Structural Definition of the Cascadia Locked Zone
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CITATION


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Abbreviations and Acronyms

AT&SML  Active Tectonics and Seafloor Mapping Lab
BOEM    Bureau of Ocean Energy Management
CEOAS   College of Earth, Ocean and Atmospheric Sciences
CTD     Conductivity, Temperature, Depth
DOI     US Department of the Interior
GIS     Geographic Information System
GPS     Global Positioning System
JDF     Juan de Fuca
MMI     Marine Mammal Institute
NAMSS   National Archive of Marine Seismic Surveys
NGDC    National Geophysical Data Center
NMFS    National Marine Fisheries Service
NOAA    National Oceanic and Atmospheric Administration
NOAM    North America
NSF     National Science Foundation
OCNMS   Olympic Coast National Marine Sanctuary
OCS     Outer Continental Shelf
ODFW    Oregon Department of Fish and Game
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<td>Office of Naval Research</td>
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<tr>
<td>OSU</td>
<td>Oregon State University</td>
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<tr>
<td>ROV</td>
<td>Remotely Operated Vehicle</td>
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<tr>
<td>USGS</td>
<td>U.S. Geological Survey</td>
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<tr>
<td>Western GECO</td>
<td>Western Geophysical Company</td>
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1. Executive Summary

The structure of the submarine Cascadia forearc can be divided into several distinct domains. The outer wedge in northern Oregon and all of Washington is a landward/mixed vergence wedge with low wedge taper, widely spaced folds, and presence of mud volcanoes. The landward/mixed vergence is linked to high, possibly near lithostatic pore fluid pressure based on geologic observations and Coulomb modeling. This domain is separated from an older complex by a significant landward vergent splay fault. In places, the older complex structures are truncated at the splay fault, suggesting a prior episode of frontal erosion, most likely during the Pliocene. Strike directions in the older complex range from NW-N. In several areas notably off SW Washington and NW Oregon, an inner forearc domain with transverse structural trends is identified. The inner structural domain most likely represents forearc response to N-S compression, which has previously been identified with scattered borehole breakouts, focal mechanisms, and onshore structural observations. We have delineated a boundary between the inner forearc and upper slope domains that may represent a change in principal horizontal compressive stress from ~N-S (inboard) to ~E-W (outboard). We propose that this stress boundary marking this stress transition, also approximately maps the downdip limit of significant interplate basal shear stress. The updip boundary may be located approximately at the transition to the landward vergent domain, where high pore-fluid pressures likely override expected coupling based on thermal considerations.

The downdip boundary shows considerable heterogeneity, with broad seaward swings off SW Washington and central-northern Oregon, and landward swings off northern Washington, and the three major structural zones of high deformation and uplift, Nehalem, Heceta, and Coquille Banks. Loci of interplate coupling in GPS models closely match evidence for strong coupling in for form of folding faulting and uplift at major structural uplifts. The lack of coupling in the landward vergent region, in several extending and lightly deformed regions are also matched in GPS and structural data. This result strongly suggests that locking heterogeneity is related to long-term forearc architecture, similar to observations for M9 events in Sumatra (2004) and NE Japan (2011). The proposed stress line is poorly correlated to along-strike variability of co-seismic subsidence. Possible causes for this mismatch include the lack of areal coverage of the coastal data, the ambiguity of the subsidence data relative to expected subsidence profiles across-strike, a long-term mismatch between strain accumulation and co-seismic slip, and or low precision of many of the earlier subsidence data. We speculate that compartmented regions of high pore fluid pressure due to poorly drained conditions, and bounded by highly coupled regions may be responsible for the two narrow regions of persistent low basal shear stress in SW Washington and north central Oregon, while subducted features may be responsible for the paleoseismic segment boundary near Cape Blanco. The coupling model developed from this work is compatible with both onshore and offshore paleoseismic data and proposed segment boundaries. We suggest that pat ruptures may die out in two poorly coupled regions of SW Washington and N. Oregon. The proposed stress transition coupled with GPS locking models suggest not only significant uncoupled regions, but significant asperities that extend further landward than smoother coupling models currently in use.
suggest. In particular, significant asperities are suggested off northern Washington, northern Oregon and Central Oregon. The landward position of the downdip locking transitions at these asperities has significant implications for expected PGA values for Seattle, Vancouver, Portland, and smaller communities on the Oregon coast including Astoria, Newport, Bandon and others.

2. Introduction

The recent Sumatra 9.15 and to an even greater extent, the Mw=9 Tohoku earthquake have demonstrated that even with 1000 years of history, and the best geodesy and instrumentation in the world, we can still fundamentally misunderstand the occurrence of subduction zone earthquakes. Long paleoseismic records are needed for an improved understanding of global plate boundary seismic behavior, particularly given the recent failure of long accepted global earthquake models. Despite decades of research, definition of the Cascadia locked zone is still the subject of great debate. Geodetic signals indicate plate coupling, but may or may not be indicative of co-seismic release. Onshore paleoseismology is improving, but large uncertainties exist when attempting to model co-seismic subsidence from paleoearthquakes. With respect to tsunami generation, one of the most significant impediments to assessing the threat posed by tsunamis to coastal communities residing adjacent to an actively subducting continental margin is the lack of knowledge of co-seismic deformation that occurs in the over-riding plate during an earthquake. This is particularly true of Cascadia where the deformation front and active accretionary wedge lie offshore and unmonitored by modern geodetic instrumentation.

Paleoseismic records presently cannot illuminate a clear definition of the map area of past ruptures, though that may be on the horizon. Consequently, for regions like Cascadia that lack an instrumental seismicity catalog that can define the “locked zone”, other methods are required. GPS geodesy, leveling, gravity, borehole breakouts and the coastal subsidence from past great earthquakes have all been applied to one degree or another as proxies for this important parameter. These lines of evidence are here assessed, along with a new proxy, the structural evidence of interplate locking, to derive an estimate of interplate locking over several time frames extending from decades to post-Miocene times. We focus on the structural and mechanical character of Cascadia's submarine forearc, as evidenced from geological and geophysical data, in an attempt to extract long-term, multi-cycle strain accumulation trends and to estimate the relationship of inelastic strain reflected in these structures to the underlying regions of inter-plate coupling.

3. Tectonic Setting and Key Elements

3.1 Overview

The Cascadia subduction zone is formed by the subduction of the oceanic Juan de Fuca and Gorda plates beneath the North American plate off the coast of northern California, Oregon, Washington, and Vancouver Island (Figure 3-1). The convergence rate is ~35–38 mm/yr. directed N60° E at the latitude of Oregon (0.4 Ma interpolation in Mazzotti et al., 2003,
depending on models and reference frames). Juan de Fuca-North American convergence is oblique, with obliquity increasing southward along the margin. The submarine forearc widens from 60 km off southern Oregon to 150 km off the northern Olympic Peninsula of Washington, where the thick Pleistocene Astoria and Nitinat Fans presently are being accreted to the margin (Figure 3-1). The active accretionary thrust faults of the lower slope are characterized by mostly seaward-vergent thrusts on the Oregon margin from 42° to 44°55' N and north of 48°08' N off Vancouver Island, and by landward-vergent thrusts between 44°55' and 48°08' N, on the northern Oregon and Washington margins (Carson, 1971, Silver 1972, Goldfinger et al., 1994; 1997, Mackay, 1995, Adam 2004). The landward-vergent province of the northern Oregon and Washington lower slope may be related to subduction of rapidly deposited and overpressured sediment from the Nitinat and Astoria Fans (Seely, 1977; MacKay, 1995; Goldfinger et al, 1997; Adam et al., 2004).

Off Washington and northern Oregon, the broad accretionary prism is characterized by a low wedge taper and widely spaced landward vergent accretionary thrusts and folds (which offscrape virtually all of the incoming sedimentary section). Sparse age data suggest that this prism is Quaternary in age and is building westward at a rate close to the orthogonal component of plate convergence (Figure 3-2; Westbrook, 1994; Goldfinger et al., 1996a; 1997). This young wedge abuts a steep slope break that separates it from the Eocene-Pliocene accretionary complex and the continental shelf. North of the mixed vergence province on the Canadian margin, vergence is mostly seaward, with a steeper narrower younger wedge, except for a landward vergent segment centered at ~49 N. (Gao et al., 2015, Yorath et al., 1987; Clowes et al., 1987; Davis and Hyndman, 1989, Spence et al., 1991a, 1991b; Hyndman et al., 1994; Yuan et al., 1994). In Southern Cascadia, mixed but mostly landward vergence and a narrower steeper wedge similarly characterize the margin (Clarke and Carver, 1992, Gulick et al., 1998). Landward vergence there may be mechanical in nature and related to a seaward dipping backstop rather than overpressured wedge sediments. Much of the western Oregon and Washington coast, and the continental shelf of Oregon is underlain by a basement of Paleocene to middle Eocene oceanic basalt with interbedded sediments known as the Crescent or Siletzia terrane. This terrane may have been accreted to the margin (Duncan, 1982), or formed by in place by rifting and extension parallel to the margin (for example, Wells et al, 1984). Much of the Oregon and Washington shelf is underlain by a moderately deformed Eocene through Holocene forearc basin sequence forming en-echelon discontinuous basins (McNeill et al. 2000). A major terrane boundary exists in southern Cascadia separating the Siletzia Terrane to the north from the Klamath Terrane to the south. The boundary, originally thought to lie along the Fulmar Fault of Snavely (1987), comes ashore at Five Mile Point, just south of Coos Bay, Oregon.
3.2 The Cascadia “Locked Zone”

The earthquake potential of Cascadia has been the subject of major paradigm changes in recent years. First thought to be aseismic owing to the lack of historical seismicity, great thickness of subducted sediments, and low uplift rates of marine terraces (Ando and Balazs, 1979; West and McCrumb, 1988), Cascadia is now thought capable of producing great subduction earthquakes on the basis of paleoseismic and tsunami evidence (for example, Atwater, 1987; Darienzo and Peterson, 1990; Nelson et al., 1995; Satake et al., 1996; 2003), geodetic evidence of elastic strain accumulation (e.g.
Mitchell et al., 1994; Savage and Lisowski, 1991; Hyndman and Wang, 1995; Mazzotti et al., 2003; McCaffrey et al., 2000, 2012; 2013; Miller et al., 2001), and comparisons with other subduction zones (e.g., Atwater, 1987; Heaton and Kanamori, 1984). Despite the presence of abundant paleoseismic evidence for rapid coastal subsidence and tsunamis, the plate boundary remains the quietest of all subduction zones (Acharya, 1992), with only one significant interplate thrust event ever recorded instrumentally (Oppenheimer et al., 1993). Cascadia represents an end member of the world’s subduction zones in both seismic activity (Acharya, 1992) and temperature. The Cascadia plate interface is among the hottest subduction thrusts because of its young subducting lithosphere and thick blanket of insulating sediments (McCaffrey, 1997).

With the past occurrence of great earthquakes in Cascadia now well established, attention has turned to magnitude, recurrence intervals, and segmentation of the margin. Geodetic leveling surveys across the onshore Cascadia forearc show that some areas are tilting landward on a time scale of 70 years. These data also indicate that in some places, tilting is occurring parallel to the arc. Mitchell et al. (1994) calculated tectonic uplift rates from the leveling data using ties to tide gauges. The uplift signal is highly variable along strike in Cascadia, with central Oregon and central Washington apparently undergoing no tectonic uplift, whereas other areas are rising at rates of 1–4 mm/yr. The geodetic uplift rates in the fast-rising areas greatly exceed the

Figure 3-2. Cross-section of the submarine forearc off central Oregon. Embedded Siletzia terrane, oblique and margin parallel strike-slip faults, are shown. Bathymetry shows widening of the accretionary prism and northward shift to mixed vergence off northern Oregon. From Goldfinger et al., (1997)
geologically determined rates of marine-terrace uplift and have thus been attributed to elastic-strain accumulation preceding a future subduction zone earthquake (Mitchell et al., 1994; Hyndman and Wang, 1995, Burgette et al., 2009). Initial elastic-dislocation models based on thermal and GPS data indicated that the locked plate boundary must lie offshore (Hyndman and Wang, 1995; Mitchell et al., 1994; McCaffrey et al., 2000; 2007), however, the meaning and existence of the high variability in rates is controversial. Hyndman and Wang (1995) attribute the variability to artifacts in data processing, whereas Mitchell et al. (1994) consider them the real products of a locked zone of varying width. Goldfinger and McNeill (2006) and Priest et al. (2009) suggest that structural evidence offshore supports long-term asperities underlying uplifted submarine structural highs offshore that coincide with areas of rapid uplift onshore. In contrast, Wells et al. (2003) proposed a forearc basin-centered asperity model for Cascadia and elsewhere. In addition, the offshore position of the locked zone has been challenged by Chapman and Melbourne (2009) who suggest that geodetic data can be interpreted as potentially co-seismic locking further landward in Washington.

In a recent effort to capture the range of views on the downdip limit, a workshop was held in December 2011 at University of Oregon (Frankel, unpublished data, 2012). The results of this meeting and subsequent discussions were utilized in a logic tree in the USGS 2014 update of the National Seismic Hazard Maps (Peterson et al., 2014). Figure 3-3 shows some of the primary models considered. The logic tree represented reviewed and published literature, and some modifications and updates to published information derived through the workshop and subsequent written communications. Discussions and the logic tree centered on three options 1) the Flück et al. 1997) thermal and geodetically based model; 2) GPS locking models; and 3) models based on the upper limit of ETS (Figure 3-3). The USGS logic tree used 50% weight for a model based on the 1 cm/yr. locking contour from averaged from two GPS models. This highly weighted model represents ~ 25% locking fraction. 20% was assigned to a contour midway between the base of the Flück model and the preferred model. A 30% weight was assigned to a model extending to the top of the ETS zone. In another recent effort, Hyndman (2013) considered the nine lines of evidence in evaluating the downdip edge: The constraints are:

1. The “locked/transition” zones on the thrust from modeling geodetic observations of current deformation (GPS, repeated leveling, tide gauges, etc.).
2. Past great earthquake rupture zone constrained by paleoseismic coastal marsh subsidence, 1700 and earlier events, i.e., “paleogeodesy”.
3. Seismic behavior limits from downdip temperatures, loosely described as the “brittle-ductile transition”, approximately 350°C for where the rupture can initiate and there is full rupture, 450°C for the landward limit of the transition zone from full rupture to zero displacement.
4. Change in seismic reflection character from a thrust marked by a thin sharp reflection to a thrust zone marked by a thick set of reflectors, i.e., brittle to ductile shear deformation.
5. Fore-arc mantle corner, landward of which there is inferred aseismic serpentinite and talc overlying the thrust.
6. Updip limit of interseismic episodic tremor and slip (ETS) slow slip that accommodates most of the long-term plate convergence.
7. Empirical association of rupture area with shelf –slope sedimentary basins defined mainly by gravity anomalies.
8. Landward downdip limit of small thrust earthquakes on the subduction thrust that indicate seismic behavior.
9. Landward limit of earthquakes on the Nootka transform fault zone as it subducts beneath the margin of Vancouver Island.

To these we may also add:
10. Landward limit of anelastic margin parallel deformation (see Priest et al. 2009).
11. Landward limit of oblique strike-slip faulting of the submarine forearc (see Goldfinger et al. 1996b, 1997).
12. Intersection of the base of crustal seismicity with the plate interface.

3.3 Proxy Uncertainties
Despite the extensive list of proxies for the downdip edge of interplate coupling, each has considerable uncertainty.

Thermal models
Downdip extent is broadly constrained by thermal geodetic models as originally presented (Hyndman and Wang, 1995). These models however are dependent on short marine heat flow probes inserted in unconsolidated sediment deposited over hydrologic systems in the wedge and slab that are completely unknown. Unfortunately, there is no comprehensive coupled model that can be used to downward continue these data to the plate interface; only very general gradient models exist to estimate the temperature at the plate interface. Hyndman (2013) estimates that uncertainties in both the critical temperatures and the thermal model to result in several 10’s of km of uncertainty.

Leveling Geodetics
The recent 2011 Tohoku earthquake showed the dramatic differences that may exist between geodetically derived coupling, which extended to a depth of ~100 km, and coseismic slip, which extended to less than half that depth (Ikeda, 2005; Ikeda et al., 2012, Goldfinger et al., 2013). Geodetic models based on leveling and tide gauges (Hyndman and Wang, 1993; 1995, Mitchell, 1994; Burgette et al (2009) provide a model representing a multidecadal view of interplate coupling that is longer than GPS (~70 years), but also just a fraction of a seismic cycle. Geodetic models indicate a potentially straightforward model of the spatial distribution of interplate coupling, but modeling this is dependent on the fraction of coupling, and also the potentially large differences between spatial coupling and interseismic release.

GPS Geodetics
GPS provides a more spatially dense view of crustal motion, however to extract the coupling signal, other non-elastic signals need to be removed, such as forearc rotation and N-S
shortening in northern Cascadia that is now well documented (McCaffrey et al., 2000, 2007, 2012, 2013). Fortunately, these signals are well enough constrained today, that they can be removed with confidence, allowing investigation of the spatial extent and variability of interplate coupling. One the problem has been reduced to this level, the uncertainties and potential difference between interseismic signals and co-seismic release are similar to those of leveling/tide gauge geodesy. As an example of this, McCaffrey et al. (2013) showed two models that both fit the levelling and GPS data, that are quite different (Figure 3-3). Model Pn1d shows strong locking out to the trench, while Pn2d shows the minimum locking and a Gaussian distribution that fit levelling, tide and GPS data reasonably well. Pn1d accumulates moment for an M9 earthquake in ~ 300 years, Pn2d in ~ 500 years. Strong locking out to the trench is a common feature of many models, however the weak coupling suggested by the landward vergent, low wedge taper and other geologic indicators suggest this is unlikely (Goldfinger et al., 1992; 1996b; 1997). Gaussian or similar models like the McCaffrey Pn2d were used by Priest et al. (2009), and Witter et al. (2011; 2012) tsunami modeling exercises after those authors favored a geologically consistent approach suggesting lower coupling in the seaward wedge. In any case, these models illustrate clearly the lack of constraint placed on the offshore region by onshore geodetic data. Given that substantially more water would be displaced by pn1d, the dramatic difference in tsunami generation between these extreme models illustrates the importance of alternate methods of addressing offshore locking models.

Figure 3-3. Locking model results for two parameterizations of locking. (see McCaffrey et al. 2013 for functions). Colors and contours are of the slip deficit rate, in mm/yr. Slip deficit rate contours are 5, 15, 25, 35 and 45 mm/yr. (A) Tapered transition zone of variable width, depth and taper but locked to trench (pn1d). (B) Gaussian distribution of locking with depth (pn2d). From McCaffrey et al. (2013).
Paleogeodetic Data from the 1700 and Previous Earthquakes

These data have potentially the least ambiguous connection to the locked plate interface and slip on it for a given earthquake. However, they are limited by two things 1) the lack of inland penetration of the data to establish a deformation curve normal to strike, and 2) the precision of the data themselves. The coastal subsidence data along strike are located at numerous sites, forming a relatively dense but largely 1d line of data points. Few sites have any inland penetration, and thus the data are better at illuminating along-strike variability than at pinning down the landward end of the locked interface. Precision of the subsidence data has been quite low, ~0.5 m on a ~1m signal for much of the data published in the late 1980’s and 1990’s, compounding the lack of spatial density (e.g. Atwater, 1987; 1992, Atwater et al., 1997). Recently, more precision has been achieved (e.g. Englehart et al. 2013), improving the precision to 10-20 cm. These high-precision data are limited to just a few sites however, and are
insufficient to re-define the spatial evidence for subsidence per event given that inland penetration is still lacking. Several other uncertainties include the fact that it is not clear exactly when the subsidence occurred. At least part of it was coseismic, but some could also occur post-seismically over hours, weeks or months. Liquefaction of broad low-lying areas can also contribute to coastal subsidence, as can the motion of sympathetic crustal structures (e.g. McNeill et al., 1998). Locking models based on the past few events are a significant step in the right direction (e.g. Leonard et al., 2010, Priest et al., 2009, Wang et al., 2013). Currently such models are limited by relatively short coastal paleoseismic records, and the lack of spatial extent in the downdip direction due to the relatively linear Cascadia coastline. What is needed to better define earthquake slip models in Cascadia is investigation of the landward extend of paleo-ruptures.

**Other Proxies**

The remaining proxies, including positioning of gravity lows, the location of the forearc mantle corner, the upper limit of ETS events, the landward limit of smaller thrust earthquakes are difficult to evaluate even qualitatively. The gravity basin model proposed by Wells et al. (2003) and a similar model by Song and Simon (2003) suggest that the principal rupture patch locations can be predicted from gravity data that outlines forearc basins. However this hypothesis is controversial, and there are many exceptions to it. For example, nearly 80% of the seismic moment in the 20th century was found by Wells et al. (2003) to coincide with these gravity low/forearc basins. However, many of the individual models were poor fits, the slip models used were of low quality, and the statistic was dominated by a good fit by the 1960 Chile M9.5 earthquake, which dominated seismic moment during historic times. Since that time, the 2004 Sumatra earthquake and 2011 Tohoku earthquake were modern well-recorded examples that did not fit the model, and in the case of Sumatra, were anti-correlated, with slip patches being located under the forearc highs. The location of serpentinitized mantle at the upper limit of the mantle wedge may provide an absolute limit to downdip strain accumulation (e.g. Hyndman, 2013), but it’s location is hypothesized mostly based on velocity models and is speculative enough that it’s use in an investigation of downdip locking is limited. The locations of shallow thrust earthquakes could prove effective were there enough of them, but few than 10 such earthquakes exist in the Cascadia region, and thus cannot provide constraints to the downdip limit. The upper limit of ETS events is likely a good proxy for the absolute limit of downdip coupling. There is a wide band separating existing locking models and the upper ETS limit however, thus there is a discrepancy between this proxy and all others at present. In addition, slip deeper than the ETS limit cannot be excluded, and the very deep rupture of the 2004 Sumatra earthquake into the upper mantle (Singh et al. 2008) suggests that the limit of co-seismic slip may at times exceed the locking limit.

The distribution of episodic tremor and slip events (ETS) downdip of the locked interface (e.g. Boyarko and Brudzinski, 2010) may also reveal evidence of locking heterogeneity and segmentation. The spatial pattern of ETS events, particularly the updip limit of the ETS zone likely has a relationship to the locked interface, though that relationship is not clear. There is a significant gap between the updip limit of ETS, and the modeled downdip limit of plate locking (e.g. Gomberg et al., 2010). At the same time, there are excess easterly velocities in the GPS
data that have been considered an artifact (after block rotation is removed, R. McCaffrey pers. comm.) or attributed to plate locking (Chapman and Melbourne, 2009). Figure 3-8 shows the range of models that exist for the Washington area.

