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To provide context for the Hanford PSHA and potential seismic sources, this chapter provides a summary of the tectonic setting of the Hanford Site region. Section 4.1 addresses the tectonic setting and geologic history. Section 4.2 addresses contemporary plate motions and the tectonic stress regime, Section 4.3 addresses late Cenozoic and Quaternary history, and Section 4.4 provides a summary of local and regional earthquakes.

### 4.1 Tectonic Setting

The Hanford Site is located in the backarc region of an obliquely convergent plate boundary (Figure 4.1). The oceanic Juan de Fuca plate presently is under-thrusting northern California and western Oregon and Washington along the east-dipping Cascadia subduction zone (Atwater 1970). Modern plate-motion models such as MORVEL (DeMets et al. 2010) predict that the Juan de Fuca plate is subducting obliquely northeast beneath coastal Washington at an average convergence rate of about 37 mm/yr relative to stable North America. Arc magmatism above the subduction zone is responsible for the development of the approximately north-south-trending Cascade volcanic chain east of the trench and forearc region (Priest 1990).

![Figure 4.1](image-url)
The modern Cascadia convergent margin in the Pacific Northwest formed after accretion of the Siletzia block or terrane to the ancestral North American continental margin approximately 51 Ma (Wells et al. 2010). Siletzia is interpreted to be a volcanic edifice that developed on oceanic crust adjacent to western North America in early Tertiary time. Siletzia alternatively is thought to have formed primarily offshore and was accreted following collision with an ancestral subduction zone located beneath what is now eastern Washington, Oregon, and Idaho (Simpson and Cox 1977; Duncan 1982; Gao et al. 2011); or it formed adjacent to North America through a rifting mechanism and subsequently became involved in the subduction process (Wells et al. 1984; Babcock et al. 1992). Once Siletzia entered the ancestral trench, subduction stalled and effectively accreted the volcanic edifice to North America. The convergent boundary subsequently stepped westward to form the modern Cascadia subduction zone (Gao et al. 2011 and references therein). Arc magmatism associated with the new subduction zone began about 42 Ma, initiating development of the ancestral Cascade Range (Babcock et al. 1992).

Following establishment of the modern Cascadia subduction zone and nascent volcanic arc, the later Tertiary geologic history of eastern Washington was dominated by eruption and deposition of the Columbia River flood basalts in middle to late Miocene time. The Columbia River Basalts (CRBs) compose one of the largest flood basalts accumulations on the Earth’s surface, covering more than 200,000 km² in Washington and Oregon and western Idaho, and representing a total volume of about 234,000 km³ (Camp et al. 2003). Although the full stratigraphic thickness of the CRBs was deposited over a period between about 17–6 Ma, the vast majority of the basalts, by volume, erupted in the first several million years (Barrash et al. 1983). The thickness of the CRBs in eastern Washington ranges between 3 and 6 km (Catchings and Mooney 1988), and is about 4.6 km (15,000 ft) thick in the vicinity of the Hanford Site (Burns et al. 2011 and references therein).

Due to the extensive and deep burial of older rocks by the Miocene basalt flows, the Eocene and Oligocene geology of eastern Washington between the accretion of Siletzia and the eruption of the CRBs is not known in detail (Campbell 1989).

Several classes of models have been proposed for the origin of the CRB eruptions (see Camp and Hanan [2008] and Gao et al. [2011] for a summary and additional references). One group of models posits that the volcanism was triggered when a mantle plume that eventually became the Yellowstone hot spot impinged on the base of the lithosphere in the vicinity of the present triple intersection of the Washington, Oregon, and Idaho state borders. Another group of models invoke backarc extensional processes to explain the CRB eruptions. A third class of models proposes that melt generation and magmatism were triggered by foundering or delamination of the lower lithosphere. Camp and Hanan (2008) favor a hybrid model in which a rising mantle plume caused delamination and partial melting of the lithosphere. Recently, Liu and Stegman (2012) proposed that a “tear” in the subducting Farallon slab beneath eastern Oregon facilitated asthenospheric upwelling and thermal erosion of the slab, effectively creating a “slab gap” and triggering partial melting of the upper mantle that resulted in the CRB eruptions.

Although eruption and deposition of the CRB profoundly affected the stratigraphy and topography of eastern Washington, the models summarized above do not directly predict creation of accommodation space for the accumulation of 3 km to 6 km of basalt flows in the central Columbia Plateau. The majority of CRB flows erupted from fissures in and near the Wallowa Mountains in northern Oregon (Camp and Hanan 2008). The northern extent of the Chief Joseph dike swarm, which was the source of the most voluminous CRB flows (Camp and Hanan 2008), is about 100 km to the east-southeast of the Columbia Plateau. In the delamination model for the triggering of CRB eruptions cited above, subsidence of areas.
surrounding the source region could be produced by destabilization of mantle lithosphere and formation of mantle “drips” that exert a downward load on the overlying lithosphere (Camp and Hanan 2008). The magnitude of subsidence predicted by numerical models of this process is much less than the observed 3 km to 6 km thickness of the CRB in the Columbia Plateau, and it does not extend greater than 100 km from the source region (Camp and Hanan 2008). Other processes more local to the central Columbia Plateau may have been responsible for accommodating deposition of 3 km to 6 km of CRB in eastern Washington.

The CRB flows in eastern Washington are deformed in a series of generally east-west-trending anticlines that are known collectively as the Yakima Fold Belt (YFB) (Reidel et al. 1994). Note that in this document the term Yakima Fold Belt or YFB refers to the tectonic province that includes the Yakima folds; the term Yakima Fold and Thrust Belt (YFTB) is reserved exclusively to indicate the YFTB seismic source zone in the SSC model (Section 8.3). The structures of the YFB dominate the post-CRB tectonics and topography in eastern Washington. The Yakima folds are asymmetric, generally north-vergent anticlines that have accommodated approximately north-south shortening. In many cases, north-vergent thrust faults are mapped at or near the base of the forelimbs of the folds (Schuster 2005), suggesting that the structures at least initially developed as fault-propagation folds above blind faults that subsequently ruptured to the surface.

The YFB is transected by the Olympic-Wallowa lineament (OWL), a major northwest-southeast topographic lineament that, as originally defined, extends from the northern Olympic Peninsula in northwestern Washington across the Columbia Basin and into northeastern Oregon (Raisz 1945) (Figure 4.2). Shorter, discrete structural trends have been inferred along the general OWL topographic lineament within the YFB, including the Cle Elum-Wallula deformed zone (CLEW [Cle Elum-Wallula lineament]), and the Rattlesnake-Wallula (RAW) alignment (see Reidel et al. 1994 and references therein for a summary). The CLEW is a 10-km-wide zone encompassing anticlines of the YFB that trend approximately N50°W, in contrast to the east-west to east-northeast trends of Yakima folds outside the CLEW. The RAW alignment is a northwest-trending alignment of structures that includes the eastern, southeast-trending part Rattlesnake Mountain, and continues along trend to the southeast to join the Wallula fault zone. Specific structures of the CLEW include Manastash, Umtanum, and Yakima Ridges, Rattlesnake Mountain, and a northwest-trending alignment of doubly plunging anticlines between Rattlesnake Mountain and Wallula Gap that is commonly included as part of the RAW alignment (Reidel et al. 1994) (Figure 4.3).

