nelson et al., 2014), but even this record may not be complete because some moderate earthquakes may not rupture to the surface and may have fault lengths that span only a part of the entire fault system.

CONCLUSIONS

Shallow, south-dipping thrust faults imaged on seismic-reflection profiles within the Seattle fault zone show that the Seattle fault forms a fault-propagation fold rather than a roof-thrust or wedge in which the main thrust faults lie deeper. Slip modeling also shows that the main Seattle fault zone thrust faults must reach shallow depths to reproduce the morphology of uplifted terraces and cause subsidence north of the fault; a roof-thrust model does not produce this subsidence or match the dip of the front of the terraces because the master fault lies too deep. The seismic-reflection profiles provide evidence that the northernmost thrust fault and synclinal axial surface forming the deformation front of the Seattle fault zone may project to the surface beneath the downtown area of Seattle, although near-surface geologic evidence consistent with this hypothesis is limited. If this location is verified, it raises the possibility that there could be ground tilting, surface rupture, or stronger ground motions than currently estimated in the downtown area. The back-thrust scarps and terrace uplifts provide evidence for at least five Seattle fault zone earthquakes in the past ~3500 yr that accommodated ~17 m of north-south shortening beneath the Puget Lowland. This cumulative shortening is about half the total expected in the Cascadia forearc, which confirms that over the past ~3.5 k.y., the Seattle fault has been one of the most active faults, if not the most active fault, in the region. Finally, kinematic modeling of Seattle fault zone earthquakes provides evidence that shallow back-thrust scarps formed in the forelimbs of major thrust faults most likely rupture in concert with earthquakes on the main faults, and the paleoseismic record provided by those back-thrust ruptures therefore reflects moderate or larger earthquakes on the main thrust faults.
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