3.4 Margin Segmentation and Along-Strike Variability

3.4.1 Evidence from Geodesy and Seismology
Leveling surveys show that some forearc areas are tilting landward on a time scale of 70 years, but also tilting parallel to the arc, indicating heterogeneity in present day locking. Mitchell et al. (1994) and later Burgette et al (2009) calculated a highly variable uplift signal along strike in Cascadia. Recent evidence of episodic tremor and slip events (ETS) downdip of the locked interface (Brudzinski et al., 2007; Boyarko and Brudzinski, 2010) also may reveal evidence of segmentation. The significance of the debate about the configuration of the Cascadia locked zone is that there may or may not be seismic segments controlled by the thermal or structural boundaries and thus control slip distribution and tsunami generation. Segmented and whole-margin ruptures should leave distinctly different stratigraphic records in both the coastal marshes and the offshore turbidite-channel systems, which we discuss below.

3.4.2 Evidence from Paleoseismology
Several studies have examined the question of rupture mode diversity, including Kelsey et al. (2005), and Nelson et al. (2006) using onshore paleoseismic data. Goldfinger et al. (2008; 2012) integrated these data with offshore evidence from turbidites, using a more extensive spatial dataset, and extending the time range to the earliest Holocene. Correlated turbidites suggest ~19 long ruptures with an average recurrence interval for full-margin paleoseismic events (900–1,100 km in length) of ~500–530 years, with a variance ranging from ~200 to 1,200 years. A series of smaller ruptures, represented by thinner turbidites of lesser areal extent, can be correlated among southern Cascadia cores, and has moderately good correspondence with the presence of events of limited extent at coastal paleoseismic sites. These smaller events define three other margin segments that have recurrence intervals of 410–500, 300–380, and 220–240 years for segments with northern terminations at approximately 46°N (Nehalem Bank), 44°N (Heceta Bank), and 43° N (Coquille Bank; Figure 3-5. Goldfinger et al. (in revision) used
Figure 3-5 Segmented rupture model, revised from Goldfinger et al. (2012). This model reflects revision of the northern boundaries of Segments B, C, and D, with subdivision to include C’ and addition of segments E and F based on new core data (this study) and tsunami modeling at Bradley Lake (Priest et al. 2014). Marine core sites controlling rupture-length estimates are shown as yellow dots. Addition of several small ruptures in northern California are shown in Segment E, and a single northern rupture is identified off Washington in Segment F, both from Goldfinger et al., (2013). Preferred latitudinal limits shown with red shading. Estimated minimum and maximum limits shown with dashed lines. Widths and up and downdip limits approximate. Widths and up and downdip limits approximate. Paleoseismic segmentation shown also is compatible with latitudinal boundaries of episodic tremor and slip (ETS) events proposed for the downdip subduction interface (Brudzinski and Allen, 2010) and shown by white dashed lines.

additional data to refine the earlier segment boundaries. Using an expanded correlation model for Washington, several interesting results emerged that suggest revision of the segmentation model of Goldfinger et al. (2012). In Figure 3-5 we show the revised segmentation model of Goldfinger et al. (2016). The northern limits of segments B and C were originally data limited, and have been revised northward. These results also include several additional Segment D ruptures, and a single northern segment rupture reported in Goldfinger et al. (2013). The
northern limit of Segment D ruptures based on observations and modeling of tsunamis at Bradley Lake, Oregon reported in Priest et al. (2014). Overall, the turbidite based earthquake record from the Washington/Canadian margin is to a large degree compatible with the paleoseismic records onshore at Willapa Bay and other locales as discussed in Goldfinger et al. (2012) Leonard et al. (2010), Enkin et al. (2013), Hamilton et al. (2015), and now includes the northern Washington coast at Waatch Marsh (Peterson et al. 2013).

3.5 Arc-Parallel Crustal Shortening of the Forearc
Paleomagnetic and GPS data have revealed a forearc block rotating clockwise 1°/my and moving northward in Cascadia (Wells et al 1998, McCaffrey et al 2007; 2013). Between the northward migration of the Oregon forearc block and the “backstop” of cratonic Vancouver Island and southern British Colombia, 4.4 ± 0.3 mm/yr. of permanent shortening occurs (McCaffrey et al., 2007). N-S compression has resulted in several E-W trending structures in the Seattle area (Figure 3-8). The Seattle Fault zone has paleoseismic evidence a large surface rupture of (~7.0 Mw) with ~7 m of slip about 1100 years ago. This earthquake resulted in landslides in the Puget Lowlands, turbidity currents in Lake Washington, and generated a local tsunami (Bucknam et al., 1992; Atwater, 1992; Karlin and Abella, 1996). The Seattle Fault zone includes the Waterman Point fault, Toe Jam Hill fault, Islandwood fault, and Restoration Point marsh all of which have paleoseismic evidence of Holocene rupture (Nelson et al., 2003, Sherrod et al., 2007, Sherrod et al., 2000). The Tacoma fault is closely related to the Seattle fault, and may rupture independently or synchronously with a similar magnitude. These faults represent secondary faults to a large blind thrust fault below the Puget Lowlands (ten Brink et al 2006, Brocher et al., 2004). Coseismic rupture may occur during earthquakes on the blind thrust or interseismically due to strain accumulation during folding of the overlying strata (Kelsey et al. 2008; 2012).

Earthquakes at such a close range to metropolitan Seattle, though likely less frequent than subduction zone earthquakes (recurrence rate ~ 1 per 7000 yr.), are a severe hazard, that may damage up to 80 bridges, a large URM building stock, and generate tsunami, liquefaction, lateral spreads and landsliding throughout the area.
Other indications of N-S compression include focal mechanisms that show thrust solutions with ~ N-S P axes, and borehole breakouts showing similar horizontal compression (Wang et al., 1995, Wang, 2000; Werner et al., 1990, McCrory et al. 2002; Figure 3-6, Figure 3-7). While these lines of evidence confirm the contemporary stress field is ~ N-S in the forearc, they are far too sparse to be useful in delineating stress domains in detail.
Figure 3-6. (a) A summary of the stress regime of the Cascadia forearc. Small solid and open circles on map are continental earthquakes (M<3) used by Mulder (1995) and Ma (1988), respectively, to determine focal mechanism solutions shown in insets. Large open circles on map and in insets are events (M≥4) since 1993 for which moment tensor solutions have been provided by the Oregon State University (http://quakes.oce.orst.edu, 1998). Large arrow pairs indicate the orientation of maximum compressive stress from various stress indicators (see text for details). (b) A summary of geodetic strain rate measurements in the Cascadia forearc. A thin solid bar indicates contraction, and a thick solid bar indicates extension. A hollow bar indicates maximum contraction where only shear strain rates were determined, assuming uniaxial contraction. The largest strain rate shown is 2.3×10^{-7} year^{-1}. From Wang, 2000.
Figure 3-7. Orientation of $s_{H\text{max}}$ as determined by borehole breakouts from 18 wells in western Oregon and Washington. The wells are located where the arrows meet, and the numbers refer to Table 2 of Werner et al. (1991). Wells 5 and 14 have bimodal directions of $s_{H\text{max}}$ shown by two sets of convergent arrows. Location of offshore faults and folds which deform Pleistocene-Miocene sediments are from Peterson et al. (1986). Figure from Werner et al. 1991.

4. Methods

4.1 Structural Mapping
We have compiled new and existing multichannel, single channel and CHIRP seismic reflection data and sidescan sonar surveys to improve the structural map on which this project is largely based. 26 geophysical surveys have been identified as relevant to the structural analysis of the Cascadia forearc and to regional tsunami modeling. The digital surveys have been integrated into both ArcMap and IHS/Kingdom digital projects to enable processing, visualization and interpretation. Of these, 14 single and multichannel campaigns representing over 30,000 km of survey have been reviewed in depth: L-5-77, L-11-80-wo, and L-5-81 collected by USGS 1977-1980; W-6-75NC, W-8-75NP, W-9-78-NC, W-18-75NP, W-29-80WO, and W-39-85WO, collected by WesternGeco 1975-1985 (available from https://walrus.wr.usgs.gov/NAMSS/, Triezenberg et al., 2016); Digicon-OSU 1989, and MGL 1212, collected through the 2012 NSF-funded Cascadia Open Access Seismic Transects (COAST) project, and industry profiles from Shell, Chevron, and Texaco (Figure 4-2). Many other surveys were used for structural mapping, including analog sparker surveys by Shell and Oregon State University, analog single and multichannel seismic surveys by University of Washington Oregon State University, University of California at Santa
Cruz, University of Hawaii, University of Washington, Scripps Institution of Oceanography, U.S. Geological Survey, and the U.S. National Oceanic and Atmospheric Administration (Figure 4-3, Figure 4-4, Figure 4-5). We used a mapping scale of 1:500,000, however many datasets have far greater resolution and the data were used at full resolution in the interpretations.

The seismic-reflection profiles vary widely in quality, depth of penetration, and navigational accuracy, and range from single-channel sparker records navigated with Loran A to 144-channel digital profiles navigated with GPS. Several detailed previous surveys have been incorporated in this compilation. For example, in preparation for Ocean Drilling Program (ODP) Leg 146, a site survey consisting of closely spaced, high-resolution multichannel seismic lines, SeaMARC 1A sidescan mapping, multibeam swath bathymetry, and ALVIN submersible dives focused on the plate boundary and accretionary wedge near the proposed drilling sites near 45° N. latitude (Figure 4-4). Within this 6,000-km² area Goldfinger et al. (1997) and MacKay (1995) mapped submarine structures in considerable detail.

Table 1. Data sources and navigation accuracy

<table>
<thead>
<tr>
<th>Data Source</th>
<th>Navigation System</th>
<th>Approx. Nav. Error</th>
</tr>
</thead>
<tbody>
<tr>
<td>U.S.G.S MCS/GLORIA</td>
<td>Transit/Loran C</td>
<td>300-1500m</td>
</tr>
<tr>
<td>OSU-U/W Sparker SCS</td>
<td>Loran A</td>
<td>1000-3000m</td>
</tr>
<tr>
<td>OSU/Digicon MCS</td>
<td>GPS</td>
<td>100m</td>
</tr>
<tr>
<td>Chevron Oil MCS</td>
<td>Transit/Loran C</td>
<td>0-500m</td>
</tr>
<tr>
<td>Shell Oil Sparker SCS</td>
<td>SHORAN</td>
<td>50 m</td>
</tr>
<tr>
<td>U.S.G.S. boomer</td>
<td>Differential GPS</td>
<td>&lt; 20m</td>
</tr>
<tr>
<td>Western Geco W-6-75NC</td>
<td>Integrated (SINS) TRANSIT, Loran, inertial</td>
<td>~ 50m</td>
</tr>
<tr>
<td>Western Geco W-8-75NP</td>
<td>Integrated (SINS) TRANSIT, Loran, inertial</td>
<td>~ 50m</td>
</tr>
<tr>
<td>Western Geco W-9-78NC</td>
<td>Integrated (SINS) TRANSIT, Loran, inertial</td>
<td>~ 50m</td>
</tr>
<tr>
<td>Western Geco W-18-75NP</td>
<td>Integrated (SINS) TRANSIT, Loran, inertial</td>
<td>~ 50m</td>
</tr>
<tr>
<td>Western Geco W-29-80WO</td>
<td>Integrated (SINS) TRANSIT, Loran, inertial</td>
<td>~ 50m</td>
</tr>
<tr>
<td>Western Geco W-39-85WO</td>
<td>GPS</td>
<td>~ 30m</td>
</tr>
<tr>
<td>Langseth MGL 1212</td>
<td>Differential GPS</td>
<td>&lt;5 m</td>
</tr>
<tr>
<td>NOAA/NOS Multibeam</td>
<td>ARG0</td>
<td>50 m</td>
</tr>
<tr>
<td>OSU 1990's Multibeam</td>
<td>TRANSIT/GPS/Loran C</td>
<td>300-1500m</td>
</tr>
<tr>
<td>OSU 50 kHz sidescan</td>
<td>GPS</td>
<td>100 m</td>
</tr>
<tr>
<td>OSU 150 kHz sidescan</td>
<td>CA code GPS</td>
<td>~ 50 m</td>
</tr>
<tr>
<td>OSU SeaMarc 1A</td>
<td>P code GPS, USBL</td>
<td>50-200 m</td>
</tr>
</tbody>
</table>
The sidescan and bathymetric surveys were navigated with a combination of GPS and in the case of older surveys, with GPS and Transit satellite navigation, with Loran C tracking used between satellite fixes for some older data. Navigation of several surveys using deep-towed sidescan towfish was by the method described by Appelgate (1988). Where spatial misfits occurred, we adjusted the sidescan data to best fit the GPS navigated MCS lines or the multibeam bathymetry where appropriate. SeaMARC 1A sidescan data were collected with a deep-towed 30-kHz system capable of imaging a 2-km or a 5-km swath width, with spatial resolutions of 1 and 2.5 m respectively. With older surveys, navigational accuracy was more variable. Loran A navigated profiles have maximum errors on the order of 1-3 km, Loran C errors are about 0-1.5 km, and Transit satellite errors range from near zero up hundreds of meters. The dense coverage of reflection profiles allowed adjustment of older lines where crossed by satellite-navigated lines (Goldfinger, 1994).

The sidescan and bathymetric surveys were navigated with a combination of GPS and Transit satellite navigation, with Loran C tracking used between TRANSIT satellite fixes. GPS navigation during high resolution sidescan surveys and DELTA dives was continuous. Navigational accuracy for GPS navigated dives and sidescan swaths is 100 m (due to selective availability, the error introduced into the civilian use signal by the Department of Defense in the 1990’s). Navigation accuracy for the seismic reflection profiles varies according to the systems in use at the time. Transit satellite errors range from near zero up to 200 m, Loran C errors are approximately 0-1.5 km, Loran A navigated profiles have maximum errors on the order of 1-3 kilometers. Shell Oil Company lines were navigated with a company SHORAN radio navigation system. Horizontal errors with this system are approximately 50 meters. Although these profiles were shot in 1961-62, their navigational accuracy is good. The dense coverage of reflection profiles allowed adjustment of older lines where crossed by satellite-navigated lines. The 1980’s NOAA/NOS multibeam bathymetric surveys were navigated with an ARGO shore based system; navigational accuracy is 50 m. The bathymetry is accurate to within 1% of the water depth across the swath. Table 1 lists the seismic reflection, bathymetric and sidescan data used in this study.

### 4.2 Inshore Bathymetric Mapping

While this study is largely based on a compilation of existing data, several surveys were conducted by us or in coloration with other groups during the project period that bear on this work. These studies are unpublished, thus we include here a brief description of the methods.

#### 4.2.1 Vessels

The 85-foot (LOA) *R/V Pacific Storm*, owned and operated by OSU’s Marine Mammal Institute (MMI), was contracted as a mapping vessel for several surveys spanning 2009-2012 for the Oregon State Waters Mapping Program (ORSWMP) and for a mapping project with BOEM for wave energy siting.
4.2.2 Equipment

OSU’s Active Tectonics and Seafloor Mapping Lab (AT&SML) provided a NavCom SF-3050 StarFire™ GPS with an ALL-AREAS (Global) correction service subscription providing decimeter precision horizontal positioning. An Applanix POS MV 320 Inertially-Aided Real-Time-Kinematic motion reference unit, Seabird SBE-19Plus CTD, and Reson SVP 70 sensor. Reson Sea-Bat 8101, Kongsberg EM 2040 and R2sonics 2025 broadband multibeam systems were used with snippets backscatter employed. AT&SML provided the CARIS HIPS and QPS FMGeocoder Toolbox© (FMGT) for bathymetric and backscatter processing.

Vessel motion was measured by an Applanix POS/MV 320 inertial measurement unit during all surveys. The POS/MV system utilizes an inertial motion unit (IMU) and L1 and L2 carrier phase measurements from multiple GPS antenna arrayed on the vessel to produce an Inertially-Aided Real-Time Kinematic (IARTK) attitude and position for the vessel. The system is used for ships position, heading, and to determine roll, pitch, yaw attitude as well as heave. The Reson 8101 utilizes real-time roll data supplied by the POS/MV, to beam steer outgoing pings such that they are formed normal to the seabed. A NavCom StarFire SF 3050 GPS system was used as the primary position system (primary GPS) input for the POS/MV. The NavCom Starfire system is a commercial satellite based differential system known as GSBAS (Global Satellite Based Augmentation System). Access to the system is through a subscription service providing positional accuracy of ~ 10 cm horizontal, and 15 cm vertical worldwide. Use of the GSBAS system eliminates the need for land based base stations, or location dependent differential signals such as the Coast Guard differential beacon system. Continuous ‘real-time” sound speed measurements were made with a sound-speed probe at the Reson 8101 transducer head, a particularly important place to measure sound speed due to the physics of forming multiple sonar beams. Water column profiles of sound-speed were collected at roughly 3 hour intervals and within the survey area using a Seabird SBE 19 CTD.

Bathymetric data were acquired using Hypack/Hysweep 2010 in .81X, .HSX, and .RAW formats. Hypack integrates the incoming bathymetric data, time stamped by the sonar sensor, and the incoming navigation data from the Starfire DGPS and PosMV motion sensors to generate a ping-by-ping data record with integrated navigation and vessel motion data. Bathymetric data from all surveys were processed using CARIS (http://www.caris.com/products/software.cfm/prodID/1) HIPS/SIPS) data processing software in order to produce tide, motion- and sound-speed-corrected, geo-referenced bathymetry and backscatter imagery. Backscatter mosaics were generated with Interactive Visualization Systems (IVS) Geocoder version 7 software to additionally produce backscatter mosaics that incorporate geometric and beam pattern corrections, as well as removing artifacts of gain and pulse length changes and topography during the survey (Fonseca and Calder, 2005; Fonseca and Mayer, 2007). The data are being collected using standard hydrographic protocols (NOAA Field Procedures Manual, 2010). Some of the primary study areas are listed below:

OOI Inshore, WA; This nearshore rocky reef complex and adjacent sedimentary shelf was mapped over three years, 2009, 2010, and 2011, under funding from the Ocean Observing Initiative. In 2001, 13 Shipek grabs samples were collected and in the northern one third of the survey area.
Cape Falcon Fault, OR; The mid-shelf site was mapped in 2011 from the R/V Pacific Storm using a Reson 8101 (240 kHz) multibeam sonar. The data coverage follows a thin zone of rock outcrop (Cape Falcon Fault) and connects 2009 OSWMP nearshore multibeam data coverage to 2002 Ocean Explorer offshore multibeam data coverage. No samples or ROV video is available within the survey coverage.

Stonewall Bank, OR; This mid-self rocky bank was surveyed in 2012 from the R/V Pacific Storm using a Kongsberg EM2040 (300 kHz) multibeam sonar. Oregon Department of Fish and Wildlife funded the survey and provided drop camera points and habitat types for seabed classification purposes. No sediment samples were collected.

Coquille Bank, OR; This outer-shelf bank was surveyed from the R/V Thomas Thompson in 2005 using a Kongsberg EM300 (30 kHz) multibeam sonar. NOAA Fisheries Northwest Fisheries Science Center funded data collection including seabed AUV dives for seabed classification. No samples were collected.

H12130 & H12131, OR; This nearshore area of Oregon’s State Waters was surveyed by Fugro Inc. under funding from NOAA Office of Coast Survey. Available data include multibeam bathymetry. Fugro collected but did not process backscatter for this survey. OSU has partially processed the backscatter but technical issues with the data have prevented OSU from deriving a backscatter map from the data for classification use. No samples or video were collected at this survey site.

4.3 Deep Water Mapping

4.3.1 Cascadia Initiative TN 265 2011

On Cruise TN 265, conducted in the R/V Thomas G. Thompson, operated by the University of Washington, we mapped approximately 28,500 km² of continental shelf, slope and abyssal plain off Vancouver Island, Oregon Washington and northern California. The depth ranges in the survey were between 26 and 3167m (mean survey depth = 1704m). The system used was the onboard Kongsberg EM302 multibeam sonar. The EM 302 system is a 30 kHz mid-range system that covers depth ranges from 10-7000m, with optimal performance between ~ 100 and 3000m. The EM302 system is a fully yaw pitch and roll stabilized beam-forming sonar, and has up to 864 soundings per ping. The swath width ranges from ~ 3-5 x water depth in normal operation. During the cruise we also acquired over 9,000 line km of Knudsen 3260 CHIRP sub-bottom profiles.
Figure 4-1. Compilation of high resolution multibeam data on the Washington margin. The major component is the TN 265 and MGL 1212 surveys conducted in 2011-2012.

Data Processing

Bathymetry data processing
Multibeam data was partially processed at sea. Sound speed corrections were applied from XBT profile data collected in real time using the Kongsberg SIS data acquisition package. XBT casts were done ~ 4 times per day. Predicted tides were applied in CARIS post processing by using NOAA zone definitions and predictions from the La Push, WA, Newport, Oregon, and Cape Blanco, OR tide stations. Soundings were “ping” edited using the CARIS swath and base surface editor tools by multibeam sonar watch standers (at sea) and by lab personnel (post cruise). All TN265 multibeam data has been gridded to 50m pixel resolution into a single Digital Elevation Model (DEM).

Bathymetry Surface Creation
We have also collected all Kongsberg EM300 or EM302 data available either online through the NGDC portal or in-house from past survey work in an effort to produce a seamless dataset of high, 50m, resolution for the survey area. This product is now complete up to July 2011. To this EM300-302 West coast surface, we’ve added modern multibeam data collected by NOS on Washington and Oregon outer coast (generally within 3nm from shore) and our own Oregon
State Waters and OOI continental shelf multibeam. Adding these datasets completes the surface to include all known sources of modern high resolution bathymetry data (excludes NOAA soundings and data described below). The bathymetry surface described above is a compilation of modern and high resolution data. It reveals seabed features in great detail over previously unknown areas, however it is not a continuous surface and has large holes where historic or lower resolution data must be used to complete the picture. Therefore our final outstanding tasks are to integrate data collected on older expeditions (Thompson Hydrosweep, Laney Chouest SeaBeam, Auriga, among others) and the NOAA soundings database to yield a regional contiguous best available bathymetry surface. To date we have produced several draft surfaces and are working to reconcile significant vertical offsets encountered when merging historic data sources with modern sources.

**Backscatter Processing**
All EM300, EM302, Oregon State Waters Reson 8101, OOI Reson 8101 backscatter data was processed using the QPS Fledermaus FM Geocoder Toolbox. FM Geocoder produces backscatter mosaics corrected for topographic effects and produces excellent quality image mosaics. To this mosaic we’ve added all available NOS and Olympic Coast National Marine Sanctuary backscatter data creating a backscatter surface of near identical footprint and coverage as the modern bathymetry product described in the previous section. All raw cruise multibeam data is available through NGDC and through the Rolling Deck to Repository (R2R) at: [http://www.rvdata.us/catalog/TN265](http://www.rvdata.us/catalog/TN265)

**Knudsen Sub-Bottom Profiles**
Knudsen 3260 3.5 kHz CHIRP Sub-Bottom data was collected in both Knudsen proprietary .keb and .keb formats, as well as .segy (RAW Format). All 9,000 line kilometers of Knudsen data was processed using UCSD’s Sioseis software package. Sioseis was used to correct of seafloor alignment errors (heave and water bottom) as well as to produce envelope (stacked and filtered) data.