In the vicinity of the Hanford Site and much of the central Columbia Basin, the 10.5 Ma Elephant Mountain Member of the Saddle Mountains Basalt is fully involved in the uplift and folding of major YFB structures such as the Rattlesnake Mountain and Saddle Mountain anticlines (Reidel et al. 1994). As discussed in Appendix E, the structural relief on the Elephant Mountain Member across the Saddle Mountain anticline is comparable to the present topographic relief, indicating that the present tectonic topography has developed in the past 10.5 million years. Reidel (1984) interpreted borehole data in the vicinity of Saddle Mountain to show that this structure was growing continuously during eruption of the CRB flows, and that the rate of fold growth decreased significantly close to the end of the CRB eruptions. Late Miocene and Pliocene deposits of the Ringold Formation in the vicinity of the Hanford Site also are involved in growth of the Saddle Mountain anticline and other Yakima folds. Geologic data discussed in Section 8.4.3.5 support the hypothesis that folding of the 10.5 Ma Elephant Mountain basalt at the Saddle Mountains and on the Hanford Site began prior to or during deposition of the Ringold Formation, indicating a Miocene or Pliocene age for the onset of deformation.
As discussed in Section 8.4.3.5, a Neogene age for the onset of folding is consistent with late Miocene changes in Pacific-North American plate motion that likely increased the rate of clockwise rotation of crustal blocks in the Pacific Northwest, which in turn is linked to north-south shortening in eastern Washington (McCaffrey et al. 2013). Work performed for this study has documented middle to late Pleistocene surface rupture along an emergent north-vergent thrust fault at the base of Rattlesnake Mountain, south of the Hanford Site (see Appendix E), thus providing local evidence for relatively young tectonic shortening in the YFB.

Two classes of models have been proposed for the downdip extent and geometry of thrust faults beneath the YFB anticlines. One class of models assumes that the thrust faults pass through the CRB and terminate downward at the contact between the CRB and underlying Paleogene and older rocks, which acts as a mechanical detachment for the folding (e.g., Figure 4 of Reidel et al. [1994]). This class of models is variously referred to as “thin-skin” deformation because the folding is confined to the CRB and thus the upper 2−5 km of the crust; or “uncoupled” (per Geomatrix 1996) because the folding does not involve the crystalline basement (i.e., it is “uncoupled” from the basement). The other class of models assumes that the thrust faults underlying the folds extend through the lower contact of the CRB, into the pre-CRB sedimentary and crystalline rocks, and terminate as discrete brittle structures at the base of the seismogenic crust. This class of models is variously referred to as “thick-skin” because the folding involves the entire thickness of the seismogenic crust, or “coupled” (Geomatrix 1996) because the thrust faults extend into and are “coupled” with the crystalline basement.
In a previous PHSA for the Hanford Site, Geomatrix (1996) included the “coupled” (thick-skin) and “uncoupled” (thin-skin) models in the source model as weighted branches on a logic tree. In the source characterization, groups of fold structures in the YFB were interpreted to share common structural features, and thus assigned common coupling values. For the majority of YFB structures, Geomatrix (1996) assessed a low probability to the coupled or thick-skin model [0.15−0.3] based on contemporary proponent models for a thin-skin or “uncoupled” origin of the YFB (e.g., Reidel et al. 1994), an assumed strength contrast between the CRB and underlying sediments interpreted to favor detachment and thin-skin deformation, and an apparent concentration of seismicity within the CRB suggesting that deformation is localized there relative to deeper rocks. Geomatrix (1996) also argued that rupture of thick-skin or “coupled” faults would likely involve the full thickness of the seismogenic crust, and thus be more likely to have preserved geomorphic evidence of surface rupturing. For example, a high weight [0.95] was assigned to the coupling model for the Toppenish Ridge anticline based on paleoseismic evidence for large individual-event displacements (Geomatrix 1996). Because similar relationships were not documented along most YFB structures, Geomatrix (1996) adopted a low weight for the coupled model in those cases.
In a review of the Geomatrix (1996) source characterization, Zachariasen et al. (2006) questioned whether sufficient data exist to provide a basis for strongly favoring either the thin-skin or thick-skin model. Zachariasen et al. (2006) developed a list of arguments in favor of the thick-skin model, including the following:

- The orientations of the Yakima folds along the CLEW are strongly influenced by the OWL, which is thought to be a deep-seated structure in the basement given its great lateral extent. Specifically, Manastash Ridge, which is one of the CLEW structures, is parallel to sub-basalt structures in the basement (Tabor et al. 1984, as cited by Zachariasen et al. 2006). According to Zachariasen et al. (2006), the influence of the OWL on the Yakima folds argues against the "uncoupled" model, in which the folds are detached well above the basement and thus have no mechanical connection or communication with basement structures.

- The Yakima fold structures are observed to locally continue beyond the limits of the CRB into basement rocks of the Cascades, arguing against thin-skin models where the folds are confined to the CRB.

- Models for growth of the Yakima folds during eruptions of the CRB, including folding during deposition of the early Grande Ronde flows (e.g., Reidel 1984), are inconsistent with the mechanical assumptions that the entire fold belt is detached above a strength contrast between the rigid basalt and underlying sedimentary rocks. Zachariasen et al. (2006) argue that syn-depositional folding during eruption and emplacement of the CRB is better explained by faults that extend below the CRB and into the underlying rocks.

- Correlation of the sub-basalt sediments encountered in deep boreholes with their exposed equivalents bordering the Columbia Plateau indicates that they are consolidated volcanic rocks, sandstones, and shales (Campbell 1989). Sonic logs from the deep Yakima Minerals 1–33 and Biss 1–29 exploration wells in the YFB (data in Czajkowski et al. 2012) document that the sub-basalt sediments have P-wave velocities ranging from about 3.8 – 4.1 km/s, which are typical of consolidated sandstones and shales (see Table 9-4 in Czajkowski et al. 2012). For comparison, P-wave velocities of unconsolidated sediments generally are less than about 2.0 km/s (Table 9-5 in Press [1966]). Although the P-wave velocity of the CRB in the wells is higher (about 5.5 km/s; Czajkowski et al. 2012) than that of the underlying sedimentary rocks, the velocities of the sub-basalt strata are greater than the reported P-wave velocity of concrete (about 3.6 km/s; Table 9-7 in Press [1966]), which indicates that they likely have substantial mechanical strength. Based on these data, there is no evidence that the sub-basalt strata are so mechanically weak that they would form a decoupling zone or detachment to accommodate thin-skin deformation of the CRB.

- Zachariasen et al. (2006) noted that seismicity occurs within the sedimentary rocks between the CRB and basement, indicating that they are brittle and not aseismic. The analysis of focal mechanisms as part of this study (Sections 8.3.2.3, 8.3.2.4, and Appendix F) also shows that earthquakes occur at depths consistent with the sediments beneath the CRB.

