Systematic characterization of morphotectonic variability along the Cascadia convergent margin: Implications for shallow megathrust behavior and tsunami hazards

Janet T. Watt and Daniel S. Brothers
Pacific Coastal and Marine Science Center, U.S. Geological Survey, 2885 Mission Street, Santa Cruz, California 95060, USA

ABSTRACT

Studies of recent destructive megathrust earthquakes and tsunamis along subduction margins in Japan, Sumatra, and Chile have linked forearc morphology and structure to megathrust behavior. This connection is based on the idea that spatial variations in the frictional behavior of the megathrust influence the tectono-morphological evolution of the upper plate. Here we present a comprehensive examination of the tectonic geomorphology, outer wedge taper, and structural vergence along the marine front of the Cascadia subduction zone (offshore northwestern North America). The goal is to better understand geologic controls on outer wedge strength and segmentation at spatial scales equivalent to rupture lengths of large earthquakes (≥M 6.7), and to examine potential linkages with shallow megathrust behavior. We use cross-margin profiles, spaced 25 km apart, to characterize along-strike variation in outer wedge width, steepness, and structural vergence (measured between the toe and the outer arc high). The width of the outer wedge varies between 17 and 93 km, and the steepness ranges from 0.9° to 6.5°. Hierarchical cluster analysis of outer wedge width and steepness reveals four distinct regions that also display unique patterns of structural vergence and shape of the wedge: Vancouver Island, British Columbia, Canada (average width, linear wedge, seaward and mixed vergence); Washington, USA (higher width, concave wedge, landward and mixed vergence); northern and central Oregon, USA (average width, linear and convex wedge, mixed and seaward vergence); and southern Oregon and northern California, USA (lower width, convex wedge, seaward and mixed vergence). Variability in outer wedge morphology and structure is broadly associated with along-strike megathrust segmentation inferred from differences in oceanic asthenospheric velocities, patterns of episodic tremor and slow slip, GPS models of plate locking, and the distribution of seismicity near the plate interface. In more detail, our results appear to delineate the extent, geometry, and lithology of dynamic and static backstop along the margin. Varying backstop configurations along the Cascadia margin are interpreted to represent material-strength contrasts within the wedge that appear to regulate the along- and across-strike taper and structural vergence in the outer wedge. We argue that the morphotectonic variability in the outer wedge may reflect spatial variations in shallow megathrust behavior occurring over roughly the last few million years. Comparing outer wedge taper along the Cascadia margin to a global compilation suggests that observations in the global catalog are not accurately representing the range of heterogeneity within individual margins and highlights the need for detailed margin-wide morphotectonic analyses of subduction zones worldwide.