**4.3.2 COAST Survey: MGL 1212 2012**

**Multibeam Bathymetry**
We created high resolution (15 x 15 m) bathymetric grids covering a portion of the Cascadia accretionary wedge using multibeam bathymetric data collected for the Cascadia Open Access Seismic Transects (COAST) and Cascadia Initiative projects (Holbrook et al., 2012; Toomey et al., 2014). Multibeam bathymetric and backscatter data were collected on the R/V *Thomas G. Thompson* using a Kongsberg EM302 multibeam system, and on the R/V *Marcus G. Langseth* using a Kongsberg EM122 multibeam system. Both ships used high-precision Differential Global Positioning System (DGPS) and real-time attitude corrections. Expendable Bathymetric Thermographs (XBTs) were deployed twice daily, at minimum, to record and correct for water-column, sound-velocity variations. Multibeam data collected during the COAST cruise were especially useful because they were collected contemporaneously with multi-channel seismic (MCS) data at a slow speed of ~4.5 knots (Holbrook et al., 2012), resulting in a much greater
along-track sounding density. Combining these two datasets along with other modern multibeam datasets has allowed for a near complete high-resolution bathymetry grid to be created for the Cascadia margin. It is this combination grid that supported our more focused study, providing regional context to submarine features.

**Seismic Reflection**

A-hull mounted Knudsen 3260 sub-bottom profiler, with a frequency sweep of 2 to 6 kHz, recorded three seismic profiles crossing two frontal thrust/anticline erosional features on the *R/V Thomas T. Thompson* cruise TN265 in 2011. Processing of these data included application of a bandpass filter and a time-varying gain. Multichannel seismic (MCS) data were acquired with 36-airguns, 6600-cubic-inch array which provided an acoustic signal recorded on a 636-channel, 8-kilometer streamer sampling the subsurface every 6.25 meters (Holbrook et al., 2012). These data were pre-stack depth migrated following conventional seismic processing steps including: band-pass filtering, trace editing, source deconvolution, multiple elimination, velocity analysis, and horizon-based tomographic velocity updates (Figure 4-3; Yilmaz, 2001).

![Explanation](image)

*Figure 4-2. Trackline map explanation for Figure 4-3, Figure 4-4, Figure 4-5.*
Figure 4-3. Northern Cascadia seismic reflection trackline map. Areas and profile segments of other figures are shown in white.
Figure 4-4. Central Cascadia seismic reflection trackline map. Areas and profile segments of other figures are shown in white.
Figure 4-5. Southern Cascadia seismic reflection trackline map. Areas and profile segments of other figures are shown in white.
5. Results

5.1 Northern Cascadia Structure

5.1.1 Overview

While many of the features identified in this initial review and the general structural trends observed thus far (Figure 5-1) are broadly consistent with previous structural studies of Cascadia's forearc - including location of landward vergent thrusts, shelf basins and large scale slope failures (Fleuh et al., 1998, Adam et al., 2004, Fisher et al., 1999; Goldfinger, 1994; Goldfinger et al. 1992a, 1992b, 1996a, 1996b, 1997, 2000; Gutscher et al. 2000, MacKay, 1995; McNeill et al., 1997). While a complete synthesis is beyond the scope of this project, the new integrated interpretation includes new features, and attempts to synthesize some of the elements of the complex structure of Cascadia.

The northern Cascadia margin off Washington is an exceptionally complex structural system, that includes 1) an older and younger accretionary wedge, separated by a sharp boundary; 2) an unusual landward and mixed vergence outer wedge; 3) an arc-parallel dextral strike-slip fault; 4) outer shelf and upper slope large listric normal faults that underlie a region of seaward gravity driven extension of the upper slope; 5) an upper slope and shelf punctuated with numerous diapirs; 6) a transition from arc parallel to arc normal structures related to arc parallel compression and the growth of the Olympic mountains and 7) WNW striking transverse strike-slip faults originating in the lower plate that deform parts of the upper plate (Figure 5-1). This complex amalgam of structures is not partitioned by time, all of these features interact and are active in the Holocene simultaneously.

<table>
<thead>
<tr>
<th>Explanation</th>
<th>Channels (nestec_15_arc)</th>
<th>activity</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>active</td>
<td></td>
</tr>
<tr>
<td></td>
<td>inactive</td>
<td></td>
</tr>
<tr>
<td></td>
<td>intermittent</td>
<td></td>
</tr>
<tr>
<td></td>
<td>unknown</td>
<td></td>
</tr>
<tr>
<td>Structure</td>
<td>Active</td>
<td></td>
</tr>
<tr>
<td>Linetype</td>
<td>Continuous</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Concealed</td>
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</tr>
<tr>
<td></td>
<td>inferred</td>
<td></td>
</tr>
<tr>
<td>Pleistocene</td>
<td>Linetype</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Continuous</td>
<td></td>
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<tr>
<td></td>
<td>Concealed</td>
<td></td>
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<tr>
<td></td>
<td>inferred</td>
<td></td>
</tr>
<tr>
<td>Pre_Pleist</td>
<td>Linetype</td>
<td></td>
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<tr>
<td></td>
<td>Continuous</td>
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<tr>
<td></td>
<td>Concealed</td>
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<tr>
<td></td>
<td>inferred</td>
<td></td>
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<tr>
<td>Stress Transition</td>
<td>Shelf Edge</td>
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</table>

Figure 5.1a. Legend for structure maps shown in Figure 5-1, Figure 5-14, Figure 5-15.
Figure 5-1. Structural geology of the northern Cascadia margin.
The fortuitous inclusion of data from the recent 2012 COAST project in this review provides eight new transects which cross the deformation front and low traction province off Washington, enabling greater structural resolution of this critical region than previously possible. Common to the new COAST transects, and to the USGS and WesternGeco transects crossing the low traction region, is the presence of a major synthetic reverse fault, identified as a splay fault located between the Pleistocene and older wedge segments (Figure 5-16).

On the Washington shelf, new bathymetric, multichannel, sidescan and high-resolution seismic data now exist that will significantly improve mapping of the Washington margin. In our original mapping of the Washington margin (Goldfinger et al., 1997; McNeill et al. 1997), we encountered significant difficulties with the Washington shelf, as this area had mainly analog and paper copies of Western Geco multichannel seismic profiles, and no bathymetric data at that time. Since then, several key studies have improved the situation. McCrory (2002) mapped a small but important section of the Washington shelf using new sidescan and boomer data collected in 1997 and 1998, and resolved the strike of some inshore structures that had baffled us in the late 1990’s. These E and NE striking structures are now confirmed by new multibeam mapping we conducted in 2009 and 2010 as part of the site surveys for the NSF supported Ocean Observing (OOI) project (http://oceanobservatories.org/). With this key puzzle piece now confirmed with multiple datasets, updating (and in some cases correcting) our late 1990’s mapping is now complete, and forms an important part of the tectonic fabric of northern Cascadia.

Using the new 2011 R/V Thompson and 2012 R/V Langseth multibeam data, we have conducted a preliminary structural interpretation of the seaward most portion of the forearc in Washington, updating and improving our 1997 interpretation. While the major elements have not changed, important details have been revealed, including a new estimate of the slip rate on the frontal thrust (see Task 3).

### 5.1.3 Two Phase Accretionary Wedge

The submarine forearc is at its widest, 150 km off the northern Olympic Peninsula of Washington, where the thick Pleistocene Astoria and Nitinat Fans presently are being accreted to the margin (Figure 5-1). The active accretionary thrust faults of the lower slope are characterized by mostly seaward-vergent thrusts on the Oregon margin from 42° to 44°55’ N and north of 48°08’ N off Vancouver Island, and by landward-vergent thrusts between 44°55’ and 48°08’ N, on the northern Oregon and Washington margins (Carson, 1971, Silver 1972, Goldfinger et al., 1994; 1997, Mackay, 1995, Adam 2004).

The new mapping presented here, further details the distribution of landward and mixed vergence structures, utilizing new multibeam bathymetry, sidescan sonar and recent seismic reflection data (Figure 5-1). Landward vergence may result from a seaward dipping backstop, but in this case, the landward-vergent province is most likely related to subduction of rapidly deposited and overpressured sediment from the Nitinat and Astoria Fans (Seely, 1977; MacKay, 1995; Goldfinger et al, 1997; Adam et al., 2004). Off Washington and northern Oregon, the broad accretionary prism is characterized by a low wedge taper and widely spaced landward vergent accretionary thrusts and folds (which offscrapes virtually all of the incoming sedimentary section). Sparse age data suggest that this prism is Quaternary in age and is
building westward at a rate close to the orthogonal component of plate convergence (Westbrook, 1994; Goldfinger et al., 1996b; 1997), however new data acquired in this project presents a revised construction rate of the wedge, discussed below. This young wedge abuts a steep lower-slope upper-slope break that separates it from the Eocene-Pliocene accretionary complex and the continental shelf. The landward limit of mixed and landward vergence coincides in places with this break, and in other places the landward and seaward structures abut directly near the rear of the lower slope wedge (Figure 5-1). The thick incoming sediment package has been folded and thrust into a broad accretionary wedge offshore Washington and northern Oregon. The rapid accretion of low amplitude landward vergent folds has resulting in a wedge with low to slightly negative surface taper, an unusual morphology exaggerated by an increase in the thickness of the accreted section with time during the Pleistocene. As a result, all of the upper slope Washington Canyons except Juan de Fuca and Quillayute feed into a margin parallel channel that in turn feeds Willapa Canyon/Channel (Figure 5-1).

North of the mixed vergence province on the Canadian margin, vergence is mostly seaward, with a steeper narrower younger wedge, except for a landward vergent segment centered at ~49 N. (Gao et al., 2015, Yorath et al., 1987; Clowes et al., 1987; Davis and Hyndman, 1989, Spence et al., 1991a, 1991b; Hyndman et al., 1994; Yuan et al., 1994). Structures in the lower slope region are relatively uniform in strike, averaging ~ 345 deg., and sub-parallel to the Washington coast. Structures landward of the lower slope-upper-slope break are highly variable in strike, but are generally more northerly trending. Several provinces of distinctly different strike include 1) a region of the central Washington inner shelf striking SW-NW averaging E-W. These structures are discussed below; 2) A region of variable strike within a seaward extending “bulge” of downslope extension; 3) a region of NW striking folds along part of the North Nitinat strike-slip fault (Figure 5-1). The boundary between the upper and lower slope provinces is in most places, a seaward vergent thrust. This structure constitutes a major “splay” fault separating the provinces, and likely was the plate boundary thrust prior to the construction of the Pleistocene wedge.

5.1.3.1 Truncation of Structures at the Upper-Slope-Lower-slope Break
In several locales, structural trends in the upper slope province are truncated or terminate against structures of the lower slope. In some cases, this truncation is acute, with 90 degree difference in strike (Figure 5-1). Most other truncations are at angles of 10-40 degrees, and are pervasive in some areas, particularly northern Oregon. The structural truncations may in a few cases be the result of downslope extension, but for the most part, this phenomenon represents a more pervasive strike difference. We suggest that the structural truncation likely represents subduction erosion along the former plate boundary thrust, likely of Pliocene age. Outboard of what was previously a steeper, and in places erosional margin, the Pleistocene wedge was accreted rapidly, creating the structural juxtapositions. The origins of the differences in strike in the older wedge are unknown.

5.1.3.2 Activity Rate of the Cascadia Splay Fault
This feature represents a profound boundary between the younger (Pleistocene to Holocene) and older wedges. In some sense it is the Cascadia Backstop, but it is also a mappable active seaward vergent thrust fault. This boundary was likely the plate boundary prior to the influx of
accreted fan material, so one would expect it to be a major player in the system still. Evidence of recent surface rupture is sparser than previously thought. Instead, close examination of each seismic crossing shows that slow, blind movement is much more common. Table 2 shows this for each of these crossings and notes. Some of the comments are general because offsets aren't quantified, and may not be quantifiable. The observations extend from 44°53' northward to almost 48°N. South of that, the splay either doesn't exist, or is buried by major submarine landslides, or both. South of the landslide province in central/southern Oregon, there is no evidence of a major splay system; a more typical fold thrust belt exists without standout splay faults.

The lack of offset in many of the basins directly seaward of the fault, even in Washington, implies a slow Quaternary slip rate. These basins currently fill at sedimentation rates of less than 1 mm/yr., and the splay fault is not keeping up with that in most places. The Pleistocene rates though are much higher in some places, so the maximum slip-rate could be higher there. The splay in places traverses basins with no Pleistocene influence, and with no detectable surface rupture. The few places where the fault appears more active suggest a heterogeneous activity, which could be modulated by other local tectonics. For example, the major normal faults in Washington add a downslope rate of convergence to the shortening rate affecting lower slope structures as they move downslope to the west, perhaps enhancing convergence there. There may be other local effects, especially given the prevalence of strike-slip faulting which is emerging from the current mapping. The splay is certainly a major structure, with large total offset imaged in the seismic sections. With minimal evidence of surface rupture, its slip-rate must still be fairly low (Figure 5-16).

Table 2. Activity Notes, Cascadia Splay Fault.

<table>
<thead>
<tr>
<th>Line</th>
<th>Type</th>
<th>Latitude</th>
<th>Slip Rate</th>
<th>Surface Rupture</th>
<th>Labeled Rupture</th>
</tr>
</thead>
<tbody>
<tr>
<td>L1</td>
<td>Multichannel, seismic</td>
<td>42°13′44″</td>
<td>Not detected</td>
<td>Not detected</td>
<td>Line crosses obliquely. Splay present at base of landslides; not faulted.</td>
</tr>
<tr>
<td>L2</td>
<td>Multichannel, seismic</td>
<td>44°10′36″</td>
<td>Not detected</td>
<td>Not detected</td>
<td>Line not visible.</td>
</tr>
<tr>
<td>W2</td>
<td>Multichannel, seismic</td>
<td>44°51′30″</td>
<td>Small</td>
<td>Observed</td>
<td>Splay present at base of landslides.</td>
</tr>
<tr>
<td>WS</td>
<td>Multichannel, seismic</td>
<td>44°52′30″</td>
<td>Large</td>
<td>Observed</td>
<td>Splay present at base of landslides.</td>
</tr>
<tr>
<td>L3</td>
<td>Seismic, analog</td>
<td>45°53′30″</td>
<td>Not detected</td>
<td>Not detected</td>
<td>Line not visible.</td>
</tr>
<tr>
<td>L4</td>
<td>Seismic, analog</td>
<td>45°31′30″</td>
<td>Small</td>
<td>Observed</td>
<td>Splay present at base of landslides.</td>
</tr>
<tr>
<td>L5</td>
<td>Seismic, analog</td>
<td>45°24′30″</td>
<td>Not detected</td>
<td>Not detected</td>
<td>Line not visible.</td>
</tr>
</tbody>
</table>

Other observations:

- Fault cuts across the lower part of Neskowin canyon where it appears to control a small landslide. No seismic crossings are present, but 35 m multibeam and backscatter show a relatively fresh looking scar, the scar though appears to be older slip at least in part.
5.1.4 Landward and Mixed Vergence

As summarized by Zhou et al. (2016), several factors have been suggested as influences in the style and vergence of fold-thrust belts, including the strength and dip of basal décollements (Davis et al., 1983; Dahlen, 1990; Cobbold et al., 2001), thickness and strength of brittle overburden (Davis et al., 1983; Dahlen, 1990; Liu et al., 1992; Teixell and Koyi, 2003), the strength ratio of hanging wall to basal décollement (Bonini, 2003; Couzens-Schultz et al., 2003), geometry and strength of the backstop (Byrne et al., 1993; Lallemand et al., 1994; Bonini et al., 2000; Rossetti et al., 2002), and loading or unloading by sedimentation or erosion (Storti and McClay, 1995; Mugnier et al., 1997; Persson and Sokoutsis, 2002; Stockmal et al., 2007; Cruz et al., 2010; Smit et al., 2010), as well as surface slope and topography (Marques and Cobbold, 2002, 2006). The landward-vergent province of the northern Oregon and Washington lower slope has most commonly been related to subduction of rapidly deposited and overpressured sediment from the Nitinat and Astoria Fans (Seely, 1977; MacKay, 1995; Goldfinger, Silver, 1972, Adam et al., 2004).

Off Washington and northern Oregon, the broad accretionary prism is characterized by a low wedge taper and widely spaced landward vergent accretionary thrusts and folds which offscrape virtually all of the incoming sedimentary section (Figure 5-2, Figure 5-3). Abundant evidence of overpressuring in the region exists. In the new mapping a number of mud volcanoes have been identified, and are primary evidence of overpressuring of the accretionary wedge. A mud volcano straddles the North Nitinat Fault near a right (restraining) bend 9 km seaward of the deformation front (Goldfinger et al. (1997). A 2 km sidescan swath revealed a submarine channel parallel to the base of the continental slope on the abyssal plain. This channel is cut and offset 150 m left-laterally by the NNF with in turn cuts and offsets the oblong fault-parallel mud volcano (Goldfinger et al., 1997). The upper slope and shelf are dotted with numerous diaper structures, also indicative of overpressures within the upper plate.

Figure 5-2. Detail of the deformation front in line So105. This line shows the existence of a strain gradient between the surface and the base of the thrust sheet at the deformation front, together with heterogeneous deformation with brittle faults forming at the surface. From Booth-Rea et al., 2008.
Figure 5-3. Preliminary interpretation of the COAST multi-channel seismic survey MGL1212, line 4. Red lines indicate landward vergent (antithetic) thrust faults resulting in fault propagation folds. Red arrows indicate relative sense of motion on the fault, while blue arrows indicate the presence and location of growth strata, suggesting growth in excess of current sediment accumulation rates and potentially recent slip on the faults.
Several models of pore fluid pressure ratio in Cascadia and similar settings have been developed. Seno (2009) modeled differential stress profiles across the fore arcs in several subduction zones based on the force balance between the shear stress on the megathrust and the lithostatic fluid pressure. The estimated values of .093 and .090 for Washington and Vancouver Island respectively (this method is further discussed below). Wang (1994) estimated Pore fluid pressure ratios of 0.6–0.8 in the wedge off Vancouver Island using a steady state 3D analytical model.

Zhao et al., 2016 suggest an unusual alternative to those discussed above. They suggest that lateral shear, related to the transverse strike-slip faults of Goldfinger et al. (1997) may be related to the province of landward vergence. However, these faults extend into the seaward vergence dominated region in central Oregon, and the model is belied by the observations of Tobin et al. (2003) who note that the dominant landward vergence in northern Oregon is broken at only one locale directly above one of these faults, and invoke draining of the excess fluid pressure to explain this very localized vergence flip. We consider this alternative unlikely.

5.1.5 Arc-Parallel Strike-Slip Faulting
We observe two sets of arc-parallel strike slip faults on the Washington margin, mid and inner shelf probably dextral faults, and strike-slip faulting near the upper slope-lower slope break which is most likely also dextral.

5.1.5.1 Inner Shelf Strike-Slip Faulting
Though much of the Washington inner shelf is dominated by transverse structures, the northern and southern parts of the shelf have strike slip faulting sub-parallel to the margin. On the northern shelf, a major structural trend extends from approximately the latitude of Destruction Island to the Juan de Fuca Channel. The structure caries considerably along strike, and includes and emergent anticline on the shelf 17 km WNW of La Push WA. The structure, here termed the Los Frayles Fault after a nearby headland, shows evidence of strike-slip character along parts of its trend, shown in Figure 5-4 and Figure 5-5. The trend continues further to the south but weakens along strike.
Figure 5-4. Multichannel seismic reflection profile across the Los Frayles Fault on the northern Washington shelf. Western Geco line WO-54. See Figure 4-3 for line location.

Figure 5-5. Two crossings of an inner shelf strike-slip fault on the northern Washington inner shelf. Upper panel is line WO-54, lower panel is line WO-4026. Developed flower structure evident west (right) of the fault in the upper panel, and less obvious in the lower panel.
5.1.5.2 Upper Slope-Lower Slope Strike-Slip Faulting

At the upper slope/lower slope break, the splay fault and a se~ 3-5 km seaward of the main splay previously discussed also shows evidence of strike-slip separation at a several locations. In the area where the Washington wedge grows rapidly wider from the narrow Vancouver Island wedge, just south of Nitinat Canyon, these structures are well expressed and may be active. Although this area is expected to be one of orthogonal convergence, the evidence for right lateral displacement is clear. Figure 5-6 shows this area, and the 2.2 km dextral offset of

Figure 5-6. Shaded relief bathymetry showing the upper slope-lower slope break at center right, separating steeper upper slope (yellow and orange) from the low taper lower slope (green). The major splay fault separates the two (heavy red line). 3.4 km seaward, a second fault showing mostly dextral separation is observed, with topographic offsets and shutter ridges shown. Separation is ~ 2.2 km.
A topographic feature on the secondary splay fault. The main splay is linear and looks relatively fresh, but does not show clear dextral offset of the Nitinat Canyon walls. This may be due to the active canyon processes (turbidity currents) keeping any visible offset small, of the dextral motion could be partitioned mainly to the secondary fault.

5.1.6 Listric Normal Faulting and Extension of the Upper Slope and Shelf

A large number of mid-slope listric normal faults have been identified off SW Washington and northern Oregon on the upper slope and outer shelf (McNeill et al., 1997). These faults were first noted by Snavely (1987) and Snavely and Wells (1991), and reported by Piper (1994) and Piper et al (1995). In this project, we have mapped these structures in additional detail and spatial accuracy, and also a number of new structures were mapped off northern Oregon, south of the previously mapped extent of such features (Figure 5-7).

Figure 5-7. Typical seaward dipping listric normal fault on the northern Washington outer shelf, shelf break at left. Profile is line WO-4030.

The original analysis of multichannel seismic reflection profiles revealed that the pervasive normal faulting on the northern Oregon and Washington continental shelf and upper slope are driving E-W extension in this region. Such structures are relatively rare in accretionary prisms, but are striking in their number and scale off Washington. These structures are active, and were mapped with seismic reflection as well as bathymetry, backscatter and sidescan sonar where available. The several of these faults have been visited using the DELTA submersible by the author in 1995 (McNeill et al., 1997). Observed scarps were single or multiple event scarps offsetting thin olive green Holocene cover sediments, clearly distinguished from the underlying medium gray Pleistocene clays. The steady addition of growth strata, with some age control, suggests these structures have been slipping at a
relatively steady rate since Miocene time, thus they are not a short term response to wedge taper adjustments due to the influx of Pleistocene sediment to the trench. Instead, these faults appear to be a relatively steady state mechanism of taper accommodation, or perhaps free slip downhill of a sedimentary section floored with unusually low basal friction at the boundary with the underlying Hoh mélange (McNeill et al., 2000).

