Location, structure, and seismicity of the Seattle fault zone, Washington: Evidence from aeromagnetic anomalies, geologic mapping, and seismic-reflection data

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ABSTRACT

A high-resolution aeromagnetic survey of the Puget Lowland shows details of the Seattle fault zone, an active but largely concealed east-trending zone of reverse faulting at the southern margin of the Seattle basin. Three elongate, east-trending magnetic anomalies are associated with north-dipping Tertiary strata exposed in the hanging wall; the magnetic anomalies indicate where these strata continue beneath glacial deposits. The northernmost anomaly, a narrow, elongate magnetic high, precisely correlates with magnetic Miocene volcanic conglomerate. The middle anomaly, a broad magnetic low, correlates with thick, nonmagnetic Eocene and Oligocene marine and fluvial strata. The southern anomaly, a broad, complex magnetic high, correlates with Eocene volcanic and sedimentary rocks. This tripartite package of anomalies is especially clear over Bainbridge Island west of Seattle and over the region east of Lake Washington. Although attenuated in the intervening region, the pattern can be correlated with the mapped strike of beds following a northwest-striking anticline beneath Seattle. The aeromagnetic and geologic data define three main strands of the Seattle fault zone identified in marine seismic-reflection profiles to be subparallel to mapped bedrock trends over a distance of >50 km. The locus of faulting coincides with a diffuse zone of shallow crustal seismicity and the region of uplift produced by the M 7 Seattle earthquake of A.D. 900–930.

Keywords: aeromagnetic maps, Cascadia subduction zone, earthquakes, fault zones, Puget Sound, seismic reflection profiles.

INTRODUCTION

Earthquake hazards from shallow crustal faults are poorly understood in most of the Pacific Northwest (Yelin et al., 1994). Crustal earthquakes occur relatively infrequently and are difficult to relate to poorly mapped faults, yet geophysical surveys indicate that faults exist in the shallow subsurface beneath many of the densely populated regions of western Oregon and Washington (e.g., Johnson et al., 1994, 1996, 1999; Blakely et al., 1995, 2000).

Much of the Puget Lowland is covered by surficial deposits, water, and dense vegetation, and information about crustal faults (Fig. 1) has come largely from marine seismic-reflection profiling (Pratt et al., 1997; Johnson et al., 1994, 1996, 1999; Brocher et al., 2001; Calvert et al., 2001; T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston, 2001, personal commun.) and potential-field surveys (Yount and Gower, 1991; Gower et al., 1985). To improve our understanding of the crustal framework of this region, the U.S. Geological Survey conducted a high-resolution aeromagnetic survey of the entire Puget Lowland region (see Appendix). By using published interpretations of high-resolution seismic data to delimit the near-surface locations of fault strands beneath the major waterways, the aeromagnetic data detail the location, length, and subsurface geometry of the Seattle fault zone along the entire southern margin of the Seattle basin. This result is a critical step in improving models of crustal deformation used to estimate earthquake hazards in this densely populated urban area. Two rupture models for the fault zone are formulated on the basis of aeromagnetic and various geologic and geophysical data.

GEOLOGIC SETTING

Geologic mapping (Yount and Gower, 1991), seismic-reflection data (Johnson et al., 1994, 1999), and geophysical models (Pratt et al., 1997) are consistent with an interpretation that the Seattle fault zone consists of multiple east-trending, north-verging thrust faults. Motion on the fault zone has displaced Eocene volcanic and sedimentary bedrock northward relative to the deep, sediment-filled Seattle basin to the north (Fig. 1) (Johnson et al., 1994). Seismic-reflection data indicate >7 km of post-Eocene throw across the fault zone (Pratt et al., 1997; ten Brink et al., 1999). Johnson et al. (1994) inferred that the Seattle fault zone has been active from 40 Ma to the present and represents an east-trending transpressive zone transferring strain from right-lateral faults located southeast and northwest of the Seattle fault zone (Fig. 1). Despite these large offsets and a long history of deformation, the locations of the Seattle fault zone and its various

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GSA Bulletin; February 2002; v. 114; no. 2; p. 169–177; 6 figures.

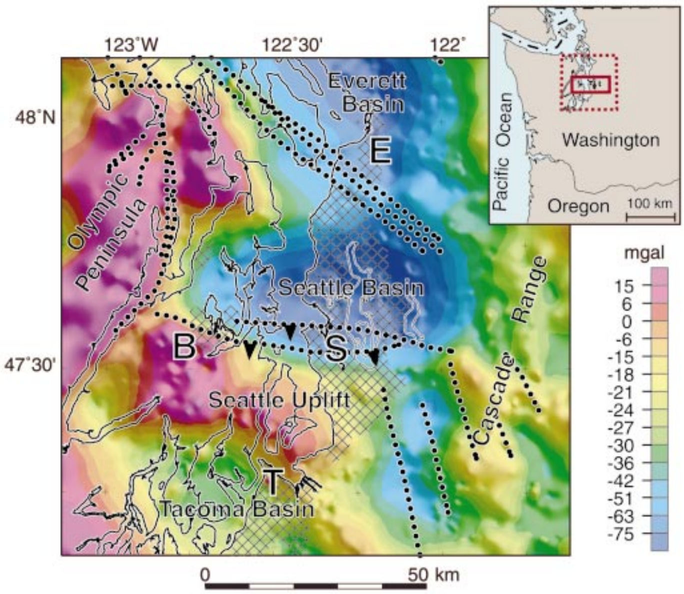


Figure 1. Regional setting. Large-scale map shows isostatic residual gravity over the Seattle basin and surrounding areas, where gravity lows reflect thick sections of basin-filling deposits. Faults generalized from Yount and Gower (1991) and Johnson et al. (1996). Crosshatch pattern indicates urbanized areas. S—Seattle, T—Tacoma, E—Everett, B—Bremerton. Inset: Dotted rectangle shows area of large-scale map of Figure 1; dashed rectangle indicates area of Figures 2, 3, 4, and 6.

strands have remained uncertain along most of its length.