INTRODUCTION

Uncertainty regarding the slip behavior (rupture extent, slip distribution, recurrence) of past megathrust earthquakes in the Cascadia subduction zone (offshore northwestern North America) leads to ambiguity in hazard assessments and limits our ability to prepare for inevitable future events. Recent subduction zone earthquakes in Japan and elsewhere have demonstrated the significant earthquake, tsunami, and landslide hazards posed by coseismic or triggered deformation of the upper plate, particularly offshore and above shallow megathrusts. For example, tsunamis produced by upper-plate fault rupture and submarine landslide generation during the 1964 Alaska earthquake (Parsons et al., 2014; Haeussler et al., 2015) and shallow megathrust rupture in the 2004 Sumatran (Guilick et al., 2011) and 2011 Tohoku-Oki (Japan) earthquakes (Fujiiwara et al., 2011; Sun et al., 2017) were responsible for the majority of fatalities resulting from each earthquake. These events emphasize the need to better understand how the upper plate responds to large shallow megathrust earthquakes in space and time. Cascadia lacks an instrumental record of large megathrust earthquakes, making estimates of past and future rupture parameters (size, slip distribution, recurrence) difficult. However, there are a number of subduction zones worldwide with extensive earthquake catalogs (i.e., Japan, Chile, and Sumatra) where scientists have suggested that the frictional properties of the megathrust, or décollement, are linked to morphotectonic evolution of the upper plate and rupture behavior (e.g., rupture extent and slip amplitude) along strike (e.g., Davis et al., 1983; von Huene and Scholl, 1991; Saffer and Bekins, 2002; Wells et al., 2003; Fujii et al., 2013; Cubas et al., 2013; McNeill and Henstock, 2014; Henstock et al., 2015; Bassett et al., 2016; Saillard et al.,...
In addition, analogue modeling experiments suggest that long-term forearc deformation controls the frequency-size distribution of megathrust earthquakes in subduction zones (Rosenau et al., 2009). For example, along-strike variations in wedge structure and taper along the Japan trench are spatially correlated with seismogenic behavior in northern Japan, the region of the destructive 2011 M 9 Tohoku-Oki earthquake (Fujie et al., 2013; Koge et al., 2014; Bassett et al., 2016). In Chile, observations of along-strike differences in wedge taper (Cubas et al., 2013) and coastal geomorphology (Saillard et al., 2017) were linked to rupture extent in the 2010 Mw 8.8 Maule earthquake and slip patterns of historical earthquakes, respectively. Bathymetric and seismic-reflection profile data along the Sunda subduction margin, offshore Sumatra (Henstock et al., 2006; McNeill and Henstock, 2014; Cook et al., 2014), document a relationship between forearc morphology, structure, and the rupture lengths of the 2004 and 2005 megathrust earthquakes. These inferred relationships based on observations along other subduction margins suggest that a systematic characterization of upper-plate morphology and structure in Cascadia may provide insights into the long-term behavior of megathrust earthquakes, particularly in regard to rupture extent and slip behavior. To date, no such margin-wide characterization of the marine forearc exists in Cascadia.

To fill this important knowledge gap, we: (1) present a margin-wide systematic characterization of first-order along-strike variation in outer wedge morphology, taper, and structural vergence based on quantitative morphological and geological analysis and geologic synthesis; (2) explore potential controls on this variability; and (3) discuss implications for shallow megathrust behavior and hazards.

The primary focus of this study is the morphology and structure of Cascadia’s outer wedge (Fig. 1A) above the shallow megathrust, where seafloor deformation during subduction earthquakes is concentrated and can produce tsunamis that threaten coastal communities. Accretionary wedges are typically divided into inner (landward) and outer (seaward) components. The outer wedge is composed primarily of accreted sediments from the oceanic plate and deformed upper-plate material, corresponding to the frontal prism and outer wedge of Kopp (2013) and the outer wedge and transition zone of Kimura et al. (2007) and Wang and Hu (2006). This part of the margin experiences significant active shortening and imbricate thrust faulting, and responds most quickly to changing boundary conditions, providing insight into active subduction processes (Kopp, 2013; Lallemand et al., 1992) and associated geohazards (earthquakes, tsunamis, and submarine landslides). In contrast, the inner wedge is typically composed of older accretionary and/or crystalline basement rocks and has a lower strain rate than the outer wedge. The transition zone between the inner and outer wedge is marked by the outer arc high (OAH; Fig. 1), a broad structural high that bounds the seaward edge of the shelf forearc basins.

Slip behavior along the megathrust is thought to be controlled by changes in frictional properties that are influenced by variations in effective stress and pore pressure (Scholz, 1998; Saffer and Tobin, 2011). According to critical taper theory (Davis et al., 1983), both pore pressure and effective stress are expressed in the morphology and deformation style of the outer wedge. The taper of a subduction wedge ($\alpha + \beta$) describes the general shape of the wedge and is the equilibrium angle between the seafloor of the wedge and the underlying subduction megathrust (Fig. 1B). The angle is a function of the strength of the wedge and that of the megathrust at the base of the wedge. Because direct measurements of megathrust fault strength parameters are rare and require drilling into the megathrust fault (e.g., Chester et al., 2013), the taper of the subduction wedge has been used to infer fault strength (e.g., Davis et al., 1983; Saffer and Bekins, 2002; Cubas et al., 2013; Koge et al., 2014).