Most listric faults sole out into a subhorizontal decollement coincident with the upper contact of an Eocene to middle Miocene melange and broken formation (MBF), known as the Hoh rock assemblage (Orange et al., 1993). The Hoh assemblage is an older accretionary complex that is relatively lightly deformed and lacks a thermal of chemical overprint of deep burial or strong deformation. The Hoh assemblage is generally describes as large blocks of sild/sandstone units divided by north trending thrust zones of with mélange units, which are offset by NE trending strike-slip faults (Rau, 1973, 1979, 1979a; Orange et al., 1993). It is not completely clear whether the Hoh rocks were underplated, or simply accreted then buried by younger units (Orange et al., 1993). The areal distribution of extensional faulting on the shelf and upper slope partially matches the subsurface distribution of the MBF for N. Oregon and SW Washington. In northern and central Washington, which is also underlain by the MBF, the normal faulting is also present but less pervasive. Evidence onshore and on the continental shelf suggests that the MBF is overpressured and mobile. For listric faults which become subhorizontal at depth, these elevated pore pressures may be sufficient to reduce effective stress and to allow downslope movement of the overlying stratigraphic section along a low-angle (0.1 deg.-2.5 deg.) detachment coincident with the upper MBF contact.

Mobilization, extension, and unconstrained westward movement of the MBF clearly drives brittle extension of the overlying sediments. No Pliocene or Quaternary extensional faults have been identified off the central Oregon margin where the shelf is underlain by the rigid basaltic basement of the Siletzia terrane. The thickness and strength of the older MBF are unknown, and therefore the depth to which extension extends is unclear: a deeper compressional regime may underlie the extending MBF. When considered together with the prominent features off Washington and the large scale slope failures in southern Oregon, these structures suggest that high pore fluid pressures may be ubiquitous in Cascadia's submarine forearc. These listric normal faults may be related to overpressuring of the section with elevated pore fluid pressures (Mourges and Cobbold, 2003), in agreement with the numerous diapirs mapped in approximately the same region (Rau and Greenock, 1974; Snively and Wagner, 1982; Palmer and Lingley, 1989; this study).

The presence of E-W trending folds on the inner continental shelf suggests that N-S compression and E-W extension are operating simultaneously. An extreme case of decoupling extending to the plate interface (10-15 km beneath the shelf) would have significant implications for the extent of coupling on the subduction zone and hence position and width of the interplate locked zone. Quaternary extension of the shelf and upper slope is contemporaneous with active accretion and thrust faulting on the lower slope, suggesting either that 1) the shelf and upper slope are decoupled from subduction-related compression; 2) that there is a heterogeneous pattern of interplate coupling, with an area of low basal shear stress on the N. Oregon-SW Washington slope, or 3) that the extension takes place during co-seismic extension or overshoot events during great earthquakes.
5.1.7 Diapirism
Numerous diapirs and breached anticlines have been mapped on the Washington shelf and onshore (Rau and Greenock, 1974; Snavely and Wagner, 1982; Wagner and Batatian, 1985; Wagner et al., 1986; Palmer and Lingley, 1989; Goldfinger et al. 1997, this study). These structures range from piercement structures rising somewhat randomly from basinal areas to breached anticlines which are common. These structures bring up lithologies similar to the Hoh assemblage, and are assumed to originate from sources in the Hoh assemblage at unknown depths (Figure 5-8). Where described onshore, the lithologies consisting of phacoidally shaped mobilized blocks of massive siltstone, sandstone, and volcanic rocks in a silt/clay matrix (Wagner et al., 1986). The preliminary map shown in subsequent figures almost certainly underrepresents the abundance of diapirs on the Washington shelf, as many cannot be distinguished from steep eroded anticlines. The pervasive presence of diapirs on the Washington shelf is unusual in that the overburden there is relatively thin compared to central Oregon for example. It appears likely that not only the presence of the underlying Hoh rocks,

![Figure 5-8. Typical diapir developed in a Miocene sedimentary basin. MBF contact at base of coherent Miocene section is clearly visible and typical of the Washington shelf. Profile from line WO-4044.](image)

but likely high overpressures are required to generate the pervasive diapirism observed, and perhaps are reduced in the better drained seaward vergent fold thrust belt of Oregon, N. California and Vancouver island.

The map pattern and structural associations of the diapirs may offer clues to the overall role they play in Washington structure. We note that some of the diapirs are directly associated with normal faulting. Figure 5-9 shows this association. In this instance, it appears that the normal fault is in part developing in response to the adjacent uplift and extension required to accommodate the space required by the diapiric intrusion.
Figure 5-9. Diapir developed in a Miocene sedimentary basin in association with an adjacent normal fault. The normal fault may in part be due to extension generated at the margins of the diapir. MBF contact at base of coherent Miocene section is clearly visible and typical of the Washington shelf. Profile from line WO-4034.

In map view, the northern Washington outer shelf diapirs are clearly concentrated along a structural trend, while they appear more random off SW Washington, although roughly half of the SW Washington diapirs have associations with the more pervasive normal faults there. The association of the diapirs and normal faults is reminiscent of that found in other regions with extensional tectonics and mobile subsurface units (e.g. Morley and Guerin, Lynch and Keller, 1998).

5.1.8 Arc-Parallel to Arc-Normal Structural Transition

One of the primary goals of this work was to improve the structural mapping of the Cascadia margin to a level that an assessment of an apparent structural shift from arc parallel to arc normal structures could be evaluated. In Washington, structures with NE and E-W trends at ~ right angles to the margin parallel fold-thrust belt exist and were first reported offshore by Wagner et al. (1985), and reiterated to a limited extent in Goldfinger et al (1997) McCrory (2002) remapped these structures and produced maps generally similar to Wagner and Batatian (1985) and Wagner et al. (1986), and superseding the Goldfinger et al. (1997) mapping on the inner Washington shelf where data was spatially more limited.

The present work uses all the seismic and sidescan data previously available, and adds to that industry digital multichannel data from Western Geco first employed by Goldfinger et al. (1997). Most of these data are now available as digital segy data at https://walrus.wr.usgs.gov/NAMSS/, though some are not and available as analog sections at Oregon State University. Integrating the single channel, multichannel, and sidescan data along with our inshore bathymetric surveys has made the multiple previous interpretations resolvable to a large degree, though some areas of uncertainty remain.

The primary notable feature of the Washington inner shelf is the pervasive presence of faults and folds transverse to the coast, with strikes ranging from NW, to E-W, to SW. These
include folds, thrust faults, and strike-slip faults. Most structures reach the surface, and are eroded, with some of the upper portions removed and forming rocky exposures on the shelf. These structures were first mapped by Wagner and Batatian (1985) and Wagner et al. (1986), and later using additional data by McCrory (2002). Our new mapping uses the previous datasets and additional new data collected after 1998, and digital multichannel industry data not available to previous workers. Our mapping is broadly similar, though the spatial depiction of the Washington shelf structures is expected to significantly improve with the new data compilation, and notably the new bathymetry available to definitively pin structural orientations in several key areas.

One structure in particular is of significance, the North Nitinat Fault (NNF). This structure was extended inshore to the Washington shelf by Goldfinger et al. (1997), and was later imaged with multibeam sonar in 2010 (Figure 5-11). This feature first appeared in the mapping of Wagner and Batatian (1985). This structure is a very large south verging thrust anticline crossing the entire shelf (Figure 5-1). Multibeam mapping shows a complex sheared anticline with clear sinistral separation along a fault exposed in the forelimb (Figure 5-11). The mapping of the inshore portion of this structure shows that it gently bends to a NE strike and continues inshore to the eastern limit of the data. McCrory (2002) named this fault the Grays Harbor Fault Zone, though it had previously been named the North Nitinat Fault in Goldfinger et al. (1997). On the mid to outer shelf, the NNF bends gently to a WNW strike and extends to the outer shelf along the northern rim of Willapa Canyon. While definitive linkage to the NNF on the slope is lacking, the prominent bending of folds in alignment with this structure is inferred to support this connection as originally proposed in Goldfinger et al. (1997). If correct, the NNF is the most significant of the transverse faults extending from the abyssal plain across the broad prism and potentially extending onshore as shown in McCrory (2002). Of the transverse structures on the Washington shelf, the NNF is the only one that does not terminate or blend conformably into the N-S structural province of the mid-outer shelf.

While the NNF crosses the entire shelf, the inner part of this structure otherwise resembles many of the other folds and faults that populate the inshore structural province. The extent and individual structural details of this province have varied somewhat with the three generations of mapping that have observed it. The present map shows that margin parallel structures dominate south of the Long Beach peninsula and gradually shift to a NW strike off the mouth of Willapa Bay. As these structures approach the prominent NNF they become subparallel to it, suggesting dominance of this structure. North of the NNF, structures are also sub-parallel to the NNF and on the outer shelf, generally return to a N-S orientation, gradually becoming NNE to 47.5N, while inner shelf structures remain ~ E-W (Figure 5-1). The mid shelf boundary between transverse and margin parallel structures trends N, intersecting the coast approximately at 47.8N near the Hoh River. Many of the structures in the inner shelf region of transverse structures appear to fade out and give way to the N-S structures to the west. In some cases, structures change strike to N-s trends, for example several folds south of the NNF. Although the spatial coverage of seismic profiles is quite high, it isn’t always clear whether the E-W trends terminate, or blend into more northerly trends across the mid-shelf boundary between the two provinces.
5.1.8.1 Thinning, Extension and Subsidence of the Mid Washington Shelf.

The pervasive listric normal faulting shown in the figures and discussed in McNeill et al. (1997) implies that at least in post-Miocene time, significant thinning of the upper North American crust must have occurred. The high sediment supply to the Washington shelf has maintained a flat erosional surface during low-stands, becoming a depositional surface in high-stand times (e.g. Nittrouer, 1978; Sternberg, 1986). While no quantitative estimates of this thinning have been done, the long-term effects of this process are visible in N-S seismic sections. Figure 5-10 shows that the MBF surface is relatively shallow on the northern and southern Washington shelf. On the central shelf, this surface deepens significantly, coincident with the locus of most extensive listric normal faulting. In a subsequent section, we show a similar relationship for the Newport syncline off Oregon.

![Figure 5-10](image)

**Figure 5-10.** N-S seismic section on the mid shelf of Washington. The base of the Miocene section/top of Hoh MBF is clearly visible, with the coherent Miocene and younger section overlying the opaque MBF. This interface is relatively shallow to the north, and deepens southward into the area of most prominent listric normal faulting at center. This surface then shallows southward onto Nehalem Bank of the southernmost Washington-Northern Oregon shelf. Line WO-01.

5.1.9 Transverse Basement Involved Strike-Slip Faulting

Using sidescan sonar, seismic reflection profiles, and swath bathymetric data Goldfinger et al. (1992; 1996a, 1996b; 1997) mapped a set of WNW-trending left-lateral strike-slip faults that deform the Oregon and Washington submarine forearc. Evidence for left-lateral separation includes offset of accretionary wedge folds, channels, and other surficial features; sigmoidal left bending of accretionary wedge folds, and offset of abyssal plain sedimentary units. Five of these faults cross the plate boundary, extending 5-21 km into the Juan de Fuca plate. Using offset of subsurface piercing points, and offset of approximately dated submarine channels, they calculate slip rates for these five faults of 5.5 to 8.5 mm/yr. Little or no offset of these faults by the basal thrust of the accretionary wedge is observed. Holocene offset of submarine channels and unconsolidated sediments was observed in sidescan records and directly by submersibles.
The new compilation of seismic and bathymetric data have added new observations to the original work (Figure 5-1). The inferred locations of the North and South Nitinat faults on the slope and shelf were examined for evidence of surface deformation. On the lower slope, few lineaments were found, though some were evident and added to the map. On the lower to middle upper slope, the new bathymetry allowed well-controlled correlation of Western-Geco multichannel profiles. In this area, it was found that previous correlations were incorrect, and the re-mapped structural trends revealed a strong strike change along the North Nitinat Fault consistent with rotation and deformation of a number of thrust ridges along the N. Nitinat Fault projection. Along the same trend, a large and well mapped anticline and fault system traverses the entire width of the shelf E-W, then oddly bends NE and extends onshore. The inshore part of this system was first mapped by McCrory (2002), and was corroborated in 2010-2011 with multibeam mapping done by our lab. This connection was originally inferred by Goldfinger et al. (1997), and now is mapped in considerable detail.

Several new strike strike-slip faults were also mapped the lower slope with similar trends to the North and South Nitinat Faults, but have not been shown to have lower plate origins, or extension to the upper slope and shelf.

As inferred by Goldfinger et al. (1997), the transverse strike-slip faults are most likely driven by dextral shearing of the subducting slab and propagate upward through the overlying accretionary wedge. Tangential hydrodynamic drag caused by oblique insertion of the slab into the mantle is a possible driving mechanism. Four sinistral faults observed in only the upper plate may be remanent traces of previous basement-driven deformation. Alternatively, a similar, though unrelated dextral shear couple driven by interplate coupling may drive these faults, and may augment deformation of the upper plate for all the sinistral faults.

The preferred model of Goldfinger et al. (1997) is a model of overall right-lateral simple shear of the submarine forearc is consistent with the observed surface faults, which may be R' or antithetic shears to the overall right-shear couple. The major strike-slip faults define elongate blocks that, because of their orientation and sinistral slip direction, must rotate clockwise. They inferred that the deformation of the submarine forearc (defined to include the lower plate) is highly strain-partitioned into arc-normal shortening, and arc-parallel strike-slip and translation. The high slip rates of the strike-slip faults, coupled with the lack of offset of these faults as they cross the plate boundary, imply that the seaward accretionary wedge is not moving at the expected convergence rate relative to the subducting plate. They conclude that the accretionary wedge is rotating and translating northward, driven by the tangential component of Juan de Fuca - North American plate convergence.

Another relevant conclusion drawn by Goldfinger et al. (1996b, 1997) is that the lower plate faults tend to deform the upper plate primarily where the plates are strongly coupled enough to transmit shear stress across the decollement and into the upper plate. The pattern observed in the landward vergent province of Oregon is that the faults are well expressed seaward of the deformation front, they offset and deform the seaward limb of the frontal thrust, then then are only weakly observed across much of the lower slope. Landward of the lower slope-upper slope break, the faults are strongly expressed as surface linear scarps, fault parallel anticlines, and offset of margin parallel accretionary wedge folds. Outside of the landward vergent province, the faults are well expressed across the full wedge.
The poor expression of the transverse faults across the lower slope in the landward/mixed vergence province was interpreted as resulting from poor interplate coupling and this poor transmission of lower plate shear into the upper plate. The new observations of strong folding and deformation of the upper plate in Washington along at least the North Nitinat Fault and weak expression on the lower slope are consistent with the early observations and coupling hypothesis based primarily on the Oregon faults.

**Figure 5-11.** New OSU-BOEM 2010 multibeam survey showing E-W striking structures on the mid Washington shelf west of Grays Harbor. Structures on the mid to inner shelf have similar trends, and actually bend NE to join NE trending structures onshore related to the Olympics (McCrary et al, 2002).

**North Nitinat Fault**

Cascadia tectonics features a number of transverse strike-slip faults that strike WNW across the accretionary prism and extend across the plate boundary into the abyssal plain. These structures are described more fully in Goldfinger et al. (1992; 1994; 1996a; 1997) and Appelgate et al. (1992). In Goldfinger et al (1997), two of these faults the North and South Nitinat Faults were described in detail. At that time the North Nitinat fault was clearly mapped on the abyssal plain and in the frontal wedge. It was tentatively extended further inboard and onto the SW Washington shelf based on limited seismic reflection data and sidescan sonar (Goldfinger et al. 1997). The fault is a strike slip fault with visible left lateral separation (Figure 5-11). In this project. We re-examined evidence for a connection between this structure and the North Nitinat Fault further to the NW. With the improved multibeam grid and digital seismic data now available, this connection has scattered lineaments and structural
truncations, but as before, is not robustly mapped across the wedge (Figure 5-1). Nevertheless, we suggest that a connection is likely, based on the similarities with similar structures in Oregon.

For example, the Daisy Bank and Wecoma faults similarly are well expressed on the abyssal plain and frontal wedge, poorly expressed on the continental slope, and well expressed at the rear of the wedge. This pattern was potentially explained in Goldfinger et al. 1997) as created by the poor interplate coupling in the landward vergent, low basal shear stress portion of the Cascadia prism, and thus poor transmission of the offset of the fundamentally basement faults into the upper plate. At the rear of the wedge, greater dewatering and strong coupling drive deformation into the upper plate again, resulting in matching upper plate structures. While we cannot definitively connect the shelf and outer wedge structures, we think this connection is highly likely. The inboard end of the North Nitinat Fault, which has a large component of south-vergent thrusting on the inner shelf, bends to the NE and extends to the inner shelf (Figure 5-11). The structure connects with the Grays Harbor Fault zone mapped by McCrory (2002). We retain the terminology North Nitinat fault here as this structure was previously named in Goldfinger et al. (1997). The Grays Harbor Fault of McCrory (2002) is explained as a part of the Olympic Mountains anticlinorium.

5.1.10 New Observations from the Washington Outer Wedge: The Missoula Floods

Utilizing new high resolution multibeam bathymetric data collected in 2011, along with chirp sub-bottom and multichannel seismic reflection (MCS) data, we recently identified remarkable erosional features on the toe of the Cascadia accretionary wedge near Willapa Canyon. These features loosely resemble slope failures of the frontal thrust, but can be distinguished by several key features: They incise the crest of the frontal thrust and encompass the landward limb; They have floors below the level of the abyssal plain, similar to plunge pool morphology; They have no evidence of landslide blocks at the base of the slope indicative of block sliding (Figure 5-12). The age of the features is constrained to be ~ 10,000-35,000 y.b.p based on post event deposition and post event slip on the frontal thrust of the Cascadia accretionary prism.

Based on morphology, dissimilarity with other submarine features, and the available age constraints, we infer that these features represent the first identification of submarine “Coulees” from the Missoula floods which occurred 13,000-19,500- y.b.p. At one of the two “Coulees” we found and mapped a scarp on the frontal thrust fault within the coulee with “60 meters of vertical offset. The short duration of the erosional event that created the coulees coupled with the presence of the fault scarp within the coulee provides a unique opportunity to evaluate the slip rate along the frontal thrust scarp. Using age constraints, fault geometry and scarp height we estimate the slip rate on this frontal thrust fault to range from 2.8 - 4.1 mm/yr. To our knowledge this is the first measurement of slip rate on the Cascadia frontal thrust. Given that this landward vergent thrust, along with the set of landward vergent thrusts to the east, must accommodate 100% of plate motion, this is a new and important constraint on the
Cascadia system. These new observations have been submitted to Geosphere as Beeson et al. (2016).

5.1.11 Slip Rate of the Cascadia Frontal Thrust: Implications for Wedge mechanics and strength

Along the inboard limb of the frontal landward-vergent thrust, the new multibeam bathymetry and MCS line each image a ~60 m-high, west-side-up, fault scarp on the west flank of the depression. The scarp can be traced to the north and south beyond the sidewalls of Feature A, confirmed by seismic profiles, and is interpreted as the landward-vergent, frontal-thrust-fault scarp of the Cascadia accretionary prism.

The ~60-m-high fault scarp in the frontal ridge at Feature A is inferred to result from slip on the frontal thrust fault in the accretionary wedge, following the erosion event that created Feature A. There is no apparent incipient thrust fault imaged in the MCS profile that could be accommodating accretionary wedge deformation farther to the west, however several normal faults are found to the west, likely due to bending moment deformation of the downgoing plate. These extensional faults are further indication of lack of significant shortening west of the frontal thrust. The scarp height and estimated fault dip (38°) from the MCS section suggests about 76 m of fault slip, a relatively small

Figure 5-12. Geologically interpreted map of area of interest with fold and thrust belts mapped along with slope failures and scour features. The inset is a zoomed image at the northern scour feature which allows visualization of a ~60 meter fault scarp offsetting the floor of the scour feature.
amount compared to the ~300 to 400 m of Holocene shortening expected along the subduction zone based on plate convergence rates (DeMets et al., 1990; DeMets et al., 2010). This relatively small estimate for post-Feature A creation fault slip suggests either a very young age for the fault itself, a young age of the offsets seen in the Feature A and B thalwegs, and therefore a very young age for the features, or a relatively slow slip rate, less than plate convergence, at the frontal thrust.

If the erosional features can be attributed to the Missoula Floods, a potential timing constraint is available to estimate the slip rate on the Cascadia frontal thrust at Feature A. Using the bathymetry and depth migrated MCS profile we estimate that the frontal thrust dips ~38° to the west at the frontal anticline within Feature A. The clear fault scarp within Feature A can be traced across the floor of feature A, as well as north and south of the erosional zone with similar heights, linking it to the undisturbed frontal thrust nearby. Because the scarp heights are the same inside and outside Feature A, we infer that the frontal thrust fault offset observed in Feature A has taken place subsequent to creation of Feature A in this ~2-km wide zone. Using the Missoula Flood age range of 19,500-13,000 y. b.p for this erosion yields a slip rate at this location on the frontal thrust fault of ~4 – 6 mm/yr., a surprisingly low rate.

In a landward vergent hanging wall block that soles out near the base of the sedimentary section (Adam et al., 2005), all of the plate convergence should be taken up in accretionary wedge deformation. Plate convergence rates are estimated to be ~40 mm/yr. (DeMets et al., 2010), yet we can only estimate 4-6 mm/yr. at the frontal thrust fault. This suggests that the slip along the frontal thrust fault is accommodating ~13% of that shortening and that other prism faults and folding should account for the remainder. Alternatively, the rate over the last 10’s of thousands of years could be slower than suggested by global plate circuits such as DeMets et al (2010) which generally cannot resolve intervals less than 0.5 MY very well (Behr et al., 2010; Titus et al., 2011). Interplate coupling models based on GPS however, are consistent with full plate locking at the expected rate (McCaffrey et al., 2013). Additionally, the basal detachment could be advancing seaward and decoupling the sedimentary section at a basal detachment, which would be difficult to observe. In this case, the frontal thrust would not see the full plate rate as some of the abyssal plain section would be transferred to the upper plate with limited structural indication in the available seismic sections. This bulk shortening was suggested by Adam et al. (2004), though without specific evidence. Normal faulting of the upper plate section however suggest extension to the west, though we cannot exclude that these faults could be older, as those observed do not reach the surface. In this case, it might be possible that the normal faults were later undercut by an unobserved basal detachment.