As noted by Zachariasen et al. (2006), the differences between the thin-skin and thick-skin models have implications for seismic hazard at a site. The thin-skin model limits ruptures to very shallow depths, which in turn limits rupture area and maximum earthquake magnitude for a given rupture length. Although the thin-skin model may generally predict that a given fault produces smaller maximum earthquakes, seismic moment balance requires more frequent smaller earthquakes on the fault for a given slip rate, the net effect of which may be a higher probabilistic hazard at a site. The thin-skin model also affects site-to-source distance calculations and may affect other aspects of site-source geometry (Zachariasen et al. 2006) (for further discussion of these issues, see Sections 8.4.2 and 8.4.3 of this report).
Studies and interpretations of structure in the YFB since the Zachariasen et al. (2006) report have generally interpreted that major map-scale structures are ultimately rooted in the crystalline basement at depths greater than about 8 km. For example, Pratt (2012) proposed that the Yakima folds originally formed in Miocene time above splay faults that branched upward in the middle to lower crust from a master strike slip fault in the basement along the trend of the OWL, essentially forming a crustal-scale flower structure (Figure 4.4). In support of this model, Pratt (2012) presented a depth-migrated seismic reflection profile across the Saddle Mountain anticline with the interpretation of a moderately south-dipping thrust fault extending through the CRB to a relatively low-angle flat at a depth of about 8–9 km (Figure 4.4), well below the lower contact of the basalt. In Pratt’s model (Figure 4.4), the low-angle reflector on the seismic line is interpreted to be linked to a south-dipping ramp that merges with the master shear zone below about 20 km. Blakely et al. (2011) inferred that the thrust fault underlying the Umtanum Ridge anticline extends through the CRB into the crystalline basement, noting that best-fit gravity models require significant positive relief on high-density basement rocks beneath the fold.

![Figure 4.4. Cross section modified from Pratt (2012) showing model for origin of the Yakima folds as splays from a master strike slip fault in the middle to lower crust along the trend of the OWL.](image)

In a Senior Seismic Hazard Analysis Committee (SSHAC) Level 2 PSHA for a series of dams along the Columbia River upstream of the Hanford Site, Jack Benjamin & Associates (JBA) et al. (2012) evaluated the thin- and thick-skin models and concluded that the preponderance of data provided a basis for a “higher degree of belief” that the thick-skin model is correct. JBA et al. (2012) assigned weights of [0.9] and [0.1] to the thick-skin and thin-skin models, respectively. Key observations cited by JBA et al. (2012) in support of this assessment include the following:

- Evidence for multiple surface-rupturing events of 3 m or more along the thrust fault bounding Toppenish Ridge is not consistent with a rupture width limited to the upper 2–5 km, as implied by the thin-skin model.
- Simple scaling relationships observed in well-studied fold and thrust belts suggest that folds with wavelengths on the order of 10 km or so, which occur in the YFB, have a vertical extent of folded and deformed crust greater than a few kilometers.
• Thin-skin models that call for rooting all of the YFB structures in a common detachment require the presence of a blind, south-dipping ramp to connect shortening above the detachment to north-directed motion and impingement of eastern Oregon (see discussion in Section 4.2). Analogues from well-studied fold and thrust belts suggest a south-dipping ramp and associated large-scale fault-bend fold or culmination should be present south of the YFB structures. To date, however, proponents of the thin-skin model have not identified or proposed a candidate structure to link a shallow detachment in the YFB with a deeper root zone to the south.

• Another class of thin-skin models for the YFB invokes processes that do not require the presence of a detachment, akin to models for the formation of “wrinkle ridges” on Mars. As noted by JBA et al. (2012), however, the geometry of the YFB structures differs in key aspects from Martian structures, and thus the “wrinkle ridges” may not be appropriate analogues. For example, wrinkle ridges on Mars and Venus commonly are associated with topographic features interpreted to be associated with mantle plumes. The YFB anticlines have been explicitly compared to planetary wrinkle ridges because Miocene growth of the folds was temporally associated with impingement of the Yellowstone plume to the east-southeast (e.g., Mège and Ernst 2001). As discussed previously, however, dynamic models of Miocene deformation associated with the Yellowstone hot spot (Camp and Hanan 2008) do not predict the development of compressive stresses and associated upper crustal shortening in eastern and central Washington during the Miocene. Also, it is important to note that the subsurface structure of wrinkle ridges on Mars and Venus is fundamentally unknown, and thus these features cannot be meaningfully used as analogues to infer the subsurface structure of the YFB.

• The YFB structures are more geometrically similar to basement-involved Laramide folds in the southern Rocky Mountains than classic detached fold and thrust belts such as the eastern Canadian Cordillera and Appalachian Valley and Ridge.

For the present seismic source characterization, the SSC TI Team has made the assessment that thrust faults underlying the Yakima folds are thick-skin or “coupled” structures that extend to the base of the seismogenic crust, effectively assigning zero weight to the thin-skin or “uncoupled” model. In addition to the compelling evidence given by Zachariasen et al. (2006) and JBA et al. (2012), the TI Team assessment takes into account the presentations by members of the technical community as represented in presentations and discussions at Workshop 1 (WS1) (PNNL 2013a) and Workshop 2 (WS2) (PNNL 2013b) that fault structures in the YFB extend to the base of the seismogenic crust. The thick-skin deformation model for the YFB also is broadly consistent with recent geodetic studies that show crustal shortening and folding in eastern Washington is driven by and kinematically related to wholesale rotation and translation of crustal blocks in eastern Oregon to the south (see discussion in Section 4.2). The deformation in eastern Washington is a part of a continuous velocity field that encompasses crustal motions in the northern Basin and Range, the forearc region of the Cascadia subduction zone, and the backarc region in eastern Washington. Because these motions are continuous and vary smoothly over length scales of hundreds of kilometers, they likely involve the entire lithosphere and thus the full thickness of the seismogenic crust is likely to be involved in the deformation. Analyses of earthquake focal mechanisms in eastern Washington (see Section 6.0 of Appendix E) document that seismicity below 8 km in the YFB reflects approximately north-south-directed crustal shortening, demonstrating that seismogenic deformation below the CRB and in the crystalline basement is kinematically related to the surface folding. The seismicity analyses (see Sections 4.4 and 8.3.2.9) further demonstrate that, although the spatial distribution of seismicity within the CRB is more swarm-like, there is no profound difference in the kinematics (e.g., orientation of axes of maximum shortening) of upper crustal seismicity within the CRB relative to deeper seismicity below the CRB (see Section 6.0 of Appendix E), supporting the hypothesis that YFB shortening is “thick-skin” and involves the entire seismogenic crust.
Although the structural relief on the CRB and first-order topographic expression of the major anticlines in YFB are strong evidence of crustal shortening, some workers have suggested that strike slip faulting also may be significant in the late Cenozoic tectonics of eastern Washington. Based on recent studies in the southeastern part of the YFB, Blakely et al. (2013) proposed that the dominant mode of deformation along the RAW alignment is dextral shear, producing right-lateral strike slip faulting. Pratt’s model (2012; Figure 4.4) is relevant to this interpretation because it implies that a Miocene-age strike slip fault or shear zone coincident with the modern OWL and RAW alignment is present at depth, and thus may be a locus of active strike slip deformation. The model geometry further implies that the thrust faults underlying the YFB are secondary splays of the master strike slip fault. It is important to emphasize, however, that Pratt’s model was proposed to account for the origin of the Yakima structures in the presumed Miocene tectonic setting. The kinematics of the faults, and possibly their relationship to the model shear zone at depth, has likely changed since they originally formed. According to Pratt (2012), “the Y[akima] F[olds] initially developed as splay faults in the early to mid-Miocene under NNW-oriented principal compressive stress, but the anticlines subsequently grew with thrust motion after the principal compressive stress rotated to N-S or NNE after the mid-Miocene.”