The location and evolution of the OAH and the overall shape of the subduction wedge is thought to be controlled in large part by distinct material-strength contrasts, or backstems, within the forearc (Byrne et al., 1993; Kopp and Kukowski, 2003). Changes in material strength and, in particular, the rigidity of the upper plate has been shown to determine depth-varying rupture behavior of megathrust earthquakes (Sallarès and Ranero, 2019). Backstems form kinematic discontinuities that are commonly described as either “static,” no longer deforming and backed by crystalline basement rocks, or “dynamic,” still deforming and backed by...
highly compacted accretionary material (Kopp and Kukowski, 2003). Backstops are commonly associated with a morphologic break in slope and/or the presence of a megathrust splay fault (e.g., Bangs et al., 2009; Cook et al., 2014; von Huene et al., 2016; Liberty et al., 2019). We provide a comprehensive map of backstop boundaries that can be used to infer first-order strength heterogeneities within the wedge and explore their relationship with along-strike variations in morphotectonics. By synthesizing wedge taper, structural vergence, and backstop properties, we start to unearth the geologic controls on margin segmentation, megathrust frictional behavior, and the distribution of potentially active splay faults.

**TECTONIC SETTING**

The Cascadia subduction zone extends more than 1300 km from the Nootka fracture zone offshore British Columbia (Canada) in the north (Audet et al., 2008) to the Mendocino triple junction offshore northern California (USA) in the south, and is the result of subduction of the Juan de Fuca and Gorda plates beneath the North American plate (Fig. 2). Along- and across-strike variations in the tectonic geomorphology of the Cascadia margin are the result of the complex interactions between these three colliding tectonic plates, each with unique physical characteristics that likely influence the rate and style of deformation. The boundary between the Juan de Fuca and Gorda plates is defined by the right-lateral Blanco fracture zone; MTJ—Mendocino triple junction; NB—Nehalem Bank; NE—Netarts embayment; OP—Olympic Peninsula; KP—Klamath Plateau; SE—Silcox embayment. State abbreviations: CA—California; OR—Oregon; WA—Washington.

![Figure 2. Topo-bathymetric map of the Cascadia margin (offshore northwestern North America) highlighting key physiographic features and the relative plate motions between the Juan de Fuca and North America.](https://example.com/image2.png)
late Holocene (e.g., Goldfinger et al., 2017; Atwater and Hemphill-Haley, 1997).

Convergence direction along the U.S. portion of the Cascadia subduction zone is predominantly oblique, and northward the margin bends westward near the Canadian border (~48°N latitude) and the convergence becomes more orthogonal (Fig. 2). Clockwise rotation of the forearc in the Pacific Northwest of the U.S. affects the relative plate convergence direction and rates south of ~48°N latitude (Wells et al., 2002; McCaffrey et al., 2007), as seen in the differences between relative motions calculated between the Juan de Fuca and North American plates (Fig. 2, black arrows; DeMets et al., 2010) versus those between the Juan de Fuca plate and the Oregon Coast Range (Fig. 2, white arrows; Wells et al., 2002; McCaffrey et al., 2007). Near Cape Blanco there is a gradual change in the rotation direction of the Oregon Coast Range relative to stable North America, with areas to the south moving to the north-northwest, opposite to subducting plate motion, and areas to the north moving north-northeast, more parallel to subducting plate motion (McCaffrey et al., 2013, their fig. 2a), resulting in an increase in the convergence rate and decrease in the degree of obliquity south of Cape Blanco (McCaffrey et al., 2007).