We prefer a scenario in which the deformation of the upper plate is distributed broadly with in the accretionary wedge and accomplished though incremental motion on the numerous thrust faults and fault bend anticlines that essentially are part of the footwall block of the frontal thrust. Although seismic profiles indicate that few if any of these thrusts are outpacing the rate of basin filling because they are blind (see Adam et al, 2004), we regard this as the simplest explanation for a deficiency of slip-rate observed at the frontal thrust. The least likely hypothesis is that the age constraint of the Missoula floods is incorrect. In this case, if we apply the full plate rate to the frontal thrust, and allow no shortening internal to the wedge (extremely unlikely), the offset would be ~ 1900 years old, a scenario for which there is no evidence, and which clearly conflicts with the demonstrated crosscutting Holocene JDF system.
5.1.11.1 Implications for the Age, Velocities and Seismic Competence of the Washington Wedge

The slow slip rate at the frontal thrust allows re-consideration of the age of the Washington wedge and its growth history. In several previous works, the outbuilding rate of the landward vergent wedge was approximated as nearly the full convergence rate. This was based on the assumption that the frontal thrust in a landward vergent system would see the full plate rate. This calculation is typified by Adam et al. (2004) who estimate a total shortening across the Washington wedge to be 12-14 km based on retrodeformed seismic sections. Using a convergence rate of 42/mm/yr., they inferred that the entire outer wedge must have been constructed in 290-330,000 years. This fast construction is, as they note, greater by a factor of 6 than the ages of lower slope anticlinal ridges estimated by Barnard (1973), Carson, (1977), and Silver (1972). If we instead use the frontal thrust horizontal slip rate of 4-6 mm/yr. as the maximum rate of wedge growth relative the undeformed abyssal plain, and assume minimal shortening landward of the rear of the lower wedge, then the 12 km of shortening of the outer wedge determined by Adam et al. (2004) would yield an age of the ~ 30 km wide wedge of ~ 2-3 MY. This estimate is much more compatible with the ages of the frontal ridges, and with the expected sharp influx of sediment to the trench at the initiation of the Pleistocene at 1.8 Ma. The remaining slip required to match the likely initiation of the outer wedge at the earliest Pleistocene may be accommodated as 1) slip on a basal detachment as shown in Adam et al. (2004); 2) as ongoing shortening in the inner wedge; 3) as undetected bulk shortening west of the deformation front; or 4) could simply be recovered as elastic strain during earthquakes without anelastic deformation. Likely some combination of these factors is involved. Our preferred model is a Pleistocene wedge as opposed to one constructed in the last 200-300,000 years. This is also more compatible with the relatively high velocities shown in the velocity models of (Fleuh et al., 1998).

5.1.11 Evidence of Co-Seismic Ground Motion in the Washington and Oregon Cascadia Wedge

During the period of this project, a parallel NSF supported project in support of the Cascadia Initiative OBS deployment was completed. That project assembled a new bathymetric grid, and conducted several cruises, mostly off Washington, to acquire new bathymetric coverage in Washington and fill other gaps on the margin. The substantial new coverage sparked the discovery of the previously described submarine coulees, and has led to further new observations relevant to this project. One layer of information in the new Cascadia structure map in progress under this project, is the mapping of submarine slides. This was just a layer done for the sake of completeness, as it seemed to have no direct relationship to the project. However, an interesting observation came from the creation of this layer. After mapping 1142 slides, we found that submarine landsliding observed in the Cascadia accretionary prism is heavily weighted towards azimuths that are directed in the seaward direction, even on fold limbs that are dominantly steeper on their landward limbs. Within the accretionary prism, a region of landward vergent folds in northern Oregon and Washington is associated with minimal surface taper of the wedge. Within this region, relatively simple anticlines commonly
have steeper forelimbs than backlimbs, averaging 2-3 degrees steeper, and these limbs face landward. Most of the forelimbs and backlimbs in the landward vergent region are formed by intact turbidite sections uplifted and folded as fault bend folds. Within this region, 220 mapped slides on fold limbs (canyon related slides are excluded) have azimuths that are 72% westerly (seaward), and 28% have easterly azimuths (landward). Of these, the limb angles are steepest on the east side for 168 of them or 76.4%, and steepest on the west side for 52 of them or 23.6%. (Figure 5-13). The unusual dominance of sliding on the lower of the two slopes is unlikely to be related to fluid flow as many anticlines are breached at their crests, and likely have similar drainage. Both limbs are composed of the same abyssal plain turbidite sequences, and thus composition, stratigraphy, and mineralogies should be much the same.

We suggest that directionality of slip in great subduction earthquakes is the best explanation for the heavily preferred seaward azimuths. A similar observation of terrestrial landslides in Japan related to the 2011 Tokachi Oki earthquake has been made (Kinoshita et al., submitted, Ugai et al., 2012). We think it may be possible to estimate the PGA required destabilize the slides in the Cascadia outer wedge. Large listric normal faults in adjacent upper slope may also be driven by accelerations and perhaps “overshoot” in the largest Cascadia earthquakes, as occurred in the Tohoku 2011 Mw 9.0 earthquake. The asymmetry of the slide azimuths strongly imply that the outer wedge participates in significant co-seismic displacement, something not previously known, and questioned in some previous work (Priest et al., 2009; Goldfinger et al., 1997 and elsewhere). The asymmetry is suggestive of failure in large events, potentially overshoot events such as the overshoot observed in the 2011 Tohoku event (e.g. Yagi et al., 2015).

Figure 5-13. Rose Diagram showing Landslide azimuths binned in 10 degree sectors, Washington margin. Strong preference for WSW azimuths is apparent.
This result seemingly contradicts the assertion in these earlier works that the outer wedge of Washington and Oregon is decoupled, based on predicted high pore fluid pressures, the low wedge taper, and presence of mud volcanoes. However, decoupling and co-seismic accelerations are not mutually exclusive, as a decoupled wedge must eventually move, either co-seismically or post-seismically. This evidence suggests the latter in at least some cases, implying that the outer wedge at least some of the time participates in co-seismic slip and therefore tsunami generation, whether or not it was “coupled”, and whether or not seismic energy is radiated from this region.

5.2 Central Cascadia Structure
In this section we discuss the structure of central Cascadia along the Oregon part of the margin. The northern part of this region has many similarities to the Washington margin, and we first discuss these features categorize as we did for Washington.

5.2.3 Two Phase Accretionary Wedge
Off northern Oregon, from 44.596 N northward, the lower slope is of mixed vergence as along the Washington margin (MacKay 1995, Goldfinger et al., 1992, 1997). The landward and mixed vergent province tapers southward from 50 km, to ~ 10 km at its southern limit. The southern limit of this province coincides approximately with the southern edge of the constructional Astoria Fan, just as the northern limit is similarly near the northern limit of the Nitinat fan. The spatial correlation of the landward and mixed vergence province with the major submarine fans currently accreting to the margin support the hypothesis of Seely, 1977 that the mixed
vergence is a function of fluid overpressures developed within these thick (2-4 km) rapidly...

Figure 5-14. Structural geologic map of the central Cascadia margin.
Figure 5-15. Structural geologic map of the central-southern Cascadia margin.
deposited Pleistocene fans. The southern limit is localized along the Daisy Bank Fault, one of the basement involved transverse strike-slip faults. The vergence flips across this fault, presumably because the draining of overpressured fluids reduces the pore fluid pressure below some threshold level, triggering the vergence change (Tobin et al. 1993; Goldfinger et al., 1992; 1996b; 1997).

Other features of the two phase wedge off Oregon are similar to those off Washington. The upper slope-lower slope break is bounded by the same splay fault, and also forms the boundary between the upper slope and lower slope structural trend provinces. This structural grain trends NW-NNW, at a 10-30 degree angle to the lower slope fold thrust belt, and extends nearly the full length of Oregon, encompassing the upper slope and shelf. Exceptions to this pattern include the three major submarine banks, Nehalem, Heceta and Coquille banks, which form three major structural uplifts (Kulm and Fowler, 1974; Kulm et al., 1973, Goldfinger et al. 1994; 1997). In these areas the seaward portions are intensely folded along N-S trends. As for portions of the Washington boundary, this implies that a period of subduction frontal erosion may have separated the accretion older complex and the Pleistocene lower slope complex.

5.2.4 Landward and Mixed Vergence
The landward and mixed vergence province off Oregon is similar to that off Washington, with the same general characteristics of very low surface taper angle, wide spacing of folds, and long linear primary folds. In several locales, the landward vergent fold thrust belt is cut by upper plate tear faults with SW trends. These appear not to be basement involved structures, and their origin is unclear, but may be related to the overall strike change of the subduction zone from N-S to NWW that occurs with the regional concave=seaward bend in the subduction zone. The landward/mixed vergence province gradually narrows from Astoria Canyon to its termination at 44.8.4 N at the Daisy Bank Fault (Goldfinger et al., 1996b, 1997). This taper mirrors the taper in thickness of the constructional Astoria Fan which is built atop the trench fill that completely fills the trench topography. The termination of the landward/mixed vergence region is abrupt, and occurs at the Daisy Bank (DBF) transverse fault. The flip from landward to seaward vergence occurs with the DBF acting as a tear fault in the frontal thrust. The draining of pore fluid pressure at this deep seated fault is presumed to localize the vergence flip, as it does at the nearby Wecoma Fault (Tobin et al., 1993; Goldfinger et al., 1996b).
Figure 5-16. A. Shaded relief bathymetry of the Cascadia accretionary complex offshore northern Oregon. Barbed black lines delineate mapped traces of the splay fault scarps (barbs point down dip) and an abrupt break in slope (Goldfinger, 1994). (inset) Bathymetric profile (A–A’) shows the abrupt break in slope separating younger accretionary wedge on the west from the older accretionary complex on the east. The younger accretionary wedge is dominated by (a) low slope; (b) landward-vergent structures; and (c) widely-spaced margin parallel folds. The older accretionary complex features (a) fold trends oriented normal to the convergence direction; (b) a steeper slope; and (c) seaward-vergent structures. B. A seaward-vergent splay fault separates the younger from the older accretionary wedge. The unmigrated USGS seismic reflection profile L-5-W077-12 (profile B–B’; Mann and Snavely, 1984) provides evidence for splay faulting and recent displacement of the seafloor.

5.2.5 Arc-Parallel Strike-Slip Faulting

5.2.5.1 Inner Shelf Strike-Slip Faulting

Most of the Oregon inner shelf is dominated by slightly oblique NNW trending structures, but the northern shelf has significant strike-slip faulting. On the northern shelf, a major structural trend extends from approximately the latitude of Northern Netarts Bay to the Columbia River. This fault, first described in Goldfinger (1994), is known as the Cape Falcon Fault. The Cape Falcon fault is a complex structural zone consisting of a straight N-S trending segment 70 km in length, and a 34 km curving segment bringing the structure ashore at the north end of Netarts Bay. It is unclear how the structure terminates or transfers slip to other structures at its northern end. This fault parallels the coast from the Columbia River southward to ~ Tillamook Bay, where it curves to the southeast and comes ashore at Cape Meares near the north end of Netarts Bay. As this fault changes strike to the southeast, it becomes nearly pure thrust, consistent with a presumed dextral strike slip structure. Along its generally northward trend on the shelf, the new observations show that it bends gently NE, and as it does so, shows evidence of extension. It appears that a broad releasing bend may characterize this section of the fault, as it broadens to several traces, and becomes involved with significant normal faulting on the middle to outer Oregon shelf and upper slope.
The exact relationships between the strike-slip and normal faults are not completely clear as of this writing, but the possibility exists that the strike slip system is more pervasive and significant than previously thought, and may influence or control the region of normal fault extensional structures known from previous work (McNeill et al., 1997). If this model is correct, the Cape Falcon structure make take on a more significant role as a crustal block boundary. McCaffrey et al. (2007) published a model of crustal block domains based on GPS data. Within the broad clockwise rotations first published in McCaffrey et al (2000), they defined several domains in the forearc that may accommodate crustal strain. One of these boundaries corresponds to the North Nitinat Fault (Goldfinger et al., 1997), recently remapped in this project. While the fault is there, the original interpretation, maintained here, is that this fault is fundamentally a lower plate structure, manifested in places in the overlying upper plate. Its role as a larger crustal block in the upper plate forearc remains unclear so far. The Cape Falcon Fault on the other hand, may be an important block boundary that has been unappreciated to date. It may be related to faults mapped onshore, in particular the Tillamook Bay structure(s). This significant structure, though not matching the offshore Cape Falcon structure exactly, may extend deep into the forearc into the Tualatin Valley, and have as much as 12 km of dextral separation (R. Wells, pers. comm).

Figure 5-17. Cape Falcon Fault zone on line WO-01 on the northern Oregon shelf. The complex transpressional zone has multiple vertical traces and varies considerably along strike.
5.2.5.2 Upper Slope-Lower Slope Strike-Slip Faulting

At the upper slope/lower slope break along the Oregon marine, the splay fault previously discussed does not show direct evidence of strike-slip separation. On the northern Oregon upper slope there is a strike-slip fault in the mid-upper slope region. We call this fault the Nehalem Fault, and it extends 73 km from 44.898N to at least 45.597 N. There are scattered bits of evidence for dextral faulting exposed on the north-central Oregon slope, including a prominent segment at 44.43N and another en echelon segment at 44.232N. These segments may be older faults exposed in folds, but appear to be inactive in recent times. The new mapping reveals no further evidence for margin parallel strike slip faulting on the central (Oregon) margin.

5.2.3 Transverse Basement Involved Strike-Slip Faulting

A number of changes and improvements have occurred while updating the structure map of Oregon. One significant change is that since the structure map on the Oregon part of the margin was completed in the mid 1990’s, four very large submarine landslides were discovered on the central and southern continental margin. This was published in Goldfinger et al. (2000), several years after the structure map was done. Now the structure map has been revised to reflect the newer interpretation of the lower slope.

Many of the faults observed there originally are, as it turns out, most likely compressional features related to the deceleration of the slides, rather than or in some cases in addition to accretionary prism imbricate thrusts. High-quality seismic reflection data in this area are sparse, making interpretation of the structure uncertain. However, some general observations may be useful. First, within the area of the large-scale slides exposed on the lower continental slope, the structural style is clearly distinct from elsewhere in the accretionary prism. In the northernmost slide, known as the Heceta slide, the interior area is hummocky in nature, showing little evidence of coherent fold thrust tectonics. It most resembles a slide surface that has been buried by some sediment cover, but barely affected by subsequent compressional tectonics (Goldfinger et al. 2000). What weak margin parallel ridge structure exists is subdued, and of much shorter wavelength that the fold thrust tectonics elsewhere in the accretionary prism. GLORIA sidescan sonar data reveal that beneath the sediment cover, a rougher more hummocky surface is imaged, not an expected observation for an accretionary prism, but expected for a slide surface ~ 100ka in age (Goldfinger et al. 2000). This hummocky surface extends well into the abyssal plain in the subsurface, where it is also imaged in available reflection data beneath the bathymetrically featureless abyssal plain.

As described more fully in Goldfinger et al. 2000), progressively more accretionary wedge-like morphology is apparent in the southern slides (Goldfinger et al. 2000). We interpret that the progressively older regions of slide debris have been increased to longer periods of compressive tectonics, shortening and compressing the slide packages along with pre and post slide abyssal plain section at the deformation front and within the slide areas. Like the Heceta Slide, these re-deformed areas have shorter wavelength and shorter strike length folds, suggestive of deformation of a less coherent section. These details of the lower slope off southern Oregon have been known for some time, but the structural map had not been
updated to reflect these newer observations which may affect the interpretation of the updip limits to plate locking in this area (Priest et al., 2009). We have excluded the portion of the slide debris that lacks a coherent fold thrust signature from the updip transition zone, as it is most likely incapable of strain accumulation. This excluded area appears to have a low gravity signature and affinities to the JDF plate (Figure 6-25).

The nature of the primary forearc basin in this area, and in Cascadia in general is perhaps not strictly a tectonic feature as implied by the term syncline. In at least some areas, such as central-Northern Oregon shelf, this feature might better be described as a monocline (the eastern “limb”), which abuts active folding of the inner accretionary prism to the west. The true nature of this feature is not revealed by current seismic data, but this alternative interpretation appears as consistent with the data as a synclinal interpretation, and is similar to a class of forearc basins commonly observed elsewhere.

The northern limit of the region of reduced deformation is marked abruptly by the Cape Falcon Fault (Goldfinger, 1994). This complex structural zone strikes N-S, and is the eastern boundary of the highly deformed Nehalem Bank, a structural deformation and uplift zone on the northern Oregon shelf. This fault shows clear evidence of dextral slip in sidescan sonar imagery, and has many of the classic indicators of such in seismic profiles, with reversals of vertical sense of motion over short distances, vertical fault traces, flower structures, and evidence of extensional and compressional features related to small strike changes. At its southern end, the Cape Falcon Fault gently turns landward and comes ashore near Cape Meares. The structure becomes strongly compressional in this area, consistent with dextral block motion of the inner vs. middle shelf.

This major fault system becomes involved with listric normal faulting on northern Nehalem Bank, and indeed these faults may be related to strike changes in the Cape Falcon fault system (Figure 6-1). While the azimuth of block motion is unclear due to gently curving traces, the northern end of this system is likely a releasing bend due to the eastward bend of the fault zone. Just outboard of the fault significant listric normal faulting is present, suggesting of a kinematic relationship, though the listric faults are NW of the CFF termination, not to the northeast as expected for a simple termination of a dextral structure. North of 46.21N, the Cape Falcon structure appears to weaken, and north of 46.33 is not observed. The northern termination of what appears to be a major structure is unclear. Similarly, the Tillamook structure becomes less robust northward, and is not observed north of 46.33N.

Several other structures inboard of the Cape Falcon Fault system appear to mirror the general style and trends of the CFF. A similar structure, here named the Tillamook Fault Zone, also strikes north-south on the northern Oregon shelf, and turns landward intersecting the coast at Tillamook head, where in may be expressed in both the topography and existing mapped structures (Figure 5-1). A third structural zone with similar trends comes ashore at the north end of Netarts Bay, here named the Netarts Bay Fault, and may influence the topography bounding the Bay, and the location of the bay itself.
5.2.1 Additional Central Cascadia Observations

Another significant improvement to the map has been the mapping of inshore structures on the inner shelf near Newport Oregon. This fortuitous mapping project arose because of the need for detailed mapping of a wave energy test site on the inner shelf. We conducted multibeam mapping, sub bottom CHIRP and boomer profiling and some sampling in May-June 2014. The results are given in Goldfinger et al. (2014) and briefly described here.

Figure 5-18 off Newport Oregon highlights a number of topographic features on the inner shelf. Of interest are the NW trending features with WNW to NW axes. These features were not known prior to 2010 when they were mapped for the Oregon State Waters Mapping Program. That project did not include subsurface investigations, so the origin of the features was not clear. One hypothesis was that they were structural in nature, another that they were sediment transport features. In the wave energy survey project, we investigated the pervasive NW trending topographic features to determine their origin and relationships to hard substrate. Survey area and tracklines are shown in Figure 5-19. We initially thought that the features were likely structural, as they are pervasive through the area, extend well to the north along the Oregon inner shelf (limits are presently unknown). These features also lie at nearly right angles to the plate convergence direction, suggesting a potential link. However, they also lie at 50-70 degree angle to the major active structures previously mapped, making a structural origin less than straightforward (Figure 5-20).

Figure 5-18. Color shaded-relief multibeam bathymetry data collected at Newport, OR. Sediment sample stations plotted over the Reson 8101 (240 kHz) multibeam bathymetry data.
In Figure 5-21, we show the NW trending features imaged with backscatter data. The low amplitude topographic highs correspond to the backscatter highs, which generally are strongest on the SW flanks of these features. The topographic highs are asymmetric, with steeper slopes to the southwest, and shallower slopes the NE. The backscatter data also follow this pattern, with high backscatter tracing the SW (steeper) flanks, and fading in intensity to the NE (Figure 5-22). The sub bottom profiles (Figure 5-23) show that the topographic highs generally do not correspond to faults as we had initially assumed. There as small diffractions in the data most likely related to steps in acoustic impedance (hard to soft) that extend downward, but are artifacts and not structures. We have not found any examples where these low amplitude asymmetric highs are linked to faulting. We also observe that the underlying unconformity is not deformed below these features, as one would expect if they were generated by faulting. There is however a slight mimicking of the upper topography that we interpret as velocity “pullup” and artifact of having more high velocity material overlying the topographic highs.

**Figure 5-19.** Sets area bathymetric data and geophysical tracklines near seal rock, OR.
Figure 5-20. Major structures of the inner shelf off Newport, or. Sets bathymetric data shown. Sets lies on the east flank of major syncline at center. NW trending bathymetric features in the sets bathymetry cross the structural grain at an obtuse angle.

Examination of the CHIRP and boomer datasets reveals that the relationship between the topographic highs and the backscatter is very strong, and the lack of underlying structure is pervasive. The low amplitude topographic highs have wavelengths of 200-600 m, and are overlain by sub-parallel but smaller features that are generally similar, which we interpret as sand waves.
Figure 5-21. Backscatter data in the sets area overlain with boomer and chirp sub bottom profile tracklines. Pervasive NW trending features are shown, with high backscatter concentrated on the SW faces of the asymmetric features.
Figure 5-22. Backscatter (top) and shaded relief imagery of a part of the sets are. NW trending asymmetric bathymetric features are shown in the seabed imagery. Steeper faces toward the sw. Scour depressions (high backscatter) form drainages, mostly on the steeper faces. Features are tentatively interpreted as subaerially formed paleo-dunes.
Looking closely at the backscatter and bathymetric data we see that the high backscatter areas are depressions as previously described, but also that they are closely related to the underlying low amplitude highs. Many of the high backscatter depressions form along the steeper SW flanks of the highs, and many also form dendritic drainages off these highs into the swales between them (Figure 5-22). The presence of drainage patterns like this suggest that the underlying features are long-lived enough that secondary erosion processes are modifying them significantly. This also provides an explanation for the lack of temporal movement of these high backscatter features. At least within the SETS study area, and perhaps elsewhere, the high backscatter depressions are linked directly to the underlying substrate. Therefore at least in this case, these features cannot be equated with the “ripple scour depressions” as discussed by Cachionne et al. (1984) but have a different association with underlying topography. It is possible however that erosion of these features may be enhanced in this area by scouring around the topographic highs, and or sediment transport controlled or modified by these features in such a way to exaggerate the bathymetric expression of the underlying features. In this case, the scour is due to increase wave energy stress related to the topography. In this way there may be a looser genetic association with the ripple scour depressions described elsewhere. In any case, this apparent linkage between surface and subsurface features suggests somewhat less temporal and therefore volumetric mobility of the surficial sand sheet that would be expected if the scours were randomly generated during individual storm events, at least in the water depths in the study area, 45-75 m. The larger asymmetric highs are generally similar

Figure 5-23. Chirp profile and corresponding backscatter across possible paleodunes, sets area. Tie line between backscatter and profile are shown. Lower panel shows profile optimized to show deeper possible unconformity surface (this potentially also could be an artifact).
in form to sand waves, and similar in orientation to the overlying smaller sand waves, yet their wavelengths (200-600 m) are very large for this type of environment. We know of no modern analog on the PNW shelf where such features are forming today. As these features appear inactive, and without a modern submarine analog, we suspect that they may in fact be Pleistocene subaerial features that are now partly exposed and being overlain and modified by recent marine processes. Candidate features that could fit this description are very large subaerial dunes such as those observed between Florence and Coos Bay today along the modern coast. The dunes of the southern Oregon coast are similar in scale, and at least some areas have a NW orientation (Figure 5-24, Figure 5-25), though E-W and SW trends are also seen in the same region. Dunes of that orientation have a much smaller wavelength and have a NW gentle face, consistent with summer N-NW winds. Dunes generally form with the gentle face on the upwind side. The asymmetry of the dunes observed on the southern Oregon coast (near Winchester OR) with a NW strike is with the gentle face facing SW, the predominant storm wind direction. The possible relict dunes in the SETS area have the opposite orientation, with the gentle face on the NE side, suggesting prevailing NE winds, very different from seasonal prevailing winds today.

Figure 5-24. Typical subaerial dune cross section.