The frontal fault of the Seattle fault zone was the likely source of a M 7 earthquake that occurred ~1100 yr ago (A.D. 900–930), causing tectonic uplift (Bucknam et al., 1992), landslides (Jacoby et al., 1992), and a local tsunami (Atwater and Moore, 1992). Uplift patterns from that earthquake are consistent with the south-side-up model for the fault. Field evidence from Bainbridge Island (Bucknam et al., 1992) indicates that the pre-uplift shoreline at Restoration Point, ~1.5 km south of Eagle Harbor, was 7 m lower than it is today, whereas the pre-uplift shoreline at a marsh near Winslow, on the north side of Eagle Harbor, was 1.5 m higher (R.C. Bucknam, 2001, written communication). Thus, near Eagle Harbor, an active strand of the Seattle fault zone must lie within narrow spatial limits near the topographic surface. A similar pattern is seen on the east side of Puget Sound: At Alki Point, south of the frontal fault, the pre-uplift shoreline was 6 m lower than it is today (R.C. Bucknam, 2001, written commun.), whereas at West Point north of the frontal fault, the pre-uplift shoreline was 3 m higher (Atwater and Moore, 1992). Considering that sea level has risen ~1 m since the uplift, net motion was dominantly south-side up both west and east of Puget Sound.

This relatively simple thrust-fault model is complicated by the recent discovery of an east-striking scarp on Bainbridge Island (Bucknam et al., 1999), referred to as the Toe Jam Hill scarp. Contrary to the long-term history on the Seattle fault zone, the topographic expression along the Toe Jam Hill scarp is consistent with a north-side-up fault, and recent geologic field evidence confirms this interpretation (Nelson et al., 1999). Moreover, a M 4.9 earthquake that occurred near Bremerton in 1997 had a focal mechanism also consistent with north-side-up movement (Weaver et al., 1999; T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston, 2001, personal commun.), and other relocated earthquake hypocenters throughout the area of the Seattle fault zone have components of north-side-up motion (T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston, 2001, personal commun.). The implications of the Toe Jam Hill fault and recent earthquakes are discussed subsequently.

The location of the deformation front and several thrusts in the Seattle fault zone are reasonably well determined in Puget Sound and other waterways by marine seismic-reflection studies (Pratt et al., 1997; Johnson et al., 1994, 1999). Geologic mapping (Yount and Gower, 1991) shows that the Tertiary strata in the hanging wall of the fault are dipping steeply to the north, and sparse outcrops can be traced eastward along strike for >50 km. However, the cover of young glacial deposits, water, and vegetation makes it difficult to map the precise location and configuration of the Seattle fault zone between the widely spaced seismic-reflection crossings, particularly beneath the highly developed regions of Seattle, Bremerton, and Bellevue. For these reasons, the Seattle area is an excellent candidate for high-resolution potential-field studies.