**METHODS**

To systematically characterize margin morphotectonics and identify spatial patterns, we first divide the offshore Cascadia subduction zone into an inner and outer wedge, two margin-parallel segments separated by the OAH. In this study, the OAH is defined as the point along a commonly broad structural high that runs along the seaward edge of the shelf forearc basins (e.g., von Huene and Scholl, 1991; Clift and Vannucchi, 2004; McNeill and Henstock, 2014). Along the Cascadia margin, the OAH has been mapped previously from seismic surveys offshore Washington (Booth-Rea et al., 2008), Oregon (McNeill et al., 2000), and northern California (Gulick et al., 1998; Clarke, 1992). We integrate existing mapping with interpretation of publicly available seismic data (see Fig. S1 in the

Figure 3. (A) Topo-bathymetric map of the Cascadia margin showing the distribution of cross-margin profiles 1–46 and the location of the outer arc high (OAH—thick white line) used for morphologic and structural analysis. Cascadia deformation front is marked by the solid red line. Topo-bathymetry is from Ryan et al. (2009). (B) Profile measurements of outer wedge width, outer wedge gradient ($\alpha$), subduction dip ($\beta$), taper ($\alpha + \beta$), and plate age at the deformation front. Black and gray dashed lines indicate first- and second-order morphotectonic boundaries identified in this study (see Fig. 4). Latitude corresponds to where the profile intersects the OAH.
Supplemental Material; Trievenberg et al., 2016; Phrampus et al., 2017) to create a consistent margin-wide map of the OAH (Fig. 3A).

**Outer Wedge Geometry and Morphologic Variability**

Slab depths in current models of shallow subduction zone geometry in Cascadia are based largely on earthquake locations and active-source seismic data (e.g., McCrory et al., 2012; Hayes et al., 2018), and the position of the plate interface can thus have uncertainties. Uncertainties in shallow slab depths range from ~1 to 3 km based on variations in velocity analysis and associated interpretation of the megathrust surface. Because crustal-scale seismic data are limited or nonexistent along portions of the Cascadia margin, particularly offshore southern Oregon and northernmost California (McCrory et al., 2012, their fig. S3), slab depths here can have uncertainties of 5 km or greater, with corresponding slab dips largely unconstrained. This uncertainty in slab geometry leads to varying outer wedge slab-dip calculations between models in Cascadia (up to 70% difference), particularly at the northern and southern extents (Fig. S1 [footnote 1]). Due to the relatively large uncertainties in models of shallow slab geometry compared to those from topobathymetry data (<1 km), we combine the methodology of McNeill and Henstock (2014) to use outer wedge width and seafloor gradient as a proxy for taper with the geomorphic approach of Brothers et al. (2013, 2019) to characterize outer wedge geometry in Cascadia.

Forty-six (46), 200-km-long, margin-perpendicular topobathymetric profiles were extracted every ~25 km along the margin between the Nootka fracture zone and the Mendocino triple junction (Fig. 3A; Fig. S2 [footnote 1]). Profiles were extracted from a Global Multi-Resolution Topography (GMRT; Ryan et al., 2009) grid of 200 m resolution. Depth values were extracted from the grid every 100 m along each profile; each profile extends 50 km seaward of the deformation front and includes a 21-point running average filter (~2.1 km rectangular window) to minimize potential ship-track artifacts in the GMRT data. Profile placement intentionally avoided the axes of major submarine canyons in order to capture the representative morphology of the outer wedge rather than localized processes related to canyronization.

For each profile, we calculate outer wedge width, gradient (α), and subduction dip (β). Outer wedge widths are calculated as the horizontal distance between the deformation front and the landward edge of the OAH (Fig. 1B), similarly to other studies of outer wedge morphology worldwide (e.g., McNeill and Henstock, 2014; von Huene and Scholl, 1991; Clift and Vannucchi, 2004). The average outer wedge seafloor gradient (α), or along-profile steepness, is calculated as the angle between horizontal and a straight line drawn from the seafloor at the deformation front to the OAH (Fig. 1B). The dip angle (β) of the subduction interface is calculated using the slab model of Blair et al. (2013), as the angle between horizontal and a straight line connecting the top of the subducting oceanic plate at the trench to that beneath the OAH (Fig. 1B).