Figure 5-25. Dune field south of Winchester, OR. These dunes are roughly the same scale as the interpreted paleo-dunes, having wavelengths of ~ 300-400 m (other nearby dune field have much shorter wavelengths). The wind direction indicated by these dunes is northerly, while the offshore possible paleo-dunes suggest northeasterly prevailing wind.
The CHIRP and boomer seismic profiles reveal a complex shallow subsurface topography that is for the most part not apparent in the surface topography (Figure 5-26). In addition to the pervasive NW trending “paleo-dunes” and probably modern sand waves observed at the surface, the shallow subsurface (< ~20 m) is characterized by a generally irregular and commonly rough surface defined by several prominent reflectors in the sub-bottom profiles. We used the IHS Kingdom seismic interpretation package to integrate the SEGY seismic data from the boomer and CHIRP surveys with bathymetric and backscatter data for interpretation. We interpret 2-5 (typically 3) significant subsurface reflectors traced throughout the study area that we have used to track the variability of what are most likely old erosion surfaces. A planned vibra-coring survey was not successful in ground-truthing the sub bottom data, thus
the following discussion lacks definitive ground truth regarding the lithology, hardness or age of the stratigraphic sequence.

While it was known or surmised that a transgressive gravel sheet existed in the subsurface, and it is presumed that a subaerial topography drowned after the LGM transgression existed, these data are among the first to show this surface in detail in the region. In the SETS area, we observe what are most likely former stream channels associated with the modern Beaver Creek, Alsea River, and Yaquina River Channels. While not enough data exist to definitively connect the mapped channels with these modern systems, they positions and trends are highly suggestive of such a connection. The former channels are now filled with probable transgressive and post-transgression sediments. In addition to the channels, the uppermost hard reflector, here interpreted as the transgressive surface in most cases, has significant topography that may have been the pre-transgression land surface, likely modified as it passed through the surf zone during rapid inundation of the latest Pleistocene meltwater pulses. This surface merges with the seafloor reflector in many places. In such cases we are unable to determine whether the surface reaches the seafloor or is thinly covered in some places, though in others this surface merges with the seafloor in the area of possible relict dunes. The shallow subsurface imaged in the SETS study resembles that described onshore where under high-stand conditions, Pleistocene sand filled channels and low spots cut into the Astoria Formation and Nye mudstones and are now exposed in the Newport area (Snavely et al., 1969). Without additional geophysical data we cannot know with certainty that the rough paleo surfaces observed at the SETS area are typical of the Oregon inner shelf, or in somewhat anomalous. We know of no particular reason, however, to suspect that this particular site is anomalous and think in more likely that it is relatively typical of the mid to inner continental shelf of at least the central Oregon margin.

5.3 Southern Cascadia Structure

The California margin proved relatively straightforward to remap, and the new map is largely compatible with that of Clarke (1990) in general, though with some important differences. Using the digital Western GECO reflection profiles that were unavailable to Clarke, we are able to see deeper in the section and map structures more comprehensively. The California margin also intersects the coastline, offering the opportunity to connect mapped structures onshore with those offshore. Several enigmatic features are observed that have been mentioned in previous publications: 1) the mid slope terrace is a prominent feature that distinguishes the California margin from all other areas in Cascadia. This terrace overlies an unusually opaque accretionary prism. Centered in the mid slope terrace area is a semicircular depression into which numerous small canyons and rilles have incised radially, feeding the larger Trinidad Submarine Canyon which incises the lower slope. The reflection profiles have now revealed that this depression is underlain by a low angle listric normal fault that also extends further to the north than the obvious bathymetric depression. There appear to be several strands of this shallow seaward dipping extensional system that may be responsible for the morphology of the mid-slope terrace, though there could be other causal factors as well. Though shallower than the Washington and Northern Oregon large listric normal faults, they are broadly similar, and have rollover anticlines paralleling the traces, and smaller extensional structures in the hanging
wall blocks, attesting to thinning and extension of this block (Figure 5-27, Figure 5-28). Further west (downdip) evidence of extension in the mid slope terrace gives way to compressional thrust faulting associated with the active accretionary prism.

Figure 5-27. Seismic reflection figure showing uninterpreted (top) and interpreted (bottom) east west section across the mid slope terrace in northern California. Seward dipping detachment fault and rollover anticline are apparent above the multiple. Western (left) structures are part of the active accretionary prism.
A recent cruise in September, 2014 surveyed this region with sub-bottom CHIRP and made AUV dives using the NOAA autonomous vehicle “Lucile”. The purpose of the cruise (in which we participated) was to make observations of cold water coral communities in exactly this extensional zone. Previous trawl surveys had recovered samples of these corals not only where expected, from rocky substrates, but also from the mid slip basin where no rocky substrates are known or expected. We hypothesize that the extensional faults shown may be venting methane rich fluids, and forming carbonate crusts at the seabed, similar to many other localities in convergent margins. This would provide the hard substrate required for the coral communities.

5.4 Coulomb Modeling of the Cascadia Forearc

The older Cascadia thermal models suggest that the megathrust reaches 350 °C mostly offshore (Hyndman and Wang, 1993, 1995). On the other hand, geodetic locking may extend to the coast and some distance inland (Chapman and Melbourne, 2009). Locking and coseismic slip however do not have to be the same, as shown by the 2011 Tohoku events, where locking appears to extend to 100 km beneath the arc, but coseismic release was offshore. Cozens and Spinelli (2012) attempt to reconcile the Cascadia data by adding fluid circulation in the JDF crust, cooling the system and broadening the locked zone, resulting in at least better agreement with
geodetic data. Within this context, we attempt to model the distribution of basal shear traction with current constraints.

**Figure 5-29.** Footprint of the new 100m grid built from a mosaic of DOGAMI, NOAA, OSU and other surfaces.

In order to describe accretionary wedge morphology, we constructed a new 100m bathymetric grid to represent the upper surface of the accretionary wedge (Figure 5-29). The grid comprises a stacked mosaic of Oregon Department of Geology and Mineral Industries coastal LIDAR surfaces (both land and shallow bathymetric - > 4m resolution), Active Tectonics and Seafloor Mapping Laboratory (ATSML) and NOAA's territorial sea gridded bathymetry (2-8m resolution), NOAA's tsunami inundation models (~10m resolution), ATSML's most recent Cascadia bathymetry grid (100m resolution), and NOAA's Coastal Relief DEM (~100m resolution). The mosaic was combined and resampled at 100m resolution with co-location precedence given to the newest and highest resolution surface using ArcGIS10. The new grid was subsequently combined with McCrory et al.'s 2004 model of slab depth (smoothed and resampled at 100m resolution) to generate a three-dimensional model of Cascadia's accretionary wedge, bounded by the deformation front and both the Crescent-Siletz terrain and a 450°C basal isotherm.

Using a matrix of initial values for basal pore fluid pressure (70-95% of lithostatic) and rock density (2300-2500 kg/m³), and assuming Byerlee's Law, where \( \mu_b \) (coefficient of basal friction) = 0.85, to be valid on the decollement, shear traction on the Cascadia megathrust was calculated from the deformation front east to the 450 degree basal isotherm as defined by Cozzens and Spinelli (2012; Figure 5-30).
average maximum shear stress of ~15MPa (Lamb, 2006; Molnar and England, 1990; Tichelaar and Ruff, 1993; Hyndman and Wang, 1993; Peacock, 1996). Employing pore fluid pressures above hydrostatic but well below lithostatic ($\lambda_b = 0.7 - 0.86$), our analysis resulted in frictional shear stress on the Cascadia megathrust meeting or exceeding 15MPa within 5-10 km of the deformation front, failing to satisfy the upper-slope to outer-shelf location of the down-dip limit of the locked zone and maximum inter-plate coupling as determined by current geodetic models. Basal pore fluid pressures of at least 90% of lithostatic are required over the entire seismogenic zone to generally satisfy the existing models of the down-dip limit of the locked zone. These initial tests support high pore fluid pressures in the accretionary wedge, particularly in Washington, as predicted by seismic velocity inversions and suggested by the presence of landward-vergent thrusts (Seely, 1977), listric normal faults and near surface methane horizons in recent multi-channel seismic data (Holbrook et al., 2012; MacKay, 1995; McNeill et al., 1997). Moreover, our analysis of frictional shear stress also suggests that it is unlikely that a sudden change in pore fluid pressure occurs across the basal decollement. With the coefficient of basal friction corrected for dissimilar pore fluid pressures on the decollement.

On the basis of heat flow and thermal modeling, megathrusts worldwide are predicted to have an

Figure 5-30. Basal shear traction (frictional shear stress) calculated using Byerlee’s coefficient of basal friction (0.85), sub-lithostatic pore fluid pressure (86% of lithostatic), a typical density of 2300 kg/m³, and H, the thickness of the wedge normal to the Juan de Fuca slab. The model is bounded to the east by a 450°C basal isotherm as defined by Spinelli, 2012. Values are reported in MPa. Green contours represent 5MPa increments. Of note is the relatively large region of low basal traction (as indicated by the color blue) and low traction gradient near the deformation front off northern Oregon and Washington.
and in the overlying wedge, even a 5% variance resulted in a pronounced eastward shift in basal shear stress magnitudes such that values of 15 MPa and greater occurred landward of the coast - a result inconsistent with both seismic and geological observations and current models of the locked zone. In addition, our analysis of frictional shear stress employing near lithostatic pore fluid pressures, typical density values, and a 15MPa threshold appears to be consistent with the upper-slope to outer-shelf position of the down dip limit of the locked zone inferred from the CAS3D-2 geodetic model (Wang et al., 2003), with the coupling parameter falling within 1 standard deviation of our determined mean for a 15MPa threshold (Figure 5-31). The down-dip trend of increasing frictional stress also suggests that the plate interface up-dip of maximum coupling (i.e. the lower slope and deformation front) contributes only marginally to strain accumulation and is passive inter-seismically, particularly in Washington where basal traction and traction gradient are low from the deformation front to 50km landward. This loading distribution is distinctly different from GPS based models which commonly put most of the locking far offshore (McCaffrey et al, 2000; 2013). GPS however is not at all sensitive to the offshore position of locking, leaving the actual locked zone position essentially unconstrained.

While Coulomb wedge theory allows for critical taper, and therefore stable coseismic sliding (coherent rupture), to occur under low wedge angle and high pore fluid pressure conditions such as those manifest in northern Cascadia’s outer forearc, a preliminary review of multi-channel seismic (MCS) data from Washington suggests that the outer wedge has not yet obtained critical taper and is unlikely to slide stably during coseismic rupture (Davis et al. 1983; Dahlen and Suppe, 1984). Furthermore, our initial analysis suggests that the tsunamigenic portion of the North American plate may be relatively narrow, shallow (with respect to submarine depth) and restricted to the upper slope and shelf in Washington. Large landward vergent thrusts forming the core of numerous widely spaced fault propagation folds, with backlimbs in excess of 1km in length, and are antithetic to the basal decollement, the low traction region off Washington likely represents a region of velocity strengthening and coseismic shortening rather a region of critical taper capable of sustaining coherent rupture with stable coseismic sliding on the megathrust. As such, this region likely exhibits partitioned slip and coseismic deformation that is largely horizontal in nature rather than vertical and therefore is unlikely to significantly contribute to tsunamigenesis in Cascadia. An important point that is sometimes misunderstood is the issue of slip to the trench. In recent work (Priest et al., 2009, Witter et al., 2011, 2012, ongoing work) we have completed hundreds of tsunami models run for various configurations in Cascadia. Whether slip to the trench is allowed or not, it is a secondary factor at most. The elastic storage of strain across the entire forearc, and its rapid relaxation is the primary driver in tsunami generation, whether slip to the trench occurs or not.
Figure 5-31. Mean location, +/- 1 standard deviation, for 15MPa limit of basal shear traction determined using all possible combinations of Byerlee's coefficient of basal friction, pore fluid pressures consistent with geological and geophysical observations (0.86, 0.90, 0.93) and typical density values (2300, 2400, and 2500 kg/m³).

6. Discussion
The primary hypothesis, derived from previous preliminary work, is that basal traction on the locked plate interface should be reflected in deformation of the upper plate, and that the
boundary between traction driven structures and structures responding more to the regional stress field may offer an indication of the long-term position of the downdip edge of the locked interface. This hypothesis has a number of potential problems, including 1) the large time range monitored by upper plate structures (post Miocene); 2) as compared to contemporaneous locking and modern geodetic evidence; the potential of multiple drivers of transverse structures that define the boundary; 3) a potential N-S strain gradient along the forearc that may alter the meaning of the structural boundary along strike; 4) multiple processes interacting contemporaneously; 5) variability of pore fluid pressure, incoming sediment chemistry, heat flow and other complicating factors along strike, and 6) the breadth of the transition zone may be highly variable along strike, and thus the transition may occur in poorly mapped areas onshore. In this section we examine the potential stress boundary in light of these factors, and compare the result to other measures of the position, width and position of plate locking derived by other measures.

6.1 Washington and Northern Oregon
The Washington-Northern Oregon margin represents and exceptionally unusual and complex submarine forearc, with contemporaneous frontal accretion, normal faulting, mid-crustal detachment, transverse strike-slip faulting, arc parallel compressive structures, and an arch in the downgoing slab due to the concave subduction zone. In Figure 6-1, the structure map is overlain by three dashed lines, broadly representing 1) the transition between the low taper, landward vergent younger wedge and the older accretionary complex; 2) the shelf edge; and 3) A boundary separating an outer structural domain of NW and NNW trending structures, and an inboard domain of structures trending between NW and SW, mostly transverse to the arc and coastline.

6.1.1 Updip Transition
In Figure 6-1, the updip transition is drawn at the boundary between seaward and mixed vergence and landward vergent structures. This boundary also is near the back edge of the Pleistocene wedge, and at the structural discordance and slope break between the upper and lower slope. These near coincident morphologic and structural boundaries suggest a rheologic boundary is likely at the same location. Although frontal erosion of the Pliocene and older upper slope complex, we presently cannot constrain the interval of missing time or the contrast in strength across this boundary. We presently are unable to constrain the potential for strain accumulation in the seaward domain. Thermal models suggest these regions should be locked (e.g. Hyndman and Wang, 1995), however the likelihood is that the pore fluid pressure is a high fraction of lithostatic, based on the multiple criteria of low wedge taper, mixed vergence, modeling of the overburden stress. While quantification of these criteria for evaluate locking fraction is elusive, we think it likely that this factor overrides the thermal models. There are simply no direct observations, and too many degrees of freedom to establish a reliable framework to evaluate this factor. The mapping of asymmetrical slide distribution strongly suggest that at least some of the time, co-seismic slip occurs with enough energy to trigger landsliding even on the outermost thrust limbs. At the same time, the mappable landslides in the mixed vergence region are relatively sparse considering the region is greater than 1M years
in age overall, and earthquakes occur there at a Holocene rate of 1 every ~ 500 years. There have been roughly 2000 great earthquakes in that time, yet only 237 mappable slides. This suggests that perhaps only the very largest earthquakes are capable of velocity weakening behavior extending to the trench (i.e. Gao et al. 2015). Simply comparing raw numbers, if one slide occurred per very large earthquake (a completely ad hoc assumption), then the largest earthquakes would represent ~ 11% of the total. Presently, the Holocene paleoseismic record of Goldfinger et al. (2012, 2013) includes 2 of 19 extreme events (T11 and T16), representing 11% of the paleoseismic record. This could well be coincidence, but suggests that the notion of slip to the trench in 1 of 10 events could be compatible with both the slide data and the paleoseismic data.

6.1.2 Downdip Transition
The Washington and northern Oregon margin structural transition is shown in Figure 6-1 and Figure 6-12. Reference source not found. (red dashed line). This transition broadly demarks the seaward margin parallel structural domain and an inboard domain of mostly transverse structures striking NW to SW. In Washington, these structures are on the central and SW shelf coming ashore at 47.775 in the north, and 46.595 in the south. North of this limit, contractional structures extend inboard to the coast, and several prominent anticlines deforming Pleistocene uniformities and the seafloor are evident (Figure 6-1; McCrory, 2002). Between these limits, the transition bulges offshore, to some extent paralleling the seaward bulge in upper slope structures that are being transported seaward by listric normal faulting (Figure 6-1). This seaward bulge was first mapped by McNeill et al. (1997) and is one of the more unusual features found on the Cascadia or other subduction margins. At its deeper end this bulge impinges on and likely influences the upper slope-lower slope transition, and the seaward vergent thrusts mapped there are to some extent, gravity driven. Further complications are the NNF, which crosses from the wedge onto the shelf, and clearly violates our interpreted structural domain boundary. Within the bulge, excepting the NNF and adjacent structures, folds and faults are sub-parallel the trend of the bulge. This clearly indicates that these trends are influenced by the gravity sliding facilitated on the MBF interface by listric normal faulting. It is likely that were this not the case, most of these structures would be assigned to the seaward margin parallel structural domain.

6.1.2.1 Northward migration of the Forearc Block and Related Arc-Parallel Compression
The inboard domain structures meet the domain boundary at angles varying from 10 to 90 degrees, with a higher proportion of high angle intersections. The hypothesis of an overall transition to the downdip edge of plate locking at the boundary between margin parallel and margin normal compression is complicated at each locale at which it has been applied. In Washington, forearc translation/rotation northward in to an unyielding Canadian forearc framework drives the N-S strain gradient in Washington, which exhibits numerous margin normal structures in the forearc. GPS studies indicate that northward motion
is 6-10 mm/yr. (McCaffrey et al. 2000). Block models suggest a longer term (post Columbia

Figure 6-1. Structural provinces of the northern Cascadia margin. Outboard of the dashed green line is the landward-mixed vergent province. Between the green and red lines is the generally NW-NNW trending seaward vergent province of the upper slope. The red line marks the approximate boundary between the upper slope-shelf province and the inner shelf province dominated by transverse structures. The shelf edge is in blue.
River Basalt) rate of ~ 8 mm/yr. for coastal Washington (e.g. Wells and Simpson, 2001). A rate of 2.2-3.3 mm/yr. is estimated for upper plate structures onshore and offshore in coastal Washington (McCrory, 2002). The clockwise rotation/translation of the Coast Range block northward impinges on the less mobile Canadian-Vancouver Island terrane which is not rotating, thus the relatively fast moving and weaker OCR block, with its thinner Siletz River-Crescent volcanics terrane absorbs most of the strain as E-W folding and thrusting, as well as oblique strike-slip faulting (Wells and Simpson, 2001; Miller et al. 2001; Johnson et al., 2004). The northern boundary of this impingement is between the northern boundary of the Olympic Mountains and Vancouver Island. The high uplift rates in the Olympics and unroofing of the subduction complex and Crescent volcanics attests to the long term and ongoing deformation there. Savage et al. (1991) estimate a modern rate of ~ 3.2 mm/yr. higher than any modern rate in Cascadia except near Cape Mendocino.

An open question is the role the Olympics play in N-S crustal shortening of the forearc. Tabor and Cady (1978) proposed what could now be considered the classical view of the Olympics. Figure 6-2 shows this view as a continuous accretionary prism. Long-term frontal accretion eventually overturns the rearmost thrusts, and erosion unroofs the system, exposing the core rocks and leaving a rim of Crescent Formation rocks (Figure 6-2). This view has been carried forward in numerous publications (e.g. Blakely et al. (2009); Brandon and Calderwood, 1990; Stewart and Brandon, 2004).

![Figure 6-2. Classical view of the formation of the Olympic Mts. Figure from Blakely et al. (2009), Modified from Tabor and Cady, (1978).](image)

The formation of the Olympics however is most likely more complex. The Olympics are roofed by the Crescent formation in thrust contact, indicating that the core rocks were underthrust beneath the Crescent/Siletzia complex, as presumably is occurring elsewhere in Cascadia (Parsons et al. 1999). Stewart and Brandon (2004) consider the wedge to be doubly vergent, however there is little evidence of this other than the enigmatic Olympics themselves, where
the seaward vergent roof thrust is overturned, and the Eocene backthrust of the Corvallis Fault in central Oregon (Goldfinger, 1990). No backthrusts have been reported elsewhere in Cascadia bounding the coastal range and inner forearc basin, and the Corvallis structure appears to me primarily strike-slip in its most recent motion. The classical model and its derivatives, as shown in Figure 6-2, are not a good match for the configuration of the forearc at present, particularly the submarine forearc. The model of a continuous accretionary prism spanning the full outer forearc from the deformation front to the rear of the Olympics is most likely incorrect. Most subduction zones have twin forearc basins, a larger inner basin adjacent to the arc, usually onshore, and a typically smaller one offshore. In Cascadia, the inner forearc basin is the Puget Willamette Lowland, and the outer forearc basin is located on the shelf. Figure 6-6 shows a seismic profile across the shelf and outer forearc basin off northern Washington, in line with the Olympic Mountains. Although the Washington shelf is complicated by listric normal faulting strike-slip faulting and diapirism, these form a second order overprint to the fundamental forearc basin structure, shown in this figure. The broad basin is lightly deformed and as such it does not conform to the classical model of Tabor and Cady (1978) in that the rear of the wedge appears to be more intensely deformed than the lightly deformed middle wedge/inner forearc basin represented in Figure 6-6. The high uplift rates attest to continuing high rates of deformation, which can only be explained by either underplating, E-W folding or a combination of these, as the intervening shelf is relatively inactive as is common in many accretionary wedges.

Figure 6-6. East west cross section based upon tomography and earthquake data through Seattle. Intraplate events of 1949, 1965, and 2001 projected onto profile. Upper plate seismicity and tremor locations are shown. Lightly deformed inner shelf are of Figure 6-6 indicated. From Stanley et al. (1999).
Figure 6-4. Velocity model of the Cascadia forearc in Washington at the southern margin of the Olympics, showing potential underthrusting of the accreted low-velocity sediments beneath the Crescent/Siletz terrane. From Parsons et al. (1999).

Figure 6-5. Pre-stack depth migration of the seaward part of line 107 and velocity model, vertical exaggeration 2:1. From Fleuh et al. (1998).
Blakely et al. (2009) show that the Saddle Mountain Fault and other structures on the eastern Olympic Peninsula are closely related to the Seattle fault, both structurally and in the coincident timing of the most recent earthquake ~ 1100 years ago. If correct, the Olympics are at least in part, also involved in N-S shortening of the forearc. This interpretation is shown in Figure 6-7 from Parsons et al. (1999), where the Olympics are not part of a backthrust doubly vergent system, but rather a large E-W fold that plunges eastward. This fold, like the other mapped in this study, those in McCrory (2002) and other sources, is accommodating N-S crustal shortening, as evidenced by the thrust faults bounding the southern and northern flanks of the Olympics.

The N-S shortening is enhanced by the fact that a slab arch exists due to the oddly concave configuration of the subduction zone. The Olympic complex sits directly on this arch, enhancing its uplift (Figure 6-7; Wells et al., 1998; Parsons et al., 1999). The Olympic fold is paired with a synformal feature presently occupied by the Juan de Fuca Straits (Snively, 1987; Parsons et al. 1999). Our mapping supports this interpretation (Figure 6-1), and clearly shows doubly vergent thrust faults offsetting young strata and the seafloor, verging north near Vancouver Island, and south near the Washington coast.
Figure 6-7. Simplified view of Washington tectonics. The Olympic Mountains impinges on the Vancouver Island margin, which has previously been accreted on to the North American craton. As a result, western Washington is deformed by a series of folds and thrust faults, driven by the clockwise rotating OCR block. From Parsons et al., (1999).