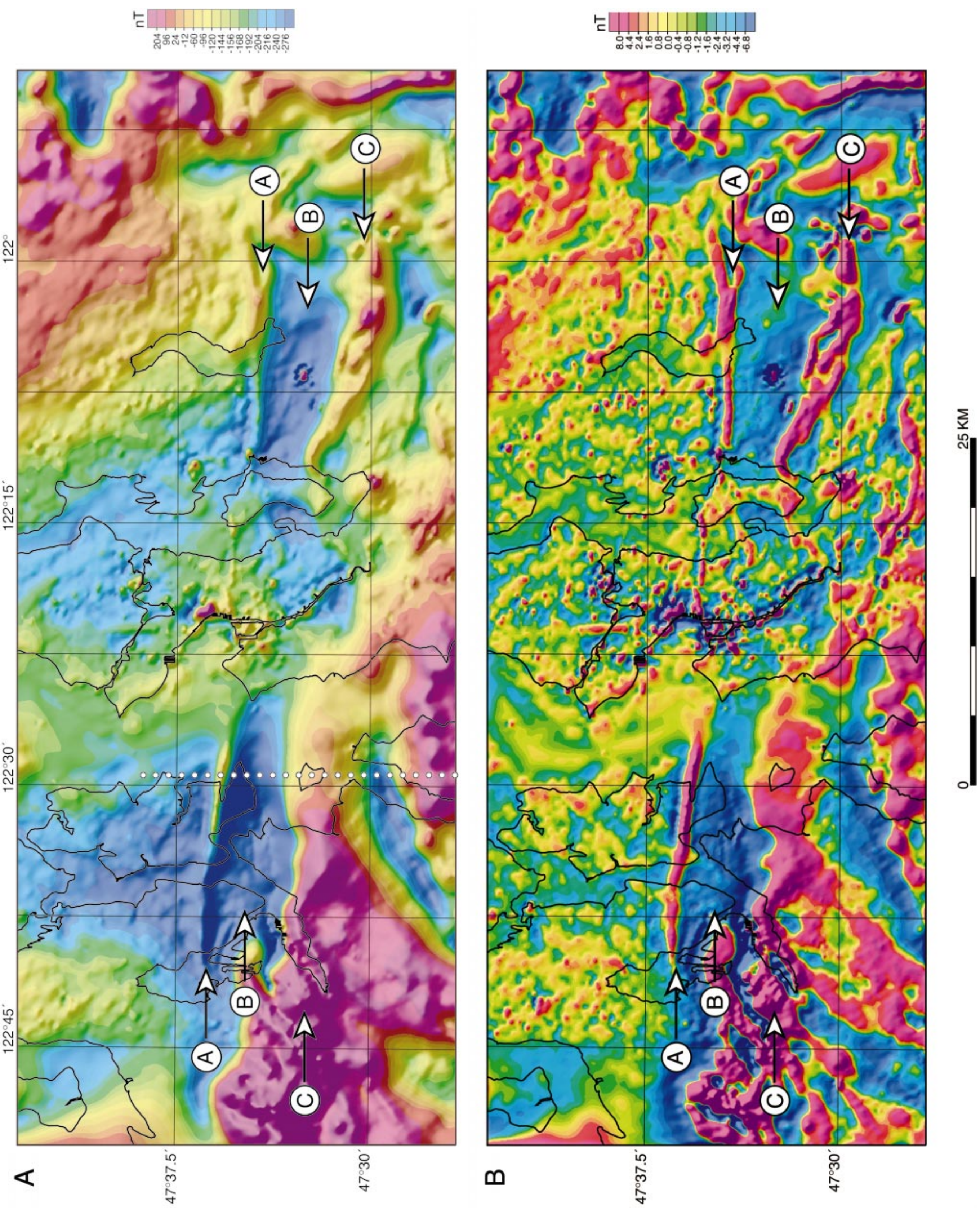
AEROMAGNETIC INTERPRETATION

The Seattle uplift (Fig. 1) is underlain at shallow depth by a complex package of Eocene and younger volcanic and sedimentary rocks. The contrasting magnetic properties of these rocks are ideal for aeromagnetic mapping of structures in the middle and upper crust. Along the Seattle fault zone, a distinctive pattern of magnetic anomalies follows the eastward trend of bedrock in the upthrown block and reliably reflects the underlying, steeply dipping stratigraphy.

Strands of the Seattle Fault Zone

Aeromagnetic data over the southern margin of the Seattle basin (Fig. 2) display a package of three east-trending magnetic anomalies. From north to south, they consist of an elongate, narrow magnetic high, a broad magnetic low, and a complex magnetic high; their east-west extent is >50 km. The northern anomaly (anomaly A in Fig. 2) is remarkably linear and narrow; it trends east from Dyes Inlet to Puget Sound and from Lake Washington to 10 km east of Lake Sammamish. On Bainbridge Island, this anomaly directly overlies a basalt conglomerate within the Miocene fluvial deposits of the Blakely Harbor Formation (Fulmer, 1975), which strikes east and dips 72°–80°N. East of Lake Washington, a similar anomaly also is caused by a steeply dipping Miocene volcanic conglomerate correlative with the Blakely Harbor Formation. These rocks were presumably deposited in the Se-

Figure 2. (A) Aeromagnetic anomalies over the Seattle fault zone and Seattle uplift. Letters A, B, and C indicate anomalies discussed in text. Dotted line shows location of magnetic profile (Fig. 5). (B) Aeromagnetic anomalies filtered in order to emphasize shallow magnetic sources. Data from A were continued upward 50 m, then subtracted from the original data.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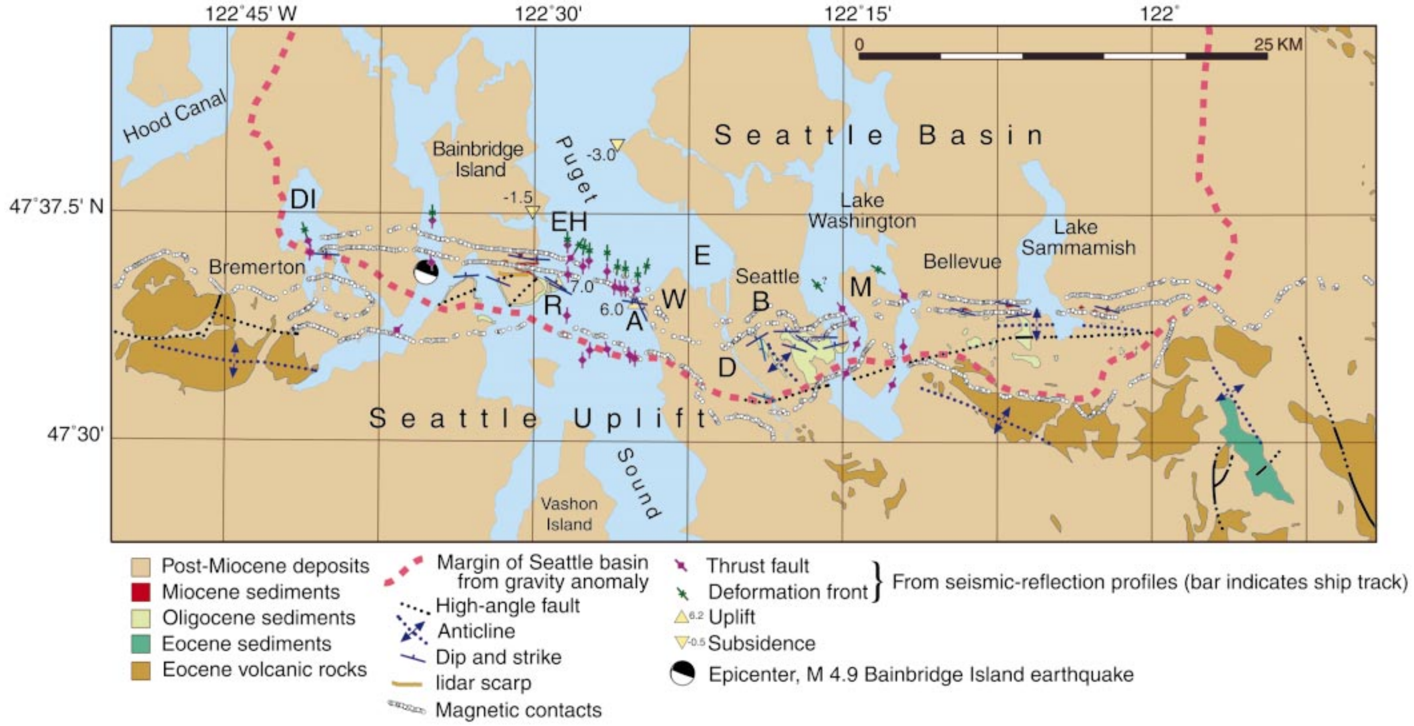
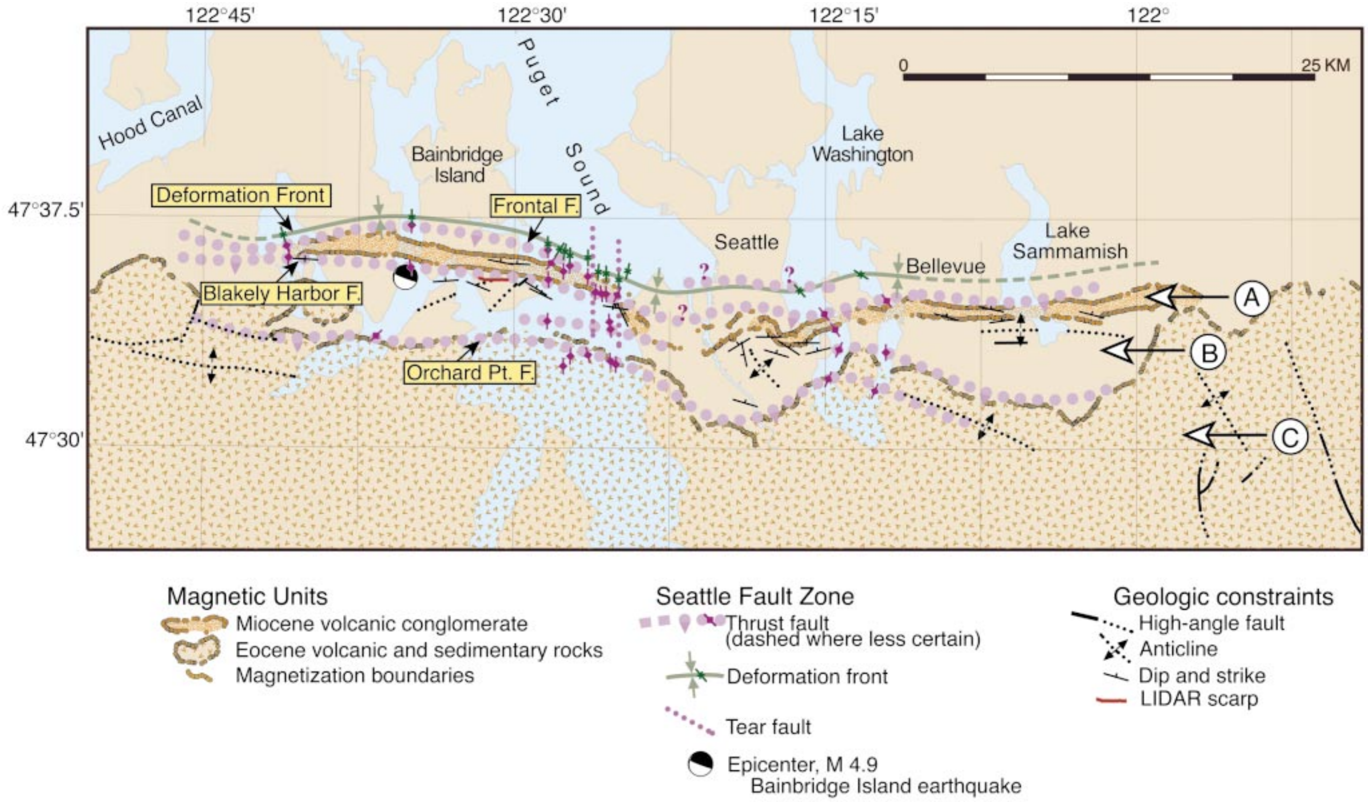