The SSC model described in Chapter 8.0 captures two key kinematic and geometric elements of Pratt’s model: 1) the SSC model explicitly includes reverse motion on faults beneath the anticlines; and 2) the SSC model includes “thick-skin” fault geometries extending to the 20-km depth beneath the folds, as proposed by Pratt (2012) (Figure 4.4). In the modern setting, the thrust faults terminate as brittle-elastic structures at 20 km or shallower depths. Although Pratt’s model assumes that the thrust faults originally were linked below 20 km to a deeper strike slip fault zone, this linkage, if once present, is now below the brittle-ductile transition zone and does not contribute to potential seismogenic (brittle-elastic) behavior of the faults in the modern tectonic setting. The SSC model for this study captures Pratt’s thick-skin geometry of the faults above the modern brittle-ductile transition zone.

High-resolution aeromagnetic data presented by Blakely et al. (2013) image the 8.5-Ma Ice Harbor dikes as a series of well-defined north-northwest-trending aeromagnetic anomalies that obliquely cross the Wallula fault zone southeast of the Wallula Gap. Blakely et al. (2013) interpret the anomalies associated with the Ice Harbor dikes to be offset a total of about 6.9 km across the Wallula fault zone, implying a long-term average rate of 0.8 mm/yr of dextral shear. In contrast, modeling of geodetic data by McCaffrey et al. (2013) suggests that the direction of maximum shortening strain in the YFB is oriented nearly perpendicular to the northwest trend of the RAW alignment, which would favor thrust and reverse faulting along the structure rather than dextral strike slip faulting. McCaffrey et al. (2013) also note that there are no discernable local strain rate gradients across the RAW alignment to indicate that it is presently a locus of active deformation. The conflicting kinematic predictions of the Blakely et al. (2013) and McCaffrey et al. (2013) models appear to arise from differing tectonic interpretations of potential field and global positioning system (GPS) data sets, respectively. Based on a review of the aeromagnetic data cited by Blakely et al. (2013) to interpret dextral offset of the Ice Harbor dikes, the TI team concluded that offsets of 1 to 2 km are permissible within the uncertainty of the data (Section 8.4.3.9.2). The TI team judged that the larger 6.9 km of integrated offset across the entire Wallula fault zone proposed by Blakely et al. (2013) is not technically defensible, however, and thus it is not included in the SSC model (Section 8.4.3.9.2). Given the lack of cross-cutting relationships to fully constrain the timing of offset of the Ice Harbor dikes, the TI team noted that most if not all of the dextral displacement interpreted by Blakely et al. (2013) could have occurred during the late Miocene, consistent with Pratt’s (2012) model for the origin of the YFB structures, and thus not be active in the modern tectonic setting, consistent with the analysis of GPS data by McCaffrey et al. (2013).
Recent and ongoing research in eastern Washington tectonics has focused on potential linkage of YFB structures in the Cascadia backarc region with Holocene-active strike slip and thrust faults in the forearc. Blakely et al. (2011) identified a series of lineaments in aeromagnetic and isostatic residual gravity maps along the OWL that cross the Cascade Range and are along the trend of the northwestern end of the YFB and Quaternary structures in the Puget Lowland (Figure 4.5). Blakely et al. (2011) proposed that these potential field lineaments “imply continuation of the [YFB] through the Cascade Range.” Specifically, they suggested that the White River-Naches fault zone could be a trans-Cascades structural link between the western Umtanum Ridge fault on the southeast with the Tacoma fault to the northwest. Similarly, they suggested that an alignment of structures they informally refer to as the Mt. Lindsay structural zone may represent kinematic linkage between the northwestern end of the YFB and the Southern Whidbey Island fault in the Puget Sound. In this interpretation, the Holocene-active Seattle fault is a westward splay from this overall structural trend. Blakely et al. (2011) explicitly compared the series of potentially linked structures along OWL, including faults in the Puget Lowland and the CLEW and RAW lineaments east of the Cascades, to the active dextral Sakhalin fault in the backarc region of the Kuril Trench (Figure 4.6). This analogy is discussed in the context of its potential implications to the SSC model in Section 8.4.3.9.

To summarize, the backarc setting of eastern Washington was established about 42 Ma following accretion of the oceanic Siletzia terrane to the ancestral continental margin and development of the modern Cascadia subduction zone. Eruptions of the CRB in Miocene time buried most of eastern Washington beneath 2–5 km of basalt. North-directed crustal shortening in late Neogene and Quaternary time has folded the CRB into a series of antiformal mountains and ridges that dominate the topography and geomorphology of the backarc region. The continuity of the YFB with structures in the forearc and regional deformation kinematics are topics of active research. The evaluation of proposed models and their implications to the SSC model are given in Chapter 8.0.

Figure 4.5. Interpreted linkage between structures of the YFB and late Quaternary-active faults in the Puget Lowland region (modified from Figure 22 in Blakely et al. 2011). Key structural features that link the forearc and backarc regions include the White River-Naches fault zone (WR-NR) and the Mt. Lindsay structural zone (ML). The overall northwest-southeast trend of the OWL is indicated by the green arrows. (See text for additional discussion.)
4.2 Contemporary Plate Motions and Tectonic Stress Regime

Active crustal deformation in eastern Washington and the Pacific Northwest is well characterized by regional gradients in the secular velocity field as measured by space-based geodesy. Analyses of GPS data by McCaffrey et al. (2007, 2013) show that much of the Pacific Northwest is moving approximately north with respect to stable North America. Specifically, the velocity field in the Pacific Northwest can be modeled as the motions of a number of rigid blocks or microplates that rotate clockwise with respect to stable North America about a pole or poles in eastern Oregon or western Idaho (McCaffrey et al. 2013; Figure 4.7), resulting in generally north-directed motions. Crustal shortening occurs where these translating blocks impinge on relatively stable regions.