Because subduction zone dip (β) beneath the outer wedge is not well constrained by existing slab models, outer wedge width and α are used as proxies for taper (α + β) to characterize along-strike morphologic variability along the Cascadia margin. Hierarchical cluster analysis was applied using MathWorks software to measurements of outer wedge width and α to provide a measure of statistical similarity at multiple spatial scales and identify first-order (>100 km) and second-order (>50 km) groupings. Earthquake magnitude-length relationships (e.g., Wells and Coppersmith, 1994) suggest that variation at these large spatial scales is likely to reflect processes controlling stress accumulation along the plate boundary that may be released in large (≥M 6.7; Aagaard et al., 2016) earthquakes.

**Outer Wedge Structural Variability**

Deformation style in the outer wedge can be classified primarily by fold vergence. Seaward-vergent folds are generated by landward-dipping thrust faults, and landward-vergent folds develop over seaward-dipping thrust faults (Fig. 1A). The structural style in subduction wedges is thought to reflect a complex interaction of changing sediment and pore-fluid properties with tectonic stress. Accretion in subduction wedges typically involves the formation of seaward-vergent folds (Seely et al., 1974); however, landward-vergent folds have been recognized along numerous subduction margins, including Alaska, Japan, Sumatra, and the Cascadia margin (Byrne and Hibbard, 1987; McNeill and Henstock, 2014; MacKay, 1995). The mechanisms controlling landward vergence are not well understood but may include the following: a frictionally weak megathrust, a strong wedge, the presence of a seaward-dipping backstop, and/or dynamic weakening during earthquakes (Byrne et al., 1993; Byrne and Hibbard 1987; MacKay, 1995; Cubas et al., 2016; Han et al., 2017).

Decades of coastal and marine research in Cascadia provide an excellent context for synthesis of wedge structure and general classification of outer wedge vergence for this study. For instance, bathymetry, seismic-reflection, seismic-refraction, and potential-field data have highlighted considerable along- and across-strike variation in the morphology and structure of the outer wedge (e.g., Goldfinger et al., 1992; MacKay, 1995; Cochrane et al., 1994; Flueh et al., 1998; Hyndman et al., 1990; Gulick et al., 1998; Fisher et al., 1999; Adam et al., 2004; Booth-Rea et al., 2008; Priest et al., 2009; Webb, 2017) and the shelf forearc basins (Wells et al., 2003; Noda, 2016; McNeill et al., 1997, 2000; McCrory et al., 2002), both of which are thought to be important elements in relation to megathrust slip (Wells et al., 2003) and tsunamiigenesis (Cubas et al., 2016). Here we integrate existing, commonly localized interpretations of wedge structure into a margin-wide characterization for comparison with along-strike morphologic variability. We note that publicly available seismic data are lacking for certain portions of the outer wedge (see Fig. S5 [footnote 1]), particularly offshore southern Oregon and northernmost California (between 42°N and 44°N latitude), making characterization of vergence in these areas questionable (see Fig. 4B). We assume the more typical seaward vergence in areas with no available constraints.
Figure 4. Results of systematic morpho-tectonic analysis. (A) Graph of outer wedge width versus gradient ($\alpha$), showing three primary groupings (A–C) based on hierarchical cluster analysis. These groupings divide the margin into four regions. (B) Map of outer wedge structural vergence, with first-order regional boundaries and associated statistical groupings shown with thick dashed black lines, and second-order boundaries with thinner gray dashed lines. Red indicates landward vergence; light blue, seaward vergence; cross-hatched red, mixed vergence. Yellow polygon outlines an area of listric normal faulting along the shelf-slope break (McNeill et al., 1997). Question marks denote portions of the outer wedge where vergence is assumed to be predominantly seaward in areas lacking seismic imagery within the outer wedge (see Fig. S5 [text footnote 1]). Magenta dashed lines outline shelf forearc basins (Wells et al., 2003). Purple lines indicate locations of seismic profiles shown in Figure 7. Cascadia deformation front is marked by the thin black line. Dashed black line indicates the 200 m isobath. Topo-bathymetry is from Ryan et al. (2009). (C) Average topo-bathymetric (black lines) and gradient (blue lines) profiles for each region, illustrating the along-strike variability in margin morphology. Green bars indicate the average outer wedge width in each region. $\alpha$—outer wedge gradient; $W$—outer wedge width; SD—standard deviation. Grey dashed line indicates location of the deformation front; thin gray lines, individual topo-bathymetric profiles for each region; black triangle, approximate location of the outer arc high; trees, approximate location of the shoreline. Yellow shading highlights the two end-member morphologies in Cascadia.
Backstop Geology and Geometry