Figure 3. Geologic compilation of the area including the Seattle fault zone and Seattle uplift, simplified from Yount and Gower (1991), Frizzell et al. (1984), and Tabor et al. (1993). Seismic-reflection interpretation from Johnson et al. (1999). Uplift data from Atwater and Moore (1992) and R.C. Bucknam (2001, written commun.); values reported as elevation of pre-uplift (or pre-subsidence) shoreline relative to present mean highest high water (MHHW). White dots indicate magnetic contacts interpreted from aeromagnetic data and discussed in text. A—Alki Point, D—Duwamish River, B—Beacon Hill, W—West Seattle, M—Mercer Island, E—Elliott Bay, R—Restoration Point, EH—Eagle Harbor, DI—Dyes Inlet.



attle basin and subsequently incorporated into the hanging wall as the Seattle fault zone propagated northward (Johnson et al., 1999).

The central magnetic low (anomaly B in Fig. 2) west of Seattle correlates with upper Eocene and Oligocene sedimentary strata of the Blakeley Formation (Fulmer, 1975). East of Seattle, the anomaly overlies undifferentiated marine and nonmarine sedimentary rocks, also late Eocene and Oligocene in age (Yount and Gower, 1991). These strata have very low magnetic susceptibility, as measured in the field with a hand-held susceptibility meter, and are intermittently exposed between Bremerton and Lake Sammamish.

The southern magnetic high (anomaly C in Fig. 2) consists of high-amplitude, steep-gradient anomalies that overlie deformed Eocene volcanic rocks in the hanging-wall block. Magnetic anomalies at the western end are caused by basaltic basement of the Crescent Formation, whereas to the east, the high-amplitude anomalies are caused by Cascade-derived andesitic rocks of the Tukwila Formation (Yount and Gower, 1991).

This tripartite package is well expressed east and west of Seattle. The intervening region, between Alki Point and Lake Washington, coincides with folds that are transverse to the Seattle fault zone (Yount and Gower, 1991). Anomalies due to geologic sources are subdued in this region, and high-amplitude anomalies from cultural sources add complexity to the magnetic pattern over Seattle. Nevertheless, the aeromagnetic triplet, as a package, can be traced beneath Puget Sound, downtown Seattle, and Lake Washington, where the sinuous shape of the aeromagnetic anomalies follows the mapped strike of beds, and both anomalies and bedding sweep around a mapped northwest-striking anticline east of the Duwamish River (Fig. 3). The southward sweep of magnetic anomalies south of Seattle is mimicked by mid- and upper-crustal seismic velocities (Brocher et al., 2001; T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston, 2001, personal commun.), indicating that this sinuous shape is a fundamental aspect of the hanging wall.