The forearc region of Oregon and Washington is recognized as a coherent microplate known as the Oregon Coast block (Figure 4.8; Wells et al. 1998; Wells and Simpson 2001). The northward translation of the Oregon Coast block is accommodated, at least in part, by active north-south shortening in the Puget Lowland region of northwestern Washington (Mazzotti et al. 2003). Geologic evidence of youthful shortening in this region includes Holocene activity on approximately east-west striking thrust and reverse faults such as the Seattle and Tacoma faults (Barnett et al. 2010, and references therein; also, see Appendix E.4 of this Hanford PSHA document), and hundreds of small earthquakes in northwestern Washington with thrust and reverse focal mechanisms (Lewis et al. 2003).

McCaffrey et al. (2013) similarly model the backarc region of central and northeastern Oregon as a clockwise-rotating rigid block with a rotation pole located in western Idaho near the latitude of the Oregon-Washington state boundary (Figure 4.7). There is a south-to-north gradient in the GPS velocities across southeastern Washington directly north of this block, and crustal motions relative to North America effectively go to zero in northeastern Washington and southeastern British Columbia (McCaffrey et al. 2007, 2013), indicating that net shortening is occurring in the intervening area, which encompasses the YFB. McCaffrey et al. (2013) explicitly model the YFB as a non-rigid deforming zone that is accommodating the northward motion and impingement of eastern Oregon.
As noted by Wells et al. (1998) and Wells and Simpson (2001), the clockwise rotation of crustal blocks in the Pacific Northwest is kinematically linked to the northwest translation of the Sierra Nevada microplate and northwest-directed extension in the northern Basin and Range of northern Nevada and southeastern Oregon (Figure 4.8). The Sierran motion and northern Basin and Range extension together accommodate approximately 20% to 25% of total Pacific-North America motion, which splays east from the plate boundary in the northern Salton trough in southern California, passes through the eastern Mojave block as the eastern California shear zone, and continues along the eastern margin of the southern Sierra Nevada as the Walker Lane belt (Unruh et al. 2003; Faulds and Henry 2008) (Figure 4.9). Geodetic data (Bennett et al. 2003) reveal that northwest dextral shear expands eastward from the central part of the Walker Lane belt into the northern Basin and Range province at the latitude of central Nevada, where it is primarily accommodated by northwest-directed extension (Figure 4.9). Both dextral shear in the northern Walker Lane belt and extension in the northern Basin and Range terminate near the boundary between the Sierran microplate and Oregon Coast block (Faulds and Henry 2008), where the style of deformation east of the oceanic plate boundary abruptly changes from distributed northwest dextral shear to clockwise rotation of rigid blocks about a proximal pole in central Idaho (McCaffrey et al. 2007, 2013; Unruh and Humphreys 2013). Distributed Pacific-North American plate motion in the interior of the Cordillera thus...
Figure 4.8. Model for large-scale translation of the Oregon Coast block (OC) by clockwise rotation about a proximal pole to the east, and accommodation of north-directed motion of the block by crustal shortening in northwestern Washington (modified from Figure 4B in Wells and Simpson 2001).

is accommodated by strike slip and normal faulting east of the Sierran microplate, clockwise rotation of rigid blocks or microplates at the latitude of Oregon, and ultimately drives north-directed shortening in the YFB and Puget Lowland regions in Washington.

In Figure 4.9, the yellow band shows the zone of distributed right-lateral shear that branches from the San Andreas fault in southern California and extends northwest across the Mojave desert as the eastern California shear zone (ECSZ). The yellow zone of dextral shear continues along the eastern margin of the Sierra Nevada microplate as the Walker Lane belt (WLB). Some of the right-lateral motion in the WLB splays northeastward into the northern Basin and Range province (NBR) in northern Nevada and southeastern Oregon, as evidenced by the east-to-west clockwise rotation and increase in length of the velocity vectors in the yellow region across Nevada.
Figure 4.9. GPS site velocities showing large-scale crustal motions across the western United States (data from the U.S. Geological Survey: http://earthquake.usgs.gov/monitoring/gps/. GPS velocities presented in a North America-fixed reference frame derived from International Terrestrial Reference Frame 2008).
Northwest-southeast extension in northern Nevada, southeastern Oregon, and southwestern Idaho accommodates fan-like opening of the yellow region of the NBR about a pole or poles in eastern Oregon/western Idaho, which in turn drives clockwise rotation of blocks in the Pacific Northwest (compare with Figure 4.7).

Available kinematic and stress indicators in south- and central-eastern Washington generally indicate approximately north-south shortening and compression, consistent with the large-scale crustal motions reflected in the GPS velocity data. A search of the 2008 World Stress Map database (http://dc-app3-14.gfz-potsdam.de/pub/stress_data/stress_data_frame.html) (Heidbach et al. 2009) reveals about 10–20 stress indicators from the published literature in the YFB (primarily focal mechanisms and borehole data), all of which consistently indicate a north-south to north-northeast–southwest orientation of the maximum compressive stress ($\sigma_1$) (Figure 4.10).

Figure 4.10. Map of stress indicator data in northern Oregon and southern Washington showing inferred orientations of the maximum compressive stress ($\sigma_1$). The long axis of the bars shows the azimuth of $\sigma_1$; symbols on the bars and colors indicate data type and faulting style, respectively (red = normal faulting; green = strike slip faulting; blue = thrust faulting). Note that the preponderance of data in the YFB and eastern Washington indicate approximately north-south to northeast-southwest $\sigma_1$, and that thrust/reverse faulting is the dominant style of deformation. Base map generated using an online tool at the 2008 World Stress Map site (Heidbach et al. 2009).
McCaffrey et al. (2013) inverted 424 earthquake focal mechanisms from the YFB for components of a reduced stress tensor, using the algorithm SATSI by Hardebeck and Michael (2006). The best-fit solution from their analysis indicates the maximum compressive stress, $\sigma_1$, trends between N4°E to N12°E and is sub-horizontal to gently plunging, which is comparable to the average north-northeast trends of stress indicators in the World Stress Map database (Figure 4.9). McCaffrey et al. (2013) also found that the intermediate principal stress trends east-west and is gently plunging, and the least compressive principal stress plunges between 57° and 80°. This stress tensor geometry is consistent with deformation that primarily accommodates north-northeast-directed shortening and vertical thickening (McCaffrey et al. 2013). A similar analysis performed for this study, in which groups of earthquake focal mechanisms from the YFB were inverted for components of a reduced incremental strain tensor rather a stress tensor, finds generally north-south- to north-northeast–south-southwest–directed shortening and vertical thickening (see Section 6.0 of Appendix E for details).

To summarize, detailed characterization of the crustal velocity field in the western United States via space-based geodesy reveals that much of the Pacific Northwest is translating northward as a series of rigid blocks that move relative to stable North America as if they rotate clockwise about a pole or poles in central Idaho. This clockwise rotation accommodates distributed northwest dextral shear in a broad zone of deformation in the western United States between the oceanic Pacific and Juan de Fuca plates to the west, and stable North America to the east (Figure 4.9). The northward motion of the Cascadia forearc region (i.e., the Oregon Coast block) and eastern Oregon drives active crustal shortening in the Puget Lowland and YFB, respectively. The north-south to north-northeast–south-southwest direction of maximum compressive stress in the YFB indicated by earthquake and borehole data is consistent with the first-order kinematics of the GPS velocity field.