According to critical taper theory, the variations in outer wedge taper characterized above should reflect changes in the strength of the wedge and/or the megathrust fault. While we cannot directly assess the strength of the megathrust, we can use three-dimensional (3-D) framework geologic mapping to characterize and infer variations in wedge strength.

Backstops within the overriding plate represent distinct material strength contrasts that play a fundamental role in the formation and evolution of accretionary wedges, particularly the OAH (Byrne et al., 1993; Kopp and Kukowski, 2003). In Cascadia, the primary (crystalline) backstop is defined by the boundary between rocks of the Wrangelia, Siletzia-Crescent, and Klamath geologic terranes and the accretionary complex. Increased lithification over time within the outer wedge has resulted in the development of one or more secondary (dynamic) backstops along the margin as well. We characterize both static and dynamic backstop geology and geometry along the margin to explore the relationship between strength heterogeneities in the upper plate and variations in morphotectonics along strike. To do this, we integrate existing regional characterizations of backstop geology and geometry with analysis of key morphologic indicators of backstop location.

Static backstop geology and geometry have been characterized at numerous locations along the margin using a combination of crustal-scale seismic, potential-field, and geologic mapping data (e.g., Hyndman et al., 1990; Clarke, 1992; Snively and Wells, 1996; Parsons et al., 1999; Fleming and Trehu, 1999; McCrory, 2000; Hayward and Calvert, 2007; Trehu et al., 2012; McCrory and Wilson, 2013). The dynamic backstop along the Cascadia margin is interpreted to be the seaward boundary of the Plio-Pleistocene wedge, and its general location along the entire margin has been inferred previously by Snively (1987) and has been mapped in more detail based on seismic-reflection imagery and deep-sea drilling information offshore portions of Vancouver Island, Washington, and Oregon (e.g., Silver, 1972; Mann and Snively, 1984; Hyndman et al., 1990; Adam et al., 2004). The dynamic backstop is difficult to discern in seismic-reflection data offshore southern Oregon due to extensive mass wasting (Goldfinger et al., 2000) and offshore northern California due to the steep gradient along the outer slope. In addition, there are no drill-hole constraints along the outer slope in these regions. Because dynamic backstops commonly correspond with a distinct change in cross-margin gradient or break in slope (e.g., Kopp and Kukowski, 2003, 2005; Cook et al., 2014; von Huene et al., 2016) along other margins worldwide, we extrapolate previously inferred dynamic backstop boundaries where those boundaries coincide with distinct morphologic breaks in slope derived from topo-bathymetry (Ryan et al., 2009). As a result, we provide a comprehensive map of backstop boundaries that can be used to describe first-order strength heterogeneities within the wedge.

RESULTS

Below we provide a general summary of the results of our systematic analysis of outer wedge morphology and the synthesis of wedge vergence and backstop geology and geometry. We then provide a detailed region-by-region synthesis as we explore the relationship between along-strike morphotectonics and strength heterogeneities in the upper plate.