We consider that the evidence of folding and faulting observed offshore, coupled with the forearc framework that is consistent with the model of Parsons et al, and inconsistent with the simple classical model of Tabor and Cady (1978) and Steward and Brandon (2004). This model is consistent also with the evidence for block motions, rotations, and a strain gradient seen in GPS data. The model can be considered a damped wave train, with the largest amplitudes being the JDF synform and the Olympics, juxtaposed against the backstop of the Vancouver Island terrane, and diminishing southward.

Within this context, the stress transition line shown in Figure 6-1 indicates the boundary between arc parallel and arc normal contraction. While McCrory (2002) distinguishes an Olympic block, and consider that the block boundary forms the northern terminus of the transverse folding in Washington, we consider that that boundary is Vancouver Island, and that the Olympic antiform is part of the fold train (see also Figure 6-8). Because of the apparent strain gradient, it is likely that this boundary lies further offshore of northern Washington, and thus meaning of the boundary in terms of its relationship to basal traction is likely not fully consistent along strike in Cascadia. The internal structure of the inner wedge and its potential for elastic strain accumulation is unclear. Seismic images are largely opaque below the MBF contact. A velocity model in Fleuh et al. (1998; Figure 6-5) suggests a high-velocity block underlies this contact. While the MBF certainly contains higher velocity blocks, it is unclear how this block behaves mechanically.
Figure 6-8. Gravity image of the northern Washington margin showing the Olympic peripheral and core units, and the upper plate bounding structures. From Blakely et al. (2009).

Given the stronger arc parallel deformation at the northern limit of the OCR/Olympic block, the balance between anelastic deformation responding to arc parallel stress, and the residual anelastic deformation driven by basal traction may be offset seaward of its position further to the south (Figure 6-1). At the same time, this area may be floored with older accretionary units as exposed in the Olympic core, and not the Hoh MBF exposed onshore and offshore in SW Washington. If so, interplate coupling could be quite strong along the northern Washington coast, and could extend inland to a greater extent than suggested by the stress line in Figure 6-1. McCrory (2002) notes that the N-S northern Washington shelf structures extending onshore imply interplate coupling may be operative there.

6.1.2.2 Crustal Thinning and Possible Decoupling in SW Washington

The thinned and subsided area on the WA shelf coincides with the seaward structural “bulge” as well as a high abundance of diapirs. This area is loosely bounded at its north end by the NNF. At the south end it is bounded by the northern margin of Nehalem Bank structural uplift which is located at the Astoria canyon. Figure 6-9 shows a N-S seismic profile extending along the mid Washington shelf from near Astoria Canyon to 47.97 N. This profile shows that the key contact between the Miocene and younger stratigraphy and the opaque MBD Hos assemblage below
has significantly subsided in the area occupied by the seaward bulge and most extensive normal faulting. While most of the listric normal faults appear to sole out at this contact (McNeill et al. 1997), others extend into the Hoh/MBF assemblage. This suggests that the broad crustal thinning involves more than just the Miocene and younger cover sequence. This subsided area is bounded by gradual uplift to the north coinciding with the Olympic uplift, and to the south by the structurally uplifted Nehalem Bank (Figure 6-9).

Coincident with the crustal thinning in the “bulge” area, the E-W structures of the inner shelf domain extend seaward to a greater extent, partially filling the inner part of the bulge (Figure 6-1). The stress transition line correspondingly follows the bulge seaward in this area, implying a seaward swing in the downdip edge of interplate coupling. Because of the complication of crustal thinning in this area, confidence in this interpretation is reduced. Long-term crustal thinning of the upper ~ 1/3-1/2 of the upper plate, at a minimum leaves only a thin ~ 5-7 km thick lower part of the upper plate to accommodate elastic strain accumulation if it is fully decoupled at the mid crustal Hoh/MBF detachment.

South of the “bulge” and potentially reduced coupling of SW Washington, margin parallel structures once again extend to the coast and onshore south of Willapa Bay (Figure 6-1). Figure 6-9 shows that this trend is accompanied by a rise in the level of the Hoh/MBF surface. This onshore trend continues south to ~ 45.7 N and encompasses Nehalem Bank, strongly deformed uplifted structural zone. We interpret the stress transition to be well onshore in this region.

6.2 Central and Southern Cascadia Margin of Oregon
6.2.1 Updip Transition

The low taper, wide fold spacing, landward/mixed vergence province as previously described in detail for the Washington margin extends southward to the northern Oregon margin. This province and the adjacent upper slope-lower slope boundary trends WSW off northern Oregon as the lower slope province narrows southward. The province terminates abruptly at the Daisy Bank Fault where the tapered province reaches zero width, and the landward vergent frontal thrust changes to seaward vergence across this structure (Figure 6-12; Goldfinger et al. 1996b, 1997). This tapering of the landward vergent province corresponds to the tapering and termination of the constructional Astoria Fan. The association between the fan and landward vergence has been previously described. Otherwise, this province and the interpretation of an updip transition is similar to that previously described for Washington.

We interpret that for the remainder of the Oregon margin, the frontal thrusts are for the most part seaward vergent, and little or no updip transition can be mapped on the basis of structural contrasts. Two exceptions to this can be seen in Figure 6-12 and Figure 6-13. Along the outer central Oregon margin, the frontal wedge is disrupted by the deep-seated mega landslide province described by Goldfinger et al. (2000). In this area, shown by a grey polygon in Figure 6-12 and Figure 6-13, structures cannot be mapped as the landslide blocks are extensive, thick and chaotic. In this area therefore, it is not clear whether the partially buried landslide debris is part of the North American plate, the JDF Plate, or whether it is competent enough to participate in strain accumulation. We suspect this is unlikely, and have mapped the updip transition to exclude this area. A small area of landward vergence also exists just to the north of the Rogue Canyon area. The landward vergence here is related to the older mega landslides, and the initiation of new thrusting over the top of the older landslide debris aprons (Goldfinger et al. 2000). We find no substantial reason to exclude this small landward vergent area from the upper plate strain accumulation, and map the updip transition at the surface in this area.

6.2.2 Downdip Transition

6.2.2.1 Low Deformation, North Central Oregon Shelf

Further south in Oregon, the stress transition moves offshore south of Nehalem bank, at the point where the Cape Falcon structure swings onshore. The offshore swing in northern Oregon is not mapped primarily on the basis of transverse structures and the juxtaposition of them against margin normal structures. This area is known as the Newport embayment and extends from ~44.9 N to ~45.6 N. In this region, the shelf edge swings well landward in a broad concave seaward configuration. Landward on the shelf, active deformation is almost completely absent. The broad and deep forearc basin there (see McNeill et al. 2000) is almost completely undeformed. For example, along strike, the major forearc basin structure, a broad synform known as the Newport syncline trends NNW from the coast near Alsea bay, to the outer shelf near Nehalem Bank. This structure has been previously described and structure contours of it were published in McNeill et al. (2000). At its north and south ends, this feature is clearly compressional, with other parasitic folds and thrusts deforming both limbs. In the central region, between the two banks, this is not the case. In that area, the feature has no deformation on its eastern limb, which instead is characterized by numerous small east dipping
normal faults (Figure 5-14 and Figure 6-12). The total throw on the aggregate of these structures is small, and they clearly play no role in the development of the structure itself, yet they are pervasive and indicate a component of extension at a high angle to the structural trend (E-W?). The western limb of the major synform is in places deformed and merges with the more active fold thrust belt to the west. Another example is a small parasitic fault fold that found in the axis of the Newport syncline. This is a small structure with no evidence of post Miocene displacement in the central region between the two major banks. At the south end however, the inactive syncline transitions to a more active structure, and the inactive parasitic thrust at its axis grows rapidly to be a prominent surface feature known as the Stonewall anticline (Figure 5-14). It splits the major forearc basin in two, and breaks the surface as Stonewall Bank (Yeats et al., 1999). Interestingly, Stonewall Bank has been the locus of several small plate boundary earthquakes, a seismic hotspot that has persisted for a number of years on the plate boundary below the stonewall area (Trehu et al., 2008; 2015). We suggest that this is probably not coincidental, and that traction on the underlying plate boundary at this site may be driving Stonewall Anticline. This association is perhaps the first direct link proposed between active upper plate structures and interplate coupling.

The pattern of activity observed for the major Newport syncline is also reflected in other structures in the same region between Nehalem and Heceta Banks. In general, structures are older, activity rates lower or altogether absent on the shelf (Figure 5-14). For a significant part of this reach of the continental shelf, there are essentially no active folds that are parasitic on the eastern limb of the main Newport syncline, and as noted above, this area is instead occupied by evidence of minor extension (Figure 5-14).

### 6.2.2.2 Heceta Bank, Coos Basin, and Coquille Bank

The Stonewall anticline, Pepetua Bank, and Heceta Bank are part of a large structural uplift on the central Oregon Cascadia margin. In this area, strong deformation has folded, faulted and uplifted this area ~ 1km since Miocene time (Kulm and Fowler, 1974). Heceta Bank may also accommodate the accretion of one or more small seamounts outboard of the western margin of the Siletzia terrane (Trehu et al. 2012). Margin parallel and NNW trending structures extend to the coast in this area (Figure 6-13), and this our stress transition is onshore in this area. Deformation of the forearc basin is strong as the previously discussed Stonewall anticline splits and deforms the inner forearc basin and underlying Siletzia rocks (Figure 6-10). Figure 6-11 shows a detailed section through central Heceta Bank showing the intense deformation and uplift outboard of the Siletzia block. In this section the forearc basin syncline is less deformed than in the section shown in Figure 6-10, perhaps due to thicker Siletzia units shielding the Miocene rocks from compressional strain.
Figure 6-10. Cross section through northern Heceta Bank and the Stonewall anticline along seismic profile WO-22. This major anticline overlies the Siletzia volcanics, which are strongly deformed in the core of the fold. See Yeats et al. (1999) for details.

Figure 6-11. Cross section through central Heceta Bank along seismic profile WO-18. The inboard syncline overlies the Siletzia volcanics, which are strongly deformed on its western limb. Magnetic profile and wells confirm the presence of the Siletzia basement.
Figure 6-12. Structural interpretation and activity rates, Northern and Central Oregon margin, updip transition, downdip transition and shelf edge are shown. Symbology as in Figure 6-1.
Figure 6-13. Structural interpretation and activity rates, Central Oregon margin. Symbology as in Figure 6-1.
South of Heceta Bank, the shelf edge swings landward in the Coos Basin. Structures in this area are generally margin parallel to near the coast, and this offer little constraint on the downdip transition. Because the deformation is much less intense however we suggest that the coupling strength may be reduced similar to the pattern previously discussed for the Newport embayment. This entire section of the margin however is underlain by the Siletzia terrane, which may exert strong controls on the pattern of deformation observed in the upper plate. The Coos Basin is terminated to the south by Coquille Bank. This bank is the smallest spatially and least deformed of the three major structural banks of Oregon, but otherwise shares the characteristics of intense deformation and uplift of the upper slope and mid shelf. Structures in this area are generally margin parallel and curve gently seaward around Coquille Bank. The onshore projection of structures place the stress transition onshore in this area. The Coquille Bank area is concurrently deforming on a NW trending anticline known as the Cape Blanco anticline. This structure deforms marine terraces onshore (Kelsey et al. 1990), and extends offshore to Coquille bank. This structure implies a component of NS compression in this area, but unlike other areas, there is not a sharp transition to EW compressive stress. Rather, the broad, low amplitude Cape Blanco Anticline uplifts contemporaneous N-S trending structures. We interpret this to represent either an underplating signal of uplift, consistent with the broad feature, a subducted feature, suggested by the gravity high that underlies it (Figure 6-25) or tectonism related to the hinge point between the Kamath and Siletzia terranes as proposed by Wells et al. (1998), as opposed to N-S compression as suggested by Kelsey et al. (1990).

6.3 Southern Cascadia Margin of Northern California

6.3.1 Updip Transition

On the remainder of the Oregon-Northern California margin, the frontal thrusts are for the most part seaward vergent, and little or no updip transition can be mapped on the basis of structural contrasts. One exception to this can be seen in Figure 6-14. A small area of landward vergence exists just to the north of the Rogue Canyon area. The landward vergence here is related to the older mega landslides, and the initiation of new thrusting over the top of the older landslide debris aprons (Goldfinger et al. 2000). We find no substantial reason to exclude this small landward vergent area from the upper plate strain accumulation, and map the updip transition at the surface in this area. Along the outer wedge on much of the northern California margin, landward vergence also exists. The landward vergence of the frontal thrusts in this area contrast with those of the Washington margin in that they do not correspond to the accretion of a submarine fan, except for the area adjacent to the small Eel Fan. These faults do not comprise a low taper, wide fold spacing fold thrust belt as in Washington and are conformable in strike with folds of the mid and upper slope. Examination of these structures reveals that the vergence appears to be controlled by the steep, seaward dipping backstop of the older accretionary complex that extends to near the deformation front. We therefore do not map these structures as part of a decoupled updip transition, and suggest that slip to the surface is possible in this area.
Figure 6-14. Structural interpretation and activity rates, southern Cascadia (California) margin. Symbology as in Figure 6-1. An additional oval area on the mid slope, inferred as weakly coupled is shown at center.
6.3.2 Downdip Transition
Along the Northern California margin, several unique features exist that complicate interpretation. Generally, structures are margin parallel and do not exhibit a stress transition. The coast line trends SSW and most structures intersect the coast and extend onshore. We therefore infer that much of the California margin downdip transition is most likely onshore. A possible exception lies between 41.8 N and 42.4 N. In this area, much like the Coos Basin, the broad forearc basin is very lightly deformed (Figure 6-14). Both north and south of this area, deformation is intense and of relatively short wavelength. We suggest that this lightly deformed area could indicate reduced basal shear stress, but could also be related to unobserved details of the underlying forearc framework. The other exception is the unusual area on the mid slope of the Eel River Basin previously described. This odd semicircular depression is underlain by a low-angle listric normal faults that also extends further to the north than the obvious bathymetric depression (Figure 5-27, Figure 5-28). There appear to be several strands of this shallow seaward dipping extensional system that may be responsible for the morphology of the mid-slope terrace, though there could be other causal factors as well. Though shallower than the Washington and Northern Oregon large listric normal faults, they are broadly similar, and have rollover anticlines paralleling the traces, and smaller extensional structures in the hanging wall blocks, attesting to thinning and extension of this block). Further west (downdip) evidence of extension in the mid slope terrace gives way to compressional thrust faulting associated with the active accretionary prism. Unlike the extended terrane of the Washington shelf, this unusual subsiding region lies sandwiched between compressional structures both outboard and inboard of it. At the same time, it also shares relatively light deformation with several other areas that may have reduced basal shear stress, though the internal structure of this area below the listric detachment surface is poorly imaged. We cannon infer low coupling in this area however based on structural relations alone.

6.4 Synthesis

6.4.1 Fit to GPS and other Datasets
We compare the structurally derived hypothetical coupling “limits” to coupling models based on other data. We focus on GPS, levelling, and coastal subsidence models, as other proxies such as heat flow and ETS limits have either low spatial resolution (heat flow) highly uncertain modeling (heat flow), or ambiguous applicability (ETS).

6.4.1.1 GPS and Levelling and Joint Inversions
GPS velocity fields have been modeled for the contribution of plate locking to the regional velocity field since ~ 1996. Campaign plus permanent site velocity fields (e.g. McCaffrey et al. 2000) provide the greatest spatial density over those based on a smaller number of permanent sites (e.g. Miller et al. 2001). The high spatial density available in Cascadia has primarily been due to the efforts of Rob McCaffrey, who has published numerous evolutionary improvements (McCaffrey et al., 2000, 2007, 2008, 2012, 2013). We focus on these models as lower density velocity fields lack the spatial resolution for comparison to the structural data.
Early models that focused on levelling data resulted in two very different views of heterogeneity or lack thereof. Figure 6-15 shows two early locking models by Mitchell et al. (1994) and Hyndman and Wang (1995). The Mitchell et al. model shows significant heterogeneity along strike, and uses little smoothing. It honors the apparently highly variable levelling data, in particular the Newport-Albany (Oregon) line, which shows no eastward tilt. In contrast, the Hyndman and Wang model applied significant smoothing to the same data, and integrated tide gauges, producing a very smooth overall model. For many years, the question has been, which one is closer to reality? Burgette et al. (2009) produced a model, again from the same levelling data, that was similar to the Mitchell model.

With the advent of GPS time series, another tool began to play a role in this issue. Figure 6-16 shows several GPS locking models from McCaffrey et al. (2007). These models show several variants and smoothing factors of the models which are based on joint inversion of the GPS and vertical levelling data. Like nearly all GPS coupling models, these assume full plate locking at the estimated convergence rate, and allow plate locking to the deformation front. This has become a common convention in Cascadia, despite geologic evidence to the contrary previously described here (Goldfinger et al. 1992, 1996b, 1997; Priest et al. 2009). These models remove the clockwise block rotation of the forearc, leaving the elastic locking signal from the megathrust, plus any residuals and other unknown signals (McCaffrey et al. 2000, 2013). A significant limitation is the near complete freedom of the updip configuration of the model. As all stations are onshore, the updip region places no constraint on the model. Many configurations have been tested, with similar fit to the data (C. Goldfinger and R. McCaffrey, unpublished data; McCaffrey et al. 2013). Most investigators have adopted these models, perhaps without realizing that the updip regions are completely unconstrained. In general though, the downdip areas are much better constrained due to their proximity to the GPS and levelling array. The down dip limits suggested by the McCaffrey et al. (2007) models are broadly similar to those of previous models based on heat flow, tide gauges and levelling (e.g. Hyndman and Wang, 1995, Wang et al., 2003). The downdip edge shows some heterogeneity depending on smoothing factors applied. Reduced smoothing results in a downdip edge (<10% locking fraction) that appears compatible with both the earlier Mitchell et al (1994) model, and our structural model in Oregon (Figure 6-16), but either misfits or does not resolve our low coupling area off SW Washington.

McCaffrey et al. (2013) presents two new models with additional data (Figure 6-17). Model Pn1d is similar to the 2007 models with minimal smoothing applied. Model Pn2d represents a significant departure in that while fitting the velocity and levelling data, it shows a locking geometry that tapers seaward and landward, and has locking contours centered on the mid slope. Closer examination of this model reveals a surprisingly good fit to the geologic data of this report. In Washington, both Pn1d and Pn2d match the low coupling area off Willapa Bay, and the strongly coupled area north of Grays Harbor. The Pn1d model shows greater heterogeneity of the downdip margin, but both generally fit the structural model. This is in contrast to previous models which generally show a broad locking contour extending onshore. In Oregon, the broad seaward swing of the Newport embayment is also fit by both models, which show a locked patch at Nehalem Bank, and another at Heceta Bank. We also see a good fit to the seaward swing at Coos Basin, and a locked patch that matches Coquille Bank. The models do not extend into California.
Figure 6-15. Early locking models of Mitchell et al (left) and Hyndman and Wang (1995) right.
Figure 6-16. McCaffrey 2007 Locking Models.
Figure 6-17. Locking model results for two parameterizations of locking (see text for functions). Colors and contours are of the slip deficit rate, in mm/yr. Slip deficit rate contours are 5, 15, 25, 35, and 45 mm/yr. (a) Tapered transition zone of variable width, depth, and taper but locked to trench (pn1d). (b) Gaussian distribution of locking with depth (pn2d). From McCaffrey et al. 2013.
Figure 6-18. (a) Uplift rates in south-to-north profile along the coast. Gray symbols are observations from leveling, purple from tide gauges, red from continuous GPS and blue from survey GPS; error bars are one-sigma. Curves are predicted from models, coded by color. Sites plotted are within 50 km of the profile line. (b) Map of residuals of vertical leveling rates along Oregon coast (model pn2d). Profile of locking and fits to data across the Cascadia subduction zone at 44.6 N, line SB-E, where uplift rates are low, for four models discussed in the text. (c) East component of the GPS velocities, error bars are one-sigma; (d) uplift rates from leveling [from Burgette et al., 2009] error bars are one-sigma; (e) locking fraction profiles (1.0 indicates fully-locked and 0.0 is freely-slipping). Gray lines show the topographic profile (in km, based on mm/yr. scale), triangles show the volcanic arc and the coast is at the westernmost observation point. Colored curves show the model predictions. From McCaffrey et al. 2013.
Figure 6-19. A. Full margin depiction of the upper and lower transitions based on structural proxies. Major structural uplifts Nehalem Bank (NB), Heceta Bank (HB), and Coquille Bank are shown. Locking contours of McCaffrey et al. (2013) model Pn2d model are shown as slip deficit in mm/yr. 10 mm/yr. = Blue dashed; 25 mm/yr. = yellow dashed; 40 mm/yr. = orange dashed. B. Structural geology and upper and lower transition lines with major submarine banks shown as in A.
In Figure 6-18, panel A shows the contributing data from which the models were developed in a coast-parallel line. This show the low uplift rated based on multiple datasets that suggest either low coupling, or perhaps a very broad transition zone, in SW Washington and Northern Oregon, with intervening high uplift and presumably strong coupling. The earlier model of Mitchell et al. (1994) closely resembles the pattern in the Pn2d model, and also is a good match for our structural model.

6.4.1.2 Coastal Subsidence

Coastal subsidence data exist in the form of subsided marsh metrics based on the vertical shifts in marsh plant species recorded in the sudden marsh subsidence events. Data exist for several events in the paleoseismic record, but the best recorded and most spatially extensive dataset exist for the AD 1700 event. Recent models include Leonard et al. (2010), and Wang et al., (2013). Leonard et al. (2010) considered models extending back ~ 4000 years, based on the onshore paleoseismology and offshore turbidite chronology of Goldfinger et al. (2012; and advance copy was supplied to L. Leonard for this purpose).

Wang et al. (2013) used paleosubsidence data for the AD 1700 event, essentially the same inputs for this event as used by Leonard et al. (2010) and attempted iterative forward models to match offshore slip models to these data. Their preferred result is shown in Figure 6-21. Both the Leonard and Wang models are limited spatially by the lack of data along the Washington margin north of the Copalis River, a significant problem. While there is paleoseismic data at the Waatch Marsh near Neah Bay (Peterson et al. 2013), these data were not available for the Leonard or Wang modeling studies, thus placing no vertical constraint on the elastic response of the upper plate. The Peterson et al. (2013) data are important as they fill a very large gap along the margin between SW Washington and Vancouver Island. These data show that large tsunami sand sheets are present at this marsh, and that tsunami overtopped the 6-8 m Neah Bay barrier ridge at a minimum, implying a ~ 10m flow depth at that location. Subsidence of 0.5-1.0m is indicated for this site, and 4-5 km inland penetration of these tsunami into the Waatch flood plain indicate very large tsunami at this site.