Computer-picked magnetic contacts (Blakely and Simpson, 1986) were used to divide the hanging wall into three magnetic lithologies located at shallow depth (Fig. 4): a thin strip of Miocene volcanic conglomerate extending eastward from Bremerton to beyond Lake Sammamish (anomaly A), Eocene volcanic rocks to the south (anomaly C), and nonmagnetic Eocene to Oligocene sedimentary rocks in the intervening gap (anomaly B). The entire package dips steeply northward, presumably offset at depth by the Seattle fault zone. We note that Miocene conglomerate does not crop out in the Seattle area on Beacon Hill where anomaly A is curved and has lower amplitude. Anomalies may be subdued in this region because the conglomerate has been forced to deeper depths or has been uplifted and eroded by the folding event related to the anticline. Despite the presence of the anticline, there is still good agreement between the geologic strikes and the strikes of the magnetic contacts over Beacon Hill (Fig. 3). Alternatively, the Miocene conglomerate deposited in this area may still be in the Seattle basin and not yet incorporated into the hanging-wall block. As discussed subsequently, this latter interpretation would imply along-strike variation in the northward propagation of the fault zone since the conglomerate was deposited in the Miocene (Fulmer, 1975).

The location of the Seattle fault zone and its various strands must be consistent with the long-distance continuity of structure and post-Eocene stratigraphy as indicated by the geology and magnetic data. In Figure 4, we have tried to ‘‘connect the dots’’ provided by seismic-reflection crossings of the fault zone in the waterways (Johnson et al., 1999), while using magnetic contacts and geologic mapping to guide us in the intervening areas. For example, the shape and continuity of anomaly A between Dyes Inlet and Alki Point shows that the Miocene conglomerate is steeply dipping and continuous in this area. Thus, east-trending faults observed in marine seismic-reflection crossings immediately north and south of the conglomerate in Dyes Inlet, west of Bainbridge Island, and in Puget Sound, cannot significantly offset the conglomerate, and we presume, therefore, that these faults have similar continuity between Dyes Inlet and Alki Point.

The deformation front, or hinge line, where Seattle basin strata are dragged upward by the frontal thrust fault, lies north of the steeply dipping magnetic conglomerate in the upthrown block. The frontal fault, as located by the reflection data, lies between the deformation front and the conglomerate unit, locally coinciding with its upper contact. The frontal fault passes through Eagle Harbor, near the northern limit of uplift associated with the M 7 earthquake ~1100 yr ago (A.D. 900–930) (Bucknam et al., 1992). The Blakely Harbor fault lies south of the conglomerate unit and locally coincides with its southern contact. The Orchard Point fault follows the contact between Tertiary sediments and an uplifted block of volcanic rocks to the south.

Figure 4. Interpretation of aeromagnetic, generalized geology, and seismic-reflection data of the area including the Seattle fault zone and Seattle uplift. Stipple patterns show interpreted magnetic terranes. Faults are indicated by purple lines, and the deformation front by a green line. Letters A, B, and C indicate anomalies discussed in text.

Is the Seattle Fault Segmented?

There is evidence from high-resolution seismic-reflection data (Johnson et al., 1999) for two north-trending tear faults in Puget Sound that offset the Seattle fault zone by several kilometers. According to Johnson et al. (1999), the tear faults cross the Miocene conglomerate between Bainbridge Island and Alki Point. Significant lateral displacement on either fault should be evident as lateral displacements of anomaly A. We have noted two points along anomaly A where right-lateral offsets are permissible (Fig. 4), but the strong continuity of anomaly A (Fig. 2) limits the amount of offset on any crosscutting faults to no more than ~1 km at this location.

The subdued nature of anomaly A beneath Seattle may relate directly to the easternmost of the proposed tear faults. The frontal fault west of Seattle has propagated northward since deposition of the Miocene conglomerate, incorporating the conglomerate into the hanging wall in the process. The frontal fault beneath Seattle, however, may not have propagated northward to the same degree, leaving the conglomerate still deep within the Seattle basin. The easternmost tear fault may mark a discordance in fault propagation.

The possibility of north-south tear faults, the mapped anticline along the Duwamish River, the sinuous nature of the magnetic anomalies between Alki Point and Lake Sammamish, and the subdued nature of anomaly A beneath Seattle all imply a degree of along-strike variation in the Seattle fault zone. Taken as a whole, we do not consider this along-strike variation to fundamentally segment the frontal fault. The M 7 earthquake that occurred ~1100 yr ago on the frontal fault uplifted the hanging wall by 7 m both east and west of the tear faults. This defining event indicates in dramatic fashion that segmentation along the Seattle fault zone does not preclude the possibility of large earthquakes in the future.

We also note a narrow anomaly extending

eastward from Elliot Bay to northern Mercer Island (Fig. 2). Part of this anomaly overlies the Interstate-90 bridge over Lake Washington and may be caused by the bridge itself, although other bridges and freeways in the area do not produce similar anomalies. Harding et al. (1988) noted a change in the dip of seismic reflectors imaged in multichannel seismic-reflection data along the west side of Mercer Island and interpreted the change in dip as indicating a possible Holocene fault located 200 m south of the bridge. Recent examination of these data indicates that vertical offset of these reflectors could not be greater than ~10 m (T. Pratt, 2000, oral commun.). We show the aeromagnetic anomaly as a queried fault in Figure 4, but additional field investigations are needed to evaluate this possibility.

SEISMICITY AND FAULT GEOMETRY

On June 23, 1997, a M 4.9 earthquake occurred ~12 km west of downtown Seattle, just west of Bainbridge Island (Weaver et al., 1999; T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston, 2001, personal commun.). The earthquake occurred directly beneath the inferred surface trace of the Blakely Harbor fault (Fig. 4). The depth of the main shock was initially estimated at 7 km (Weaver et al., 1999), but recent hypocenter relocations by T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston (2001, personal commun.) indicate a depth of 11.5 km. The focal mechanism (Fig. 4) indicated either north-side-up motion on an approximately vertical east-trending surface, or northward displacement on a nearly horizontal surface. Aftershocks followed above the main-shock hypocenter, suggesting that the former solution is more likely, but neither solution is consistent with the long-term history of the Seattle fault zone, which would predict south-side-up motion on the frontal fault.