Outer Arc High

The OAH, which marks the transition zone between the inner and outer wedge, is traced along the entire margin in seismic-reflection data and is commonly, but not always, co-located with the shelf break along the 200-m isobath (Figs. 3A, 4B). Offshore Vancouver Island and Washington, the OAH coincides with a distinct shelf break at ~200 m water depth. South of 46°N, the OAH commonly deviates from the shelf break at 200 m and instead follows a subtle mid-slope break that cuts across the Netarts and Siltcoos embayments (Figs. 2, 3). South of Cape Blanco, the OAH and 200 m shelf break do not meet.

Outer Wedge Taper and Vergence

Outer wedge width and α calculated along 46 topo-bathymetric profiles vary considerably along the Cascadia margin (Fig. 3B). The width of the outer wedge varies between 17 and 93 km, and outer wedge α ranges from 0.9° to 6.5° (Fig. 3B). In general, there is an inverse relationship between outer wedge width and α (Fig. 4A), like that documented globally (e.g., Davis et al., 1983). The dip angle (β) of the subducting plate (McCrory et al., 2012; Blair et al., 2013) within the outer wedge varies along the margin between 3.2° and 13.2°, with associated taper (α + β) values ranging from 6.7° to 15.9° (Fig. 3B). In general, the dip angle of the subducting slab steepens at both the northern and southern ends of the margin near the Nootka and Mendocino fracture zones, respectively.

Hierarchical cluster analysis of outer wedge width and α provides a measure of statistical similarity at multiple spatial scales. Here we focus on profile groupings that span ≥100 km along strike—a spatial scale roughly equivalent to rupture lengths of large earthquakes. Three first-order groupings (A–C, Fig. 4; Fig. S3 (footnote 1)) and five second-order groupings (Fig. S3) are apparent along the margin. When plotted in map view, the three first-order groupings divide the margin into four broad geographic regions, with regions 1 and 3 being statistically similar (Fig. 4B; Fig. S4a). Of the five second-order groupings, only four define regions with extents >50 km (Fig. S4b). Those four second-order groupings divide regions 3 and 4 into two smaller subregions (Fig. 4B). A plot of profile 𝛼 versus β (Fig. S6) also shows three distinct first-order groupings, providing verification that outer wedge α and width are an appropriate proxy for taper in Cascadia.

We calculated mean topo-bathymetric profiles for each of the four broad regions (Fig. 4C) to highlight regional differences in mean profile shape and distinct mid-slope breaks. The geometry (width, gradient, and taper) of the outer wedge along the Cascadia margin varies along strike (Fig. 3B), with end-member geometries located offshore Washington and southern Oregon–northern California in regions 2 and 4, respectively (Fig. 4C). Results...
from margin-wide synthesis of structural vergence (Fig. 4B) show that each region also displays a distinct pattern of structural vergence. The following regional discussion of the results will focus predominantly on the first-order morphologic variation.

**Backstop Geology and Geometry**

Synthesis of backstop geology and geometry (Fig. 5) provides a comprehensive regional characterization of wedge strength heterogeneities along the Cascadia margin. While the general dip direction of backstops is shown with the blue arrows in Figure 5, apparent dip values have been estimated only at a few select sites along the margin where two-dimensional geophysical cross-sections exist. These dip estimates and associated references will be described in more detail in the following section.

Each morphotectonic region is characterized by a unique backstop configuration. In particular, static backstop geology, geometry, and distance from the trench vary considerably along strike (Figs. 5, 6; Table 1). Wedge taper evolves in response to changes in megathrust fault strength and/or wedge strength (Davis et al., 1983). We commonly observe distinct mid-slope breaks (changes in $\alpha$) along individual (Fig. S2 [footnote 1]) and regionally averaged (Figs. 4C, 6) topo-bathymetric profiles that correspond to dynamic backstop boundaries. This morphologic response and associated change in wedge taper both reinforce our suggestion that these boundaries reflect lateral strength heterogeneities in the wedge and enable extrapolation of these boundaries along the margin. In summary, the varying backstop configurations along the Cascadia margin are interpreted to represent material-strength contrasts within the wedge that may help to regulate the along- and across-strike taper and structural vergence in the outer wedge.