### 4.3 Late Cenozoic and Quaternary History

Post-Columbia River Basalt Group (CRBG) evolution of the Columbia Basin is recorded by folding and faulting in the YFB and deposition of sediments in the basins (Reidel et al. 1994). The study region, which includes much of the Columbia Basin and more locally the Pasco Basin (Figure 4.11), has been the site of deposition of clastic sediments since the cessation of basaltic volcanism approximately 8.5 Ma (DOE 2002; Fecht et al. 2004). Post-CRBG stratigraphic units provide other datums (deposits and geomorphic surfaces) that can be used to evaluate patterns and rates of deformation. Reidel et al. (1994) state that alluvial-lacustrine sediments deposited primarily by the Columbia River system show that the YFB structures were growing and displacing river channels and that lateral distribution of facies, changes in depositional style, and structural deformation of these sediments can be used to evaluate the structural evolution of the YFB.
4.3.1 Post-CRB Regional Stratigraphy

The oldest suprabasalt deposits include fluvial sands and gravels and lacustrine sands, silts, and clays of the late Miocene to Pliocene upper Ellensburg Formation, Ringold Formation, and Snipes Mountain Conglomerate (Newcomb 1958; Lindsey 1995; Gustafson 1978; Reidel et al. 1994; DOE 2002). The Thorp Gravel in Kittitas Valley is believed to be a time-equivalent to the upper Ringold Formation in the Pasco Basin (Waitt 1979). According to Fecht et al. (2004) the region experienced a base-level change near the end of or immediately following Ringold deposition that resulted in rapid rejuvenation of rivers, which incised into and cut laterally across the sedimentary fill (Newcomb et al. 1972). This resulted in a hiatus in the stratigraphic record. A younger sequence of mainstream fluvial sands and gravels and associated overbank sands and silts were then deposited in the newly modified river valley. Tributary sidestreams also aggraded sediments and in a few locations discharged sediments into the main trunk streams of the Columbia Basin. The mainstream and sidestream fluvial sediments form the Cold Creek unit and are Plio-Pleistocene in age. A standardized stratigraphic nomenclature for post-Ringold deposits was established by DOE (2002) based on regional stratigraphic and sedimentary facies observations (Figure 4.12). In addition to the Cold Creek unit (or CCU, formerly referred to as the Plio-Pleistocene unit), these include the Hanford formation (deposits from multiple cataclysmic Ice Age flooding events) and Holocene deposits. Around the margins of the basin on the flanks of the basalt ridges, Pleistocene alluvial fan deposits also are present (Baker et al. 1991).

Brief descriptions of the key suprabasalt units used to provide constraints on the timing and rates of late Cenozoic deformation in the study area are provided in Appendix E.
Figure 4.12. Hanford Site stratigraphy. Photos show examples of strata, both in drill core and outcrop, which are representative of the Gable Gap study area. Borehole number and depth of sample below ground surface (bgs) are shown where appropriate (from Bjornstad et al. 2010).
4.3.2 Summary of Late Miocene, Pliocene and Quaternary History

The evolution of the Columbia River system since the Miocene provides insights into the structural evolution of the study region. Reidel and Tolan (2013) summarize the evolution of the Columbia River system since the Miocene as a consequence of flood basalt volcanism, Cascade Arc volcanism, and tectonism in the Pacific Northwest. During the waning phase of CRBG eruptions (12.5–8.5 Ma), the Columbia River flowed south across the YFB. The post-CRBG pre-Ringold channel (upper Ellensburg Formation and Snipes Mountain Conglomerate) of the Columbia River extended across the western Pasco Basin, entering at Sentinel Gap and exiting near Sunnyside Gap (Reidel et al. 1994) (Figure 4.13a). The Ringold Formation in the Pasco Basin records later phases of an eastward-shift in the channel (Figure 4.13b). Lowermost Ringold sediments are interpreted to have been deposited in the Pasco Basin approximately 8 Ma by an ancestral Columbia River that entered the Pasco Basin through Sentinel Gap and exited through a structural wind gap between Rattlesnake Mountain and Red Mountain. The flow margin of the Ice Harbor flow influenced the course of the river at this time. Sometime after 8 Ma, the ancestral Columbia River abandoned the Yakima Valley and began flowing through Wallula Gap (Figure 4.13c). Although the reason for the abandonment of the westward course is debatable, Reidel and Tolan (2013) favor a hypothesis that it reflects continued growth of Naneum Ridge and the Horse Heaven Hills, along with continued subsidence of the Pasco Basin. The growth of Naneum Ridge uplift also may explain a 30-km westward shift of the Yakima River from the Konnowock Pass to Union Gap on the western flank of the Naneum Ridge. As illustrated in Figure 4.13c, the Yakima River following the shift of the ancestral Columbia River to Wallula Gap flowed initially through Badger Coulee, joining the Columbia River at Kennewick-Pasco, Washington.

Fluvial deposition through much of early to middle Ringold time (8 Ma to approximately 5.5 Ma) was dominated by aggradation of pebble- to cobble-gravel. At about 5.5 Ma, an abrupt change to sand-dominated deposition occurred (Reidel et al. 1994; Lindsey 1996). The cause is unknown, but it is speculated to be related to some combination of uplift of the YFB structures and renewed Cascadian volcanism that resulted in the aggradation of the ancestral Columbia River canyon west of Hood River, Oregon (Reidel and Tolan 2013).

During late Ringold time, repeated influx of hyaloclastic debris resulted in the rapid aggradation of the ancestral Columbia River canyon west of the vicinity of present-day Hood River, Oregon (Tolan and Beeson 1984). This rapid aggradation of the ancestral Columbia River might have been a major contributing factor that led to the formation of lakes within Pasco Basin during late Ringold time (Fecht et al. 1987). Strata in the Pasco Basin record the formation and aggradation of three separate facies during the period between approximately 5 Ma and 3 Ma (Smith et al. 2000, see Figure 4.3 in Appendix E).