However, the north-side-up solution for the 1997 earthquake was consistent with the morphology of a Holocene fault scarp visible on a lidar image of Bainbridge Island (Bucknam et al., 1999; Nelson et al., 1999) (Figs. 3 and 4), which shows up-to-the-north morphology and lies approximately within the surface projection of the preferred plane of the earthquake focal mechanism. The Toe Jam Hill scarp, as it is referred to, is ~2 km in length, ranges from 1.5 to 10 m in height (Bucknam et al., 1999), and is parallel to the Blakely Harbor fault. The north-side-up morphology of the Toe Jam Hill fault scarp is consistent with formation along a back thrust (Nelson et al., 1999), possibly the southern margin of a "pop-up" formed by bending of the hanging-wall block as it rides up the frontal fault from a deeper detachment.

Taken together, the Bainbridge Island earthquake sequence and the Toe Jam Hill scarp suggest complexity in the spatial relationship between the frontal fault, the Blakely Harbor fault, and earthquakes responsible for the Holocene scarp. We have modeled magnetic anomalies along the east shore of Bainbridge Island to help determine the geometry of these faults (Fig. 5). Recent interpretations of seismic-reflection data (Pratt et al., 1997; ten Brink et al., 1999; Johnson et al., 1999; Brocher et al., 2001; Calvert et al., 2001) from this part of Puget Sound all agree that strands of the Seattle fault zone dip southward, with dips ranging from 40° (ten Brink et al., 1999) to 60° (Johnson et al., 1999; Calvert et al., 2001). This range of southward dip was used to set dip limits on all north-verging faults in Figure 5. The volcanic conglomerate was assigned a dip of 74°N and an up-section thickness of 1 km on the basis of geologic field mapping (Yount and Gower, 1991). The conglomerate was also assigned a magnetization of 0.8 A/m, the maximum allowable according to 35 field measurements using a hand-held susceptibility meter. The Crescent Formation was assigned a magnetization of 1.7 A/m, consistent with published values for Coast Range basalt elsewhere (0.1–3.5 A/m from Bromery and Snavely [1964], 2.75 A/m from Finn [1990], and 3 A/m from Pratt et al. [1997]).

Figures 5A and 5B show the observed magnetic anomaly and a permissible magnetic model, respectively. Figures 5C and 5D show two interpretations based on the magnetic model. Both include three thrust faults, in accordance with Figure 4, and predict that the frontal fault has offset the Miocene conglomerate ~1 km at a depth of ~1.5 km. The Eocene Crescent Formation is multiply offset, with a total vertical throw of ~7 km at this location, in agreement with seismic interpretations (ten Brink et al., 1999; Pratt et al., 1997). In both interpretations, the 1997 Bainbridge Island earthquake occurred in the footwall of the frontal fault.

The length and height of the Holocene scarp suggest that the earthquake(s) that caused the scarp would have been larger than M 6 (Hemphill-Haley and Weldon, 1999) and, if the rupture surface were near vertical, it may have ruptured into the footwall of the frontal fault. The spatial relationship and shared sense of motion between the Holocene Toe Jam Hill scarp and the 1997 Bainbridge Island earthquakes support an argument that a steeply dipping fault system might have been responsible for both the scarp and the earthquake (Fig. 5C). The 1997 earthquake, however, was only M 4.9, and the resulting rupture surface probably did not reach the topographic surface and may not have cut the frontal fault. Kinematically relating the Toe Jam Hill scarp to the 1997 earthquake remains problematic.

Alternatively, the Toe Jam Hill scarp on Bainbridge Island may represent a back thrust to the main strand of the Seattle fault zone (Fig. 5D). In this model, the scarp reflects movement on the frontal fault and is related to the long-term slip rate of the Seattle fault zone. This model implies that the 1997 Bainbridge Island earthquake and aftershocks are not directly related to the Toe Jam Hill scarp but rather occurred within the lower plate of the frontal fault, perhaps on a fault stressed by the advancing wedge above the thrust sheet (Fig. 5D).

The focal mechanism of the 1997 Bainbridge Island earthquake is also consistent with a nearly horizontal plane of slip, with the upper block having moved northward relative to the lower block. The zone of aftershocks above the main shock may reflect deformation along an active axial surface in the bending, overriding block, in the manner described by Shaw and Suppe (1996) for blind thrusts in the Los Angeles basin.

The first model (Fig. 5C) has the advantage of kinematically linking the Toe Jam Hill scarp to the 1997 Bainbridge Island earthquakes, whereas the second (Fig. 5D) has the potential to quantitatively relate paleoseismic slip rates to crustal structure of the Seattle fault zone. The different tectonic models describe possible fault rupture dimensions and kinematic links with important implications for evaluating the location, potential magnitude, and recurrence rates of moderate to large earthquakes on the Seattle fault zone.

The 1997 Bainbridge Island earthquake was representative of a general pattern of crustal earthquakes within the Seattle fault zone. Figure 6A shows epicenters of upper-plate earthquakes of M 1.5 or larger occurring in the vicinity of the Seattle fault zone since January 1980. Figure 6B shows hypocenters of selected earthquakes projected onto a vertical plane that lies normal to the fault zone, taking into account the sinuous trace of the fault zone. Most earthquakes in this part of the Puget Sound occur deeper than ~15 km (emphasized by the dashed line in Fig. 6B). On the basis of a larger set of relocated hypocenters, T.M. Van Wagoner, R.S. Crosson, K.C. Creager, G. Medema, and L. Preston (2001, personal commun.)

found that nearly 60% of hypocenters fall between 15 and 25 km, with a mean depth of 17.6 km, placing most crustal earthquakes in the Seattle region within Crescent Formation basement. Within the Seattle fault zone, hypocenters depart significantly from this depth range, forming a near-vertical zone that extends well above the 15 km depth to near the topographic surface. These shallow earthquakes do not lie along the thrust fault predicted from our magnetic model, nor along thrust faults described in earlier studies (e.g., Johnson et al., 1994; Pratt et al., 1997; Brocher et al., 2001). This recent seismic activity may reflect one or more basin-loading faults stressed by the advancing wedge of the thrust sheet, or deformation within an upper plate that moves along a nearly horizontal plane of slip. In any case, the pattern of recent seismicity appears to be a departure from the long-term deformational history of the Seattle fault zone.