**Outer Wedge Segmentation**

In this section, we discuss each of our morphotectonic regions in detail, emphasizing the unique outer wedge taper, vergence, and backstop configurations.
Figure 6. Regional conceptual models illustrating the first-order along-strike variability in morphology and structure along Cascadia’s marine forearc in relation to outer wedge taper and strength heterogeneities in the upper plate. Potential megathrust splay faults are identified based on work by von Huene et al. (2016) that shows that these structures are commonly associated with backstop boundaries. Cross-section colors indicate geologic terranes and backstop boundaries shown in Figure 5. Trees depict approximate position of the shoreline within each region. References for regional models are discussed in detail in the text. Note that outer wedge taper is lowest in region 2 and increases to the north and south. OAH—outer arc high; DF—deformation front.

that result in segmentation of the outer wedge. A summary of this first-order segmentation is provided in Table 1 and illustrated in Figures 3–7.

Region 1—Vancouver Island (47.8°–49.5°N)

Region 1 encompasses the eight northernmost cross-margin profiles (1–8) from the Nootka fracture zone southwest to the northern margin of the Juan de Fuca Canyon (Fig. 3). The OAH in this area largely coincides with a distinct shelf-slope break at ~200 m water depth (Figs. 3, 4). The outer wedge offshore Vancouver Island has a mean width and gradient of 52 km and 2.2°, respectively, with a moderately steep toe (~5°) at the deformation front (Fig. 4C). The average taper of region 1 is 11.7°. The outer wedge here has a relatively consistent taper, with a comparatively linear, sloping seafloor; submarine canyons (Fig. 2) cut across the slope in the southern half of the region between profiles 4–8 and are more incised south of profile 7. None of the submarine canyons in this area are currently (i.e., during present sea-level highstand) connected to fluvial sediment sources.

Fold vergence imaged in seismic data within the outer wedge is predominantly seaward, with the exception of relatively localized fold-and-thrust structures along the deformation front (Figs. 4B, 7A; Hyndman, 1995; Clowes and Hyndman, 2002; Johns et al., 2012; Riedel et al., 2018). The frontal thrust within this region exhibits alternating patterns of landward and seaward vergence in a prominent

Locking depth
- 350 °C (Hyndman and Wang, 1995)
The morphotectonics of region 1 is associated with the highest margin-normal convergence rates (Fig. 2) and a distinct backstop configuration (Figs. 5, 6, 7A). Offshore Vancouver Island in region 1, rocks of the Siletz-Crescent and Wrangellia terranes form the static backstop (McCrory and Wilson, 2013), which is located between 50 and 100 km from the deformation front (Figs. 5, 6). Cross-sections of the region based on seismic data depict this backstop as dipping landward (~30°) to the northeast (Davis and Hyndman, 1989; Hyndman et al., 1990). Davis and Hyndman (1989) noted a distinct mid-slope break within the outer wedge on seismic profiles and suggested the change in gradient marks the landward edge of the Pleistocene wedge, and therefore the dynamic backstop (Figs. 5, 7A). Both the static and dynamic backstops in region 1 dip landward. While the average topo-bathymetric profile shows a mid-slope break within the outer wedge (Figs. 4C, 6), the mid-slope break in map view and along individual profiles (Fig. S2 [footnote 1]) is continually discontinuous. The lack of a distinct continuous mid-slope break in topo-bathymetry within the outer wedge in region 1 suggests that the material strength of the outer wedge here may increase gradually, rather than abruptly, toward the coastline.

**Region 2—Washington (46°–47.8°N)**

As the deformation front bends southward along the Washington margin in region 2 (profiles 9–19), the taper decreases to an average of 8.1° and is reflected in a considerable widening of the outer wedge (~83 km) and a decrease in the average seafloor gradient to 1.4° (Figs. 3, 4). Interestingly, the seafloor gradient