The regional uplift, which began ~3–4 Ma, marked the end of widespread sediment deposition and the beginning of stream and river entrenchment that created the present Columbia River Gorge, the beginning of the entrenchment of the meanders of the major tributaries of the Columbia River (e.g., the Yakima River between Ellensburg and Yakima), and the period of time when more than 100 m of Ringold sediments were removed from the Pasco Basin (Reidel and Tolan 2013). Well-developed pedogenic soils subsequently formed calcrite caps in the Ringold deposits, as well as upper Ellensburg, and other upland basalt surfaces.
The final phase in the development of the Columbia River system occurred during the Pleistocene. Cataclysmic floods (“Missoula” or “Spokane” floods) were generated by episodic, cataclysmic releases of large volumes of water from glacial Lake Missoula. Multiple episodes of cataclysmic Ice Age floods are recorded in the Pacific Northwest beginning as early as 1.5–2.5 Ma based on evaluation of surface exposures and borehole studies in southeastern Washington (Bjornstad et al. 2001). During the most recent Ice Age starting ~30 ka BP, the Purcell Trench lobe of the continental ice sheet spread down the Pend Oreille River near the Idaho-Montana border. This lobe dammed the Clark Fork River, impounding

Figure 4.13. a) Detail of the Pasco Basin area immediately after Elephant Mountain time. b) Detail of the Pasco Basin area immediately after Ice Harbor time. c) Detail of the Pasco Basin area for middle to late Ringold time. (From Reidel and Tolan 2013)
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This dam then broke out episodically throughout marine oxygen isotope stage (MIS)-2 (29–14 ka) with the most recent floods occurring during the late Wisconsin. Hydraulic damming of flood waters at Wallula Gap periodically generated a temporary lake, Lake Lewis, that resulted in deposition of bedded, fine-grained sediment (commonly referred to as slackwater deposits or Touchet Beds) in the Pasco, Walla Walla, and Yakima Basins. The maximum height of Lake Lewis at Wallula Gap was approximately 380 m (Last et al. 2012). A discussion of the ages of the various flood-related deposits observed in the study region is provided in Section 4.2.4 of Appendix E.

Prior to the Pleistocene, the Columbia River took a more direct path from Sentinel Gap to Wallula Gap through Gable Gap (Bjornstad et al. 2010). After the floods it achieved its present course from Priest Rapids Dam to Richland, Washington, the Hanford Reach of the Hanford Reach National Monument. The lower Yakima River also achieved its present course as the floods ended. Gravels from the Pleistocene floods were deposited at the entrance of Badger Coulee forcing the Yakima River to flow north through the Rattlesnake Mountain-Red Mountain water gap and join the Columbia at the present confluence (Reidel and Tolan 2013). Fecht and Marceau (2006) provide a description of the glaciofluvial and fluvial sediments and bedforms that are present in the river valley that formed after Pleistocene largescale cataclysmic flooding.

Reidel et al. (1994) state that all geologic structures within the Columbia Basin had developed their present relief by the end of the Pleistocene and that evidence of continued growth is mainly in the frontal fault zones (i.e., imbricated thrust or reverse faults associated with the steep forelimbs of asymmetrical anticlines). Although erosion and deposition of slackwater deposits and eolian deposits that occurred as a result of these floods greatly influenced the geomorphology of the study region and in many places obscures the record of Quaternary tectonic deformation of YFB structures, evidence of Pleistocene and Holocene faulting has been documented at many localities (see Section 8.4).

4.4 Seismicity in the Hanford Site Region

The seismic hazard at the Hanford Site comes from local and regional crustal sources and from distant sources associated with the Cascadia subduction zone (see hazard sensitivity results in Chapter 10.0). These two types of earthquake sources have different characteristics and for the purpose of calculating earthquake recurrence parameters are maintained in the Hanford PSHA as two separate catalogs (see Chapter 6.0). Details about the geographic and temporal extent of the two catalogs are given in Section 6.1. General characteristics of the regional and local seismicity for the crustal and subduction catalogs are described in this chapter, details of the earthquake catalogs and their analyses are given in Chapter 6.0, and the use of the seismicity data for purposes of seismic source characterization is given in Chapter 8.0.

4.4.1 Crustal Seismicity

The geographic extent of the local and regional crustal earthquakes included in the crustal catalog for the Hanford PSHA is shown in Figure 4.14 and covers an area between latitude 45°N to 49°N and longitude -121.5°E and -117.5°E. To the west of the crustal catalog region, earthquakes are mainly associated with the Cascades, the Cascadia subduction zone, and active crustal faults in the Puget Lowland, such as the Seattle, Tacoma, South Whidbey Island, and Devils Mountain faults (Johnson et al. 1996, 1999; Sherrod et al. 2004; Kelsey et al. 2004; Sherrod et al. 2008). The seismicity in the Hanford region is divided into the historical record (pre-1936), the early instrumental period (1936–1969), and the modern instrumental period (post-1969).
4.4.1.1 Studies of Individual Earthquakes

The strongest earthquakes that occurred in the region surrounding the Hanford Site were the 1872 Chelan earthquake and the 1936 Milton-Freewater earthquake (Figure 4.14).

The December 15, 1872 Lake Chelan Earthquake

The December 15, 1872, is the largest historical earthquake known to have occurred in the Pacific Northwest east of the Cascade Mountains (Bakun et al. 2002). It is known that the earthquake was strong enough to cause landsliding and opening of ground fissures, and that it was felt at large distances (Montana, Idaho, British Columbia, Oregon, and throughout Washington). At the time of the earthquake, the region around Lake Chelan was sparsely populated and the descriptions of the effects of the earthquake are subject to interpretation. Crider et al. (2003) report a range in magnitude estimates from various researchers as 6.5–7.4. Bakun et al. (2002) conducted the most comprehensive analysis of the macroseismic intensity data for the earthquake and noticed that there are differences of up to 2 MMI between intensity assignments by different authors and such differences are sufficient to determine uncertainty in the epicentral location of the earthquake. The location used in the earthquake catalog is that obtained by Bakun et al. (2002) near the south end of Lake Chelan; its magnitude derived from macroseismic intensity (M\text{I}) was 6.8, with a range of 6.5 to 7 at the 95% confidence level. The Lake Chelan earthquake has not been attributed to a particular geologic feature, although several structures have been investigated as possible sources (e.g., Straight Creek fault, Leavenworth fault, Entiat fault, and Chiwaukum graben). A zone of seismicity persists in the region (Figure 4.14).

The July 15, 1936 Milton-Freewater Earthquake

The July 15, 1936 Milton-Freewater earthquake is estimated to be M\textsc{L} 6.1 (WCC 1980b) to M\textsc{L} 6.4 (Bott and Wong 1993). The maximum observed intensity was estimated in excess of MMI VII (JBA et al. 2012). The original epicentral location, based on isoseismal data, was placed about 10 km northeast of Milton-Freewater (Neumann 1938). However, later detailed studies by Woodward-Clyde Consultants (WCC 1980b) suggested that the earthquake originated north-northeast of Walla Walla, Washington, on the Hite fault, rather than at the epicenter defined by intensity data. A reassessment of intensity data by Mann and Meyer (1993) placed the epicenter on the Wallula fault. This preferred location proposed by the WCC (1980b) study is based, in part, on their investigation of the April 8, 1979 earthquake (Figure 4.14), for which a high-quality epicentral determination is possible. The 1979 earthquake fault plane solution indicates either oblique, right-lateral reverse motion on a N30°E fault, or alternatively, oblique, left-lateral reverse motion on a N40°W fault. Although both orientations align with known fault systems (the Hite fault and the Wallula fault), the WCC (1980b) report prefers the N30°E orientation because it agrees with the geologically determined sense of slip, which is contrary to that observed for the Wallula fault system. In addition, the epicenters of the 1936 and 1979 earthquakes align with the trend of the Hite fault and the N30°E fault plane solution (WCC 1980a).
Figure 4.14. Crustal seismicity in the Hanford Site region. Shown are the epicenters of earthquakes included in the crustal catalog for the project and the magnitudes are expected magnitude E[M] values, which might differ from the original catalog values (see Section 6.5.2). Shown are the epicenters of the 1872 Lake Chelan earthquake (Bakun et al. 2002), the 1936 Milton-Freewater earthquake (WCC 1980b), and the April 8, 1979 earthquake. Blue lines indicate the seismic source zones used in the PSHA analysis (Section 8.3) and red lines are the fault sources (Section 8.4).
4.4.1.2 Instrumental Seismicity