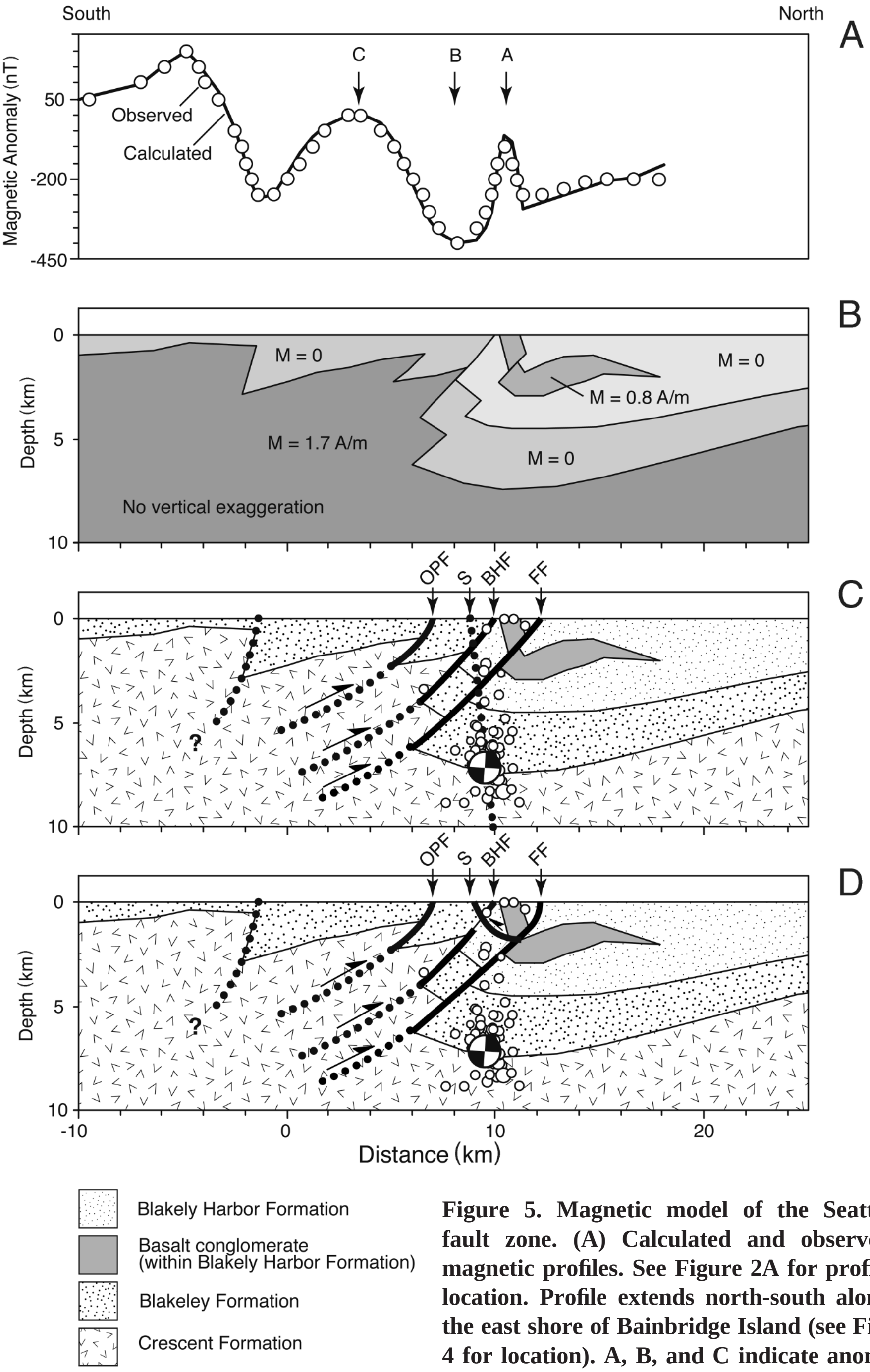


Figure 5. Magnetic model of the Seattle fault zone. (A) Calculated and observed magnetic profiles. See Figure 2A for profile location. Profile extends north-south along the east shore of Bainbridge Island (see Fig. 4 for location). A, B, and C indicate anomalies discussed in text. (B) Magnetic model. Sources are assumed to be infinitely extended in the east and west directions. Dip of frontal fault (50°S) determined from seismic-reflection data (ten Brink et al., 1999; Johnson et al., 1999). Dip (74°N) of Miocene conglomerate (labeled by its magnetization $M = 0.8$ A/m) from geologic mapping (Yount and Gower, 1991). The Crescent Formation is the unit with a magnetization of 1.7 A/m. (C) Interpretation favoring a structural connection between Bainbridge Island earthquakes and Holocene scarp. OPF—Orchard Point fault, BHF—Blakely Harbor fault, FF—frontal fault, S—Holocene scarp seen in lidar topographic data. (D) Interpretation favoring the Holocene scarp as a back thrust to the main thrust sheet. First-motion solution and earthquake locations from Weaver et al. (1999).