The coastal subsidence models raise a question of ambiguity of the data with respect to slip models. Because the coastal subsidence data are by definition close to the coast, they essentially represent a 1d dataset that does not sample inland to any significant extent. Levelling data do sample inland, but are available only at transects defined by the infrequent E-W highway routes, and the coastal highway. They also represent a ~ 70 year time range, which is generally assumed to be compatible with the 10-15 year span for GPS. When modeling paleoseismic data, the 1d linear dataset is a very poor constraint on the position of the locked plate boundary. This is illustrated in Figure 6-20 which shows a generic elastic dislocation model of a subduction zone. Typically, the locus of coseismic subsidence lies roughly below the downdip edge of the rupture patch, with the subsidence either diminishing landward, or extending into a damped wave landward. Therefore simply mapping the 1d subsidence values along the coast cannot constrain the locked patch configuration without additional information. When levelling data is added, the situation improves along and near the available transects, however as the 2011 Toholu earthquake demonstrated, plate locking and coseismic release do not have to coincide. In that case, the plates were coupled to some degree all the way to the

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arc at a depth of ~100 km (Ikeda et al. 2012). The coseismic release was offshore however. The interseismic signal was subsidence, and the co-seismic signal was also subsidence. Eventually, post-seismic relaxation began to bring the coastline back up, a process that is apparently still ongoing.

Figure 6-20. Generic elastic slip dislocation model. (Top) Cross-section view across the subduction zone with the shaded parts corresponding to the lithosphere. The thick line represents the locked interface, which slips during the giant megathrust earthquakes. (Bottom) Hypothetical pattern of coseismic uplift and subsidence and its geometrical relationship to slip on the locked interface.

In a real case, factors including fault dip angle and slip distribution affect the actual pattern of uplift and subsidence. From Meltzner et al., 2006.

Figure 6-21 shows the results from the forward modeling of coastal subsidence data of Wang et al. (2013). Their preferred model shows a series of locked patches along the margin. The vertical data are shown under the model figure. The large patch off the Washington coast is largely unconstrained, and the Waatch marsh data were not yet available, leaving no data north of the Copalis River. Their model shows a weak locking in SW Washington, which is similar to our structural model, the Mitchell levelling, and the McCaffrey Pn1d model. Further to the south however, the model is a poor match to these models. The Nehalem Bank area, strongly coupled in the GPS, levelling and structural models, is poorly coupled in the coastal model. The Newport embayment, poorly coupled in the structural and GPS/levelling models, is strongly coupled in the coastal model. This pattern continues to ~42N, with the coastal model essentially a mirror image of the structure, levelling, and GPS/levelling models. This is a surprising and unexpected result. Leonard et al. (2010) presents results for additional events, extending back ~4000 years. These models importantly explore the variability in subsidence per event.
Figure 6-21. AD 1700 forward models of slip from Wang et al., 2013.
Figure 6-22. Estimated coseismic subsidence in events 1–9 (T1–T9), plotted against latitude (data details in Table DR1 [see footnote 1]). Gray diamonds represent high-quality estimates from statistical microfossil analyses; gray circles are medium-quality estimates from relative organic content, with macrofossil data on both sides of the contact and/or relative fresh/brackish diatom concentration; white
circles are low-quality estimates from relative organic content, with macrofossil data for one/no sides of the contact. Thick black lines and gray shading represent the weighted average (moving average over 1°) and uncertainty, respectively, of estimated coseismic subsidence. Upward/downward arrows with question marks indicate unquantifiable estimates of uplift/subsidence. Thinner lines (labeled for event 2) show the predicted subsidence for megathrust slip of 10 m and 30 m (light-gray lines) and for the release of 500 yr. of accumulated strain (dark-gray dashed line). See Table 1 for data sources.

In Leonard et al. (2010), the 1700 AD event shows a pattern similar to that of the Wang model, though without the low subsidence at ~ 45 N. Like the Wang model, it is essentially a mirror image of the locking models of McCaffrey, Mitchell, and out structural proxy model. If we then look and the Leonard et al. (2010) results for previous events T3, T4, T5, T6, T7, and T8 and T9, the larger of the events shown in Figure 6-21, we see that T3, a large regional event, shows much the same pattern as AD 1700, as does T5, T5, T6 and T7. T8, one of the largest events in this time range, shows a similar N-S pattern, but less subsidence in SW Washington, ~ 1 m as opposed to 1.5-2 m. T9 shows a similar pattern to T8, but with even less variability along the N-S profile.

The aggregate findings from previous studies and our results suggest a significant mismatch between locking models based on GPS, leveling and our structural proxy, and the coastal subsidence data. We suspect that since the subsidence patterns from event to event are quite similar, that the mismatch is unlikely to be due to variable precision from mixed methods, errors, or other issues. We consider it more likely that that mismatch is due to the previously described ambiguity presented by the coastal subsidence data. In areas where co-seismic subsidence is low, but the levelling, GPS, and structural indications show strong locking, it may be that locking extends further inland than tested in the forward models of Wang et al. (2013). Elastic dislocation models used by Leonard et al. (2010) are based on the very smooth model of Hyndman and Wang (1995) and Wang et al. (2003) shown in Figure 6-15b. They consider the variability observed in the subsidence data to be an effect of slightly variable distance from the smooth locked interface. Given that our locking model agrees closely with the more recent GPS models, and the Mitchell et al. (1994) and Burgette et al. (2009) levelling based models, we suggest that the observed heterogeneity is likely real, and that the Cascadia locked zone is highly heterogeneous.
Figure 6-23. Map showing asperity sensitivity testing, Cascadia margin. Red contour shows slip model tested based on asperities located at structural up lifts (Nehalem Bank and Heceta Bank, outlined in yellow). Grey contour outlines gravity basin model of Wells et al. (2003).

Figure 6-24. Coastal uplift/subsidence data for the AD 1700 earthquake from coastal paleoseismic data. Grey trace is subsidence generated by Wells et al. (2003) gravity based model. Red trace is generated by asperities located at mapped structural uplifts.
6.4.2 Comparison to Other Models

6.4.2.1 Paleoseismic Models

Onshore and offshore paleoseismic evidence from 41 Holocene Cascadia earthquakes strongly suggests that along-strike segmentation plays a significant role in Cascadia (Goldfinger et al. 2012, 2016 in revision). We can briefly compare the locking models discussed thus far with the along strike segmentation observed in coastal and offshore data.

Expanded and more detailed paleoseismic data for Washington (Goldfinger et al. 2016 in revision) have revealed several interesting results that suggest revision of the segmentation model of Goldfinger et al. (2008, 2012). We observed a persistent pair of faint beds in the stratigraphic interval between regional beds T10 and T11. As these beds were among our strongest regional correlators, there was little doubt of the presence and position of the two smaller beds. In related work, we collected several new cores from northern Oregon piggyback basins in 2015. The best of these, in Oceanus Basin, was easily correlated to Hydrate Ridge cores, and then to core TT053-20 in Astoria Channel. There are several beds in these cores in the same stratigraphic interval, between regional T10 and T11. The thickest of these beds, known as T10b and T0f are interpreted extend further north than Hydrate Ridge, which was apparently a data limited boundary. We tentatively correlated these beds to northern Washington based on their persistent stratigraphic position and roughly compatible model ages In Goldfinger et al. (2016 in revision) we revised the segmentation model of Goldfinger et al. (2012), including the northward extension of T10b and T10f described above.

The most significant revision is the northward extension of Segment C ruptures. There had previously been a 150 km gap in the offshore paleoseismic data between Hydrate Ridge and Astoria Canyon. The 2015 Oceanus cores filled this gap, approximately in its center. The straightforward correlation of beds younger than ~ 5000 ybp (the maximum depth of the new cores) allowed extension of “Segment C” ruptures in that age range northward ~ 100 km. The new work also included an old core, TT053-20 that had not been previously analyzed from Astoria Canyon. This core also appears to include Segment C events observed at Oceanus Basin. The segment C events appear to terminate somewhere between Nehalem Bank/Astoria Canyon, and Willapa Canyon (probable) and Quinault Canyon (certain).

Goldfinger et al. (2016 in revision) also include several additional Segment D ruptures, and a single northern segment rupture reported in Goldfinger et al. (2013). They also revise the northern limit of Segment D ruptures based on observations and modeling of tsunamis at Bradley Lake, Oregon reported in Priest et al. (2014). Overall, the turbidite based earthquake record from the Washington/Canadian margin is to a large degree compatible with the paleoseismic records onshore at Willapa Bay and other locales as discussed in Goldfinger et al. (2012) Leonard et al. (2010), Enkin et al. (2013), Hamilton et al. (2015), and now including the northern Washington coast at Waatch Marsh (Peterson et al. 2013). Segment C ruptures however have not been observed at onshore sites off northern Oregon as yet. The revised segmentation model is shown in Figure 3-5.

If we compare the segment model in Figure 3-5 to the available locking models, we see that the northern termination of segment D coincides with Coquille Bank and Cape Blanco Oregon.
Cape Blanco has previously been implicated as a segment boundary based on subducted topography (summarized in Goldfinger et al. 2012), and may be associated with obvious structural boundaries that include the Blanco Fracture zone, two subducting pseudo faults, and the Siletzia-Klamath terrane boundary (Burgette et al. 2009). These features form clear heterogeneity along the plate interface that may serve as barriers or asperities for rupture initiation. Approximately seven Holocene paleoseismic ruptures may terminate near this boundary, which also has a geodetic signature of strong coastal uplift. Along the northern margin, where structural segmentation of the JDF plate is not apparent, significant basement structure is partially masked by thicker sediment supply. The new northern boundary of Segment C coincides with Nehalem Bank, and major structural uplift that extends into SW Washington. It appears that these events do not extend to the north, and therefore appear to die out in the area of low coupling in SW Washington or at Nehalem Bank itself. The two events in Segment C’ appear to die in somewhere in the 100 km span between Hydrate Ridge and the Oceanus Basin. They may die in the low coupling area of the Newport embayment, though this is poorly constrained. North of Grays harbor, the paleoseismic records suggest a large and persistent asperity stretching along the Washington coast to Vancouver Island, consistent with a model of primary control by sediment thickness on the subducting plate (Goldfinger et al, 2003; 2008, 2012; Enkin et al., 2013; Hamilton et al. 2015).

The structural stress transition line and the GPS/levelling locking models show low coupling in SW Washington and central-northern Oregon, at approximately the same location paleoseismic data show rupture terminations of the Segment C and C’ ruptures of Goldfinger et al. (2016 in revision). The geodetic and structural models are generally in good agreement for the southern part of Cascadia as well. The spatial constraints on the termination of Segment D ruptures are insufficient to correlate them with either Coquille Bank or the possible reduced coupling suggested to the south of Coquille Bank.

6.4.2.2 Gravity Models

It has been proposed that readily identified structural or geophysical features - such as forearc basins or gravity anomalies - correlate with slip in historical earthquakes (Wells et al., 2003; Song and Simon, 2003). This has led to the suggestion that the locations of these features might predict the slip patches of future (or past) ruptures. Wells et al. (2003) proposed that the basins are formed by basal subduction erosion during great earthquakes, and thus long-term subsidence is expected there. Their model does not specify the mechanisms by which these areas would be highly coupled as opposed to the more dense gravity highs that are adjacent to the basins. In their paper, Wells et al. (2013) show a number of examples of slip models in comparison to mostly free-air gravity data. While the models purport to show correlation, most show rather poor correlation. In addition, many of the slip models are quite old and poorly determined. Apparent correlations between megathrust slip and forearc structure or gravity usually rely on finite fault models derived from teleseismic or limited geodetic or tsunami data, but the uncertainty in slip models is not usually addressed in this context. Inversions based on teleseismic body waves, surface waves, sparse geodetic measurements, or tsunami waves often give highly variable solutions, and the results are subject to further
uncertainty depending on model fault geometry and structure. The models that appear to be the best fit are those shown for Nankai and for Chile 1960. Wells et al. (2003) estimate that 79% of the seismic moment in the 20th century was released from slip patches beneath forearc basins. This statistic however is strongly influenced by the 1960 Mw9.5 event. It does indeed appear to be a good fit to the forearc basin/gravity low, but since this single events accounts for a large fraction (30-40%) of the seismic moment of the 20th century, it drives the statistical inference that concludes Wells et al. (2003). We conclude that the gravity data presented in Wells et al. (2003) are a poor fit to slip models for many of the earthquakes presented, many are poorly determined, and several are good fits.
Figure 6-25. A. Structure map underlain by gravity surface of Wells et al. (2003). Onshore the surface is Bouguer gravity, offshore it is free-air. Major structural uplifts Nehalem Bank (NB), Heceta Bank (HB), and Coquille Bank are shown. Locking contours of McCaffrey et al. (2013) model Pn2d model are shown as slip deficit in mm/yr. 10 mm/yr. = Blue dashed; 25 mm/yr. = yellow dashed; 40 mm/yr. = orange dashed. B. Structural geology and upper and lower transition lines with major submarine banks shown as in A.
Wells et al. (2003) presented a model for Cascadia in which they proposed that gravity lows along the submarine Cascadia forearc could predict asperities in great earthquakes, following their hypothesis as previously mentioned. Figure 6-25 shows this model. If we compare this to the locking models from structure and GPS, we see that they are poorly correlated. Their model shows a segment boundary off northern Washington that is not observed in the paleoseismic data, nor in the structural and GPS/levelling locking models. The Wells model also shows a single large asperity extending from the Olympics to central Oregon, where the paleoseismic data have at least one substantial segment boundary near Nehalem Bank/SW Washington. They show a segment boundary corresponding with Heceta bank, which could be compatible with structural and geodetic models, but for the opposite reasons in the sense that Heceta Bank is a locked patch in the structural/geodetic models, and an unlocked patch in Wells et al. (2003). Wells et al. also have locking patches in Coos Basin, and unlocked at Coquille Bank, the opposed of the structural and geodetic models, and show three patches south of 42N in California where the structure models are ambiguous, and recent locking models from GPS do not exist.

The gravity surface presented in Wells et al. (2003) is a Bouguer model onshore, but free-air offshore. The lack of topographic or regional corrections in this model mean that it is highly influenced by these factors. For example, Nehalem Bank is a highly deformed structural uplift, but appears to be a part of a continuous gravity low the Wells et al. model, which is partly due to this, and also due to the color map used in the Well et al. figure (Figure 6-25). The potential field data, with different color maps, shows this in Figure 6-26.

Figure 6-26. Cascadia potential-field anomalies and geology. A: Aeromagnetic anomalies, transformed to magnetic potential (in nT·km; Blakely, 1995, p. 343–346). B: Bouguer gravity anomalies onshore, free-air anomalies offshore. C: Generalized geology. Blue dashed lines in A bound magnetic anomalies interpreted here as partially caused by hydrated mantle. Black horizontal line pattern shows location of
magnetic anomalies of highest amplitude. White dashed line is location of seismic transect (Bostock et al., 2002) showing evidence of serpentinized forearc mantle (in yellow rectangle). From Blakely et al., 2005.

**Figure 6-27.** Map showing asperity sensitivity testing, Cascadia margin. Red contour shows slip model tested based on asperities located at structural up lifts (Nehalem Bank and Heceta Bank, outlined in yellow). Grey contour outlines gravity basin model of Wells et al. (2003).

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**Figure 6-28.** Coastal uplift/subsidence data for the AD 1700 earthquake from coastal paleoseismic data. Grey trace is subsidence generated by Wells et al. (2003) gravity based model. Red trace is generated by asperities located at mapped structural uplifts.
Figure 6-27 and Figure 6-28 show the results of sensitivity testing of the Wells et al. (2003) gravity based model, and our structural model. Figure 6-27 show the two models in map view, and Figure 6-28 shows coastal subsidence data plotted with expected subsidence and uplift from the two models. We find that the gravity based model is negatively correlated with the trend of subsidence data, while the model based on asperities located at the major submarine bank/structural uplifts is a reasonable match for the paleoseismic vertical trends.

More recent great megathrust earthquakes in Sumatra and the Solomon Islands for which detailed coral and GPS geodesy show that megathrust coseismic slip distributions do not follow the gravity/forearc basin model of structural or geophysical control. For example, abundant coral and GPS geodetic data show that slip during the 2005 Mw 8.7 Nias-Simeulue rupture in Sumatra did not occur beneath the forearc basin, but was instead focused beneath the outer arc high. Thanks to the presence of outer arc islands that allowed abundant measurements of surface deformation in the region surrounding the slip patch, the slip distribution for this megathrust rupture is among the best yet obtained - and the data show that slip occurred well updip of the forearc basin and squarely on a positive gravity anomaly.

The three main slip patches of the Mw 9.15 Sumatra earthquake, while less well determined, also do not lie below forearc basins for the most part. The southern patch straddles the forearc high, forearc basin boundary, while the northern patches lie below structural/gravity highs. Goldfinger et al. (2006) noted the lack of correlation with gravity lows for the great earthquakes in Kamchatka, 1952, Alaska, 1964 as well, but also noted the apparent good fit for Chile, 1960.

Bassett et al. (2015) show that the higher density rocks responsible for the gravity high in the NE Japan forearc correlate formed the asperity in Tohoku and correlate well with the main slip patch from 2011. The argue that the higher density rocks have a higher shear modulus and thus are more capable of storing elastic strain, a simple straightforward hypothesis, and therefore would be expected to be the loci of strain accumulation. This is essentially a well determined updated version of the original asperity hypothesis. Their correlations from the 2011 earthquake are shown in Figure 6-29 and Figure 6-30.
Figure 6-29. A, Residual topography. Dashed red and grey lines mark the forearc segment boundary and the trench-axis respectively. Triangles show arc volcanoes. Hk, Hokkaido; E, Erimo seamount; T, Tokyo; K, Kashima. B, Residual gravity anomalies. The black dashed line marks the intersection of the subducting slab with the forearc Moho. C, Mean forearc density anomalies. Contours (5-km increment) show forearc crustal thickness. D, Profiles (grey lines) perpendicularly traversing the forearc segment boundary showing the south-to-north increases in topography (~0.8 km), residual gravity anomaly (~60 mGal) and mean forearc density anomalies (~150–200 kg m⁻³). The mean for each ensemble of profiles is plotted in black. From Bassett et al., 2016.
Figure 6-30. A, Instrumental earthquake record. Grey plus symbols show the epicentres of earthquakes in the JMA catalogue (1923–2015) with $M_I \geq 6.5$. Dashed ellipses show the aftershock area of thrust earthquakes in the Global Centroid Moment Tensor (http://www.globalcmt.org) catalogue (period 1976–2014) with $6.5 \leq M_w < 8$. No aftershock areas cross the forearc segment boundary (dashed red line). B, Rupture areas for large ($M_w > 7.0$) megathrust earthquakes between 1896 and the 2011 Tohoku-oki earthquake. C, Coseismic slip contours (10-m increment) for the 2011 $M_w 9$ Tohoku-oki earthquake. The 20-m slip contour (thicker contour line) defines the Tohoku asperity. D, Inter/postseismic deformation. Contours show interseismic back-slip rate (increment 2 cm yr$^{-1}$). Arrows show 1-year postseismic displacements of seafloor GPS sites. The fast seaward motion of site FUKU is associated with shallow afterslip. From Bassett et al., 2016.

7. Conclusions

Structural mapping of faults and folds in Cascadia is possible due to the very high density of seismic reflection profiles, the near complete coverage of multibeam bathymetry, and numerous sidescan sonar surveys. The initial regional mapping completed in 1994-1997, has been updated with additional data, and significantly improved by the availability of much of the data in digital form, and the availability of integration and analysis and integration packages such as HIS Kingdom.

The structure of the submarine Cascadia forearc can be divided into several distinct domains. The outer wedge in northern Oregon and all of Washington is a landward/mixed vergence wedge with low wedge taper, widely spaced folds, and presence of mud volcanoes. The landward/mixed vergence is linked to high, possibly near lithostatic pore fluid pressure.
based on geologic observations and Coulomb modeling. This domain is separated from an older complex by a significant landward vergent splay fault. The older complex has a steeper taper, and comprises a structural domain that is commonly discordant in strike direction with that of the lower slope domain. In places, the older complex structures are truncated at the splay fault, suggesting a prior episode of frontal erosion, most likely during the Pliocene. Strike directions in the older complex range from NW–N. In several areas notably off SW Washington and NW Oregon, an inner forearc domain with transverse structural trends is identified. The inner structural domain most likely represents forearc response to N-S compression, which has previously been identified with borehole breakouts, focal mechanisms, and onshore structural observations. We have delineated a boundary between the inner forearc and upper slope domains that may represent a change in principal horizontal compressive stress from ~N-S (inboard) to ~E-W (outboard). We propose that this stress boundary marks this stress transition, and also approximately maps the downdip limit of significant interplate basal shear stress. We further propose that the termination of ruptures from offshore paleoseismic data at two segment boundaries off northern Oregon and SW Washington may be due to ruptures dying onto these two poorly coupled regions.

The updip boundary may be located approximately at the transition between the upper slope seaward vergent domain and the landward vergent domain. The likely presence of near lithostatic fluid pressures in the lower slope domain likely overrides the thermal models that suggest locking on this area based on dewatering of smectite clays. Such dewatering likely adds to the overpressuring conditions.

The downdip boundary shows considerable heterogeneity, with broad seaward swings off SW Washington and central-northern Oregon, and landward swings off northern Washington, and the three major structural zones of high deformation and uplift, Nehalem, Heceta, and Coquille Banks. Loci of interplate coupling in at least one GPS model closely match the lack of coupling in the landward vergent region, and strong coupling at the major structural uplifts. The proposed stress boundary, which represents Miocene and younger deformation, closely matches current GPS locking models, and Coulomb wedge modeling, strongly suggesting that locking heterogeneity is related to long-term forearc architecture. Recent investigations of coseismic slip in recent Mw >9 earthquakes suggest similar models likely operate off Sumatra for the 2004 rupture, and NE Japan in the 2011 Tohoku event.

The proposed stress line is poorly correlated to along strike variability of co-seismic subsidence. Possible causes for this mismatch include the lack of areal coverage of the coastal data, the ambiguity of the subsidence data relative to expected subsidence profiles across-strike, a long-term mismatch between strain accumulation and co-seismic slip, and or low precision of many of the earlier subsidence data. Mechanisms potentially responsible for several poorly coupled (creeping) sections of the plate interface, such as a rough plate interface, high abundance of dehydrating clays, subducted features and rapidly accreted sediments are either not present or spatially broader than the relatively narrow creeping regions. On exception is the likely subduction of several topographic highs offshore Cape Blanco which likely control this boundary. We speculate that compartmented regions of high pore fluid pressure due to poorly drained conditions, and bounded by highly coupled regions may be responsible for the two narrow regions of persistent low basal shear stress. These areas may also be responsible for segment boundaries in northern Oregon and SW Washington,
with several paleoseismic terminations suggest that some ruptures may die out in these poorly coupled regions.

The proposed stress transition coupled with GPS locking models suggest not only significant regions of poor coupling, but significant asperities that extend further landward that smoother coupling models currently in use suggest. In particular, significant asperities are suggested off northern Washington, northern Oregon and Central Oregon. The landward position of the downdip locking transitions at these asperities has significant implications for expected PGA values for Seattle, Vancouver, Portland, and smaller communities on the Oregon coast including Astoria, Newport, Bandon and others.

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