The present-day crustal seismicity within the region of the Hanford Site is mostly concentrated in three areas: the YFB, the Entiat-Chelan seismic zone, and to a lesser extent, the epicentral area of the 1936 Milton-Freewater earthquake (Figure 4.14). The YFB is associated with small- to moderate magnitude earthquakes, many of which occur within the CRB above the underlying Tertiary sedimentary rocks and the pre-Tertiary basement rocks. Gomberg et al. (2012) analyzed 5,160 earthquake records for M > 0.0 from the Pacific Northwest Seismograph Network (PNSN) for the period from 1970 to 2010 to explore the relationship of seismicity to regional tectonics and near-surface processes. Their study indicates that much of the seismicity (about one-third of the total events) can be characterized as being spatially clustered (i.e., swarm-like), with approximately 40% to 50% of the total events occurring in the shallow crust within or above the CRB; the remainder are interpreted to occur in the deeper crust as the result of regional tectonic deformation associated with plate convergence and clockwise rotation (Gomberg et al. 2012). This observation is in agreement with earlier studies (e.g., Rohay and Davis 1983; WCC 1980b) that characterize YFB seismicity as shallow, low-magnitude swarms rather than classical mainshock-aftershock earthquake sequences. Figure 4.15 shows the epicenters of earthquakes in the upper 3 km within the CRB and earthquakes having depths greater than 3 km. The small magnitudes and swarm-like spatial distribution of the shallow seismicity reported by other researchers are supported by the analyses conducted for this study.

Although small-magnitude earthquake swarms in the YFB constitute a persistent feature, several studies conclude that seismicity in the YFB cannot be clearly associated with particular faults or fault planes (Zachariasen et al. 2006; JBA et al. 2012; Gomberg et al. 2012), with the possible exception of seismicity associated with Saddle Mountain. Several earthquake swarms repeat through time, although swarms appear to migrate spatially. Swarms typically last from days to months. Examples include swarms between the Saddle Mountain and Frenchman Hills structures (which hosted the December 20, 1973, M 4.4 Royal Slope earthquake), the Wooded Island swarms (Wicks et al. 2011; Gomberg et al. 2012), and the Coyote Rapids swarm (October 25, 1971, M 3.8 Coyote Rapids earthquake) (Rohay and Davis 1983). Gomberg et al. (2012) postulate that the spatial and temporal relations between earthquake swarms suggest a causal mechanism associated with near-surface processes related to fluids; however, they fail to find strong evidence for hydrologic processes as their source. Wicks et al. (2011) propose a model for the 2009 Wooded Island swarms that attributes shallow seismicity to slip on a shallow thrust fault and a near-horizontal fault. They suggest that the near-horizontal fault is a bedding-plane fault located between basalt flows.

For the Hanford PSHA, a subset of the highest-quality records for earthquakes from the catalog assembled for this project were relocated using two independent algorithms by Cliff Thurber and Felix Waldhauser (see Section 5.3 and Appendix F). These high-resolution earthquake relocation analyses were conducted to evaluate the spatial resolution of the hypocenter locations in the catalog and provide a data set that the SSC TI Team could use to assess any spatial associations between seismicity and fault sources. Cliff Thurber (Appendix F) used seismic tomography, both conventional and double-difference, to obtain earthquake relocations from a set of P-wave arrivals of well-recorded earthquakes obtained from the PNSN catalog. The double-difference method minimizes the residuals between observed and predicted travel time differences for pairs of earthquakes and improves the resolution of earthquake locations. Differences between the PNSN catalog locations and the relocations are relatively small: the mean absolute values of the shifts are 0.66 km in the east-west direction, 0.64 km in the north-south direction, and 1.42 km in the focal depth. The estimated location uncertainties from conventional
Figure 4.15. Instrumental seismicity in the site region for those earthquakes having reliable depths in the Hanford PSHA crustal catalog. Left panel shows earthquakes in the shallow crust with depths less than or equal to 3 km. Right panel shows earthquakes having depths greater than 3 km. Magnitude values used in the plots are E[M] values (see Section 6.5.2).
tomography are on the order of 500 to 750 m in epicenter and 1 to 1.5 km in depth for good-quality
PNSN events; the double-difference tomography returns smaller location uncertainties. The relocations
obtained by Felix Waldhauser and those by Cliff Thurber are generally consistent and differences in the
standard deviations are small: 0.9 km in the east-west direction, 0.7 km in the north-south direction, and
1.5 km in depth.

A conclusion of these relocation efforts is that the spatial clusters of shallow hypocenters are robust
features of the seismicity. In addition, Waldhauser and Thurber’s relocation results led them to conclude
that seismicity could not be readily associated with particular fault structures, supporting the conclusions
expressed in earlier studies (e.g., Zachariasen et al. 2006; JBA et al. 2012). As discussed in Section
8.1.4.3, the SSC TI Team examined the 3-D distribution of earthquake hypocenters and the downdip
geometries of the faults of the YFB. They also concluded that spatial associations of the seismicity with
the faults at depth are not apparent.

4.4.2 Cascadia Subduction Zone Seismicity

Earthquakes included in the Hanford PSHA Cascadia subduction zone catalog are shown in
Figure 4.16, and the manner in which the catalog was developed is given in Chapter 6.0. The Cascadia
subduction zone seismicity consists of seismicity associated with the plate interface and the intraslab (also
called Benioff zone) sources. The region of the catalog includes the coastal regions of Oregon,
Washington, and portions of British Columbia, and extends east to the Cascade Mountains. Both crustal
and subduction earthquakes occur within this region, so the removal of crustal events is done based on the
location of their hypocenters with respect to the depth of the top of the slab, as modeled by McCrory et al.
(2006). Section 6.1.3 describes this process in more detail. The observed seismicity for the intraslab
source provides the fundamental basis for defining that source’s geometry and recurrence behavior
(Section 8.2.2). In contrast, the plate interface has been notably quiescent during the period of
instrumental recordings, so its location and geometry is not readily determined by observed seismicity.
As discussed in detail in Section 8.2.3, a variety of geophysical and geologic data are used to assess the
geometry and recurrence behavior of the plate interface source.
Figure 4.16. Cascadia subduction zone seismicity included in the subduction earthquake catalog (Chapter 6.0). Contours of depth to the top of the slab and the depths of hypocenters are shown.
4.5 References


4.32


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