SUMMARY

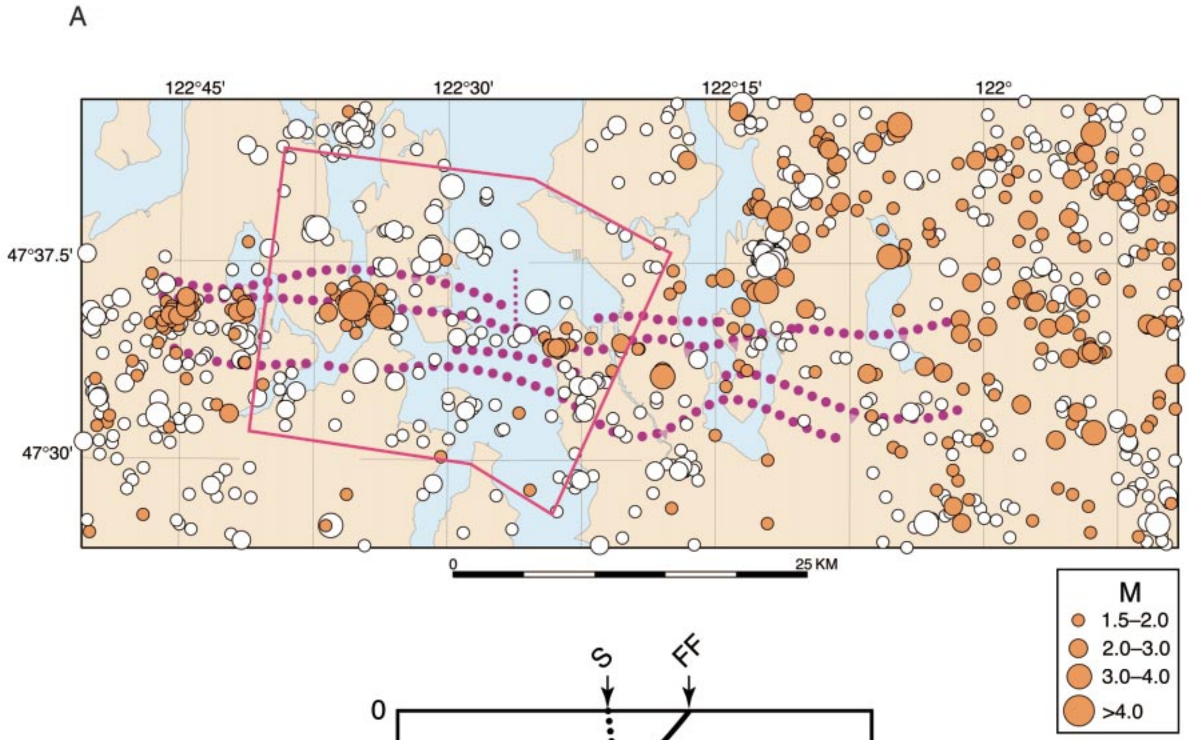
New, high-resolution aeromagnetic data provide constraints on the location, length, and geometry of the Seattle fault zone. The correlation of aeromagnetic anomalies with tilted upthrown-block stratigraphy defines the location of the Seattle fault zone within narrow limits over a distance of 50 km. Thus determined, the fault zone on Bainbridge Island coincides with recent seismicity, a postglacial fault scarp, and the M 7 earthquake that occurred ~1100 yr ago (A.D. 900–930) on the Seattle fault zone. The details of the relationship between earthquake sources in the footwall and those in the hanging wall and the role of the frontal fault at depth remains to be resolved, as well as the kinematic relationships among current earthquakes, paleoseismic evidence, and fault geometry.

APPENDIX

The aeromagnetic survey (Blakely et al., 1999) was flown along north-south lines spaced 400 m apart and along east-west control lines spaced 8 km apart. Flight altitude was 250 m above terrain, or as low as permitted by safety considerations. A theoretical flight surface, based on a digital topographic model, was computed in advance of the survey, and real-time, differentially corrected Global Positioning System (GPS) navigation was used during flight to maintain the desired flight surface. Two ground-based magnetometers were used to monitor and correct for time-varying magnetic fields. Total-field anomalies were computed based on the International Geomagnetic Reference Field updated to the date of the survey.

ACKNOWLEDGMENTS

We are grateful to Ralph Haugerud, Tom Pratt, Brian Sherrod, Derek Booth, Kathy Troost, Tom



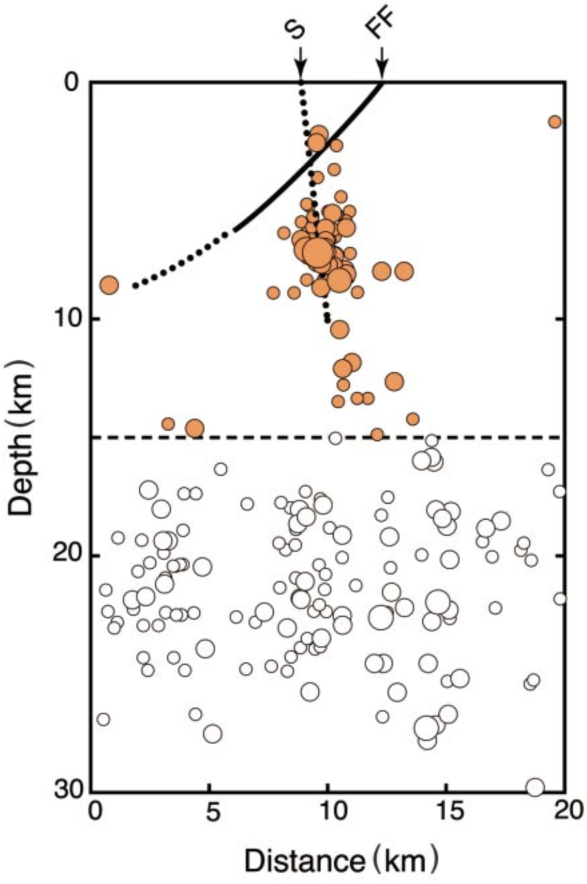


Figure 6. Upper-plate earthquakes of the Seattle fault zone. (A) Epicenters of M 1.5 and larger earthquakes occurring since January 1980 are shown by circles, where size of circle indicates relative magnitude. Dotted purple line shows predicted traces of the Seattle fault zone, taken from Figure 4. (B) Cross section across the Seattle fault zone. Earthquakes within pink box of Figure 6A projected parallel to fault zone. S—Holocene scarp seen in lidar topographic data, FF—frontal fault (see also Fig. 5, C and D, for relationship to other faults). The dashed line at 15 km emphasizes the fact that most hypocenters are below this depth.

Brocher, and Uri ten Brink for discussions that helped formulate many of the ideas in this paper. Early reviews by Bob Jachens and Tom Brocher and later reviews by Bob Crosson, Silvio Pezzopane, Peter La Femina, and Chuck Connor were particularly helpful.

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Manuscript Received by the Society December 26, 2000
Revised Manuscript Received August 9, 2001
Manuscript Accepted August 21, 2001

Printed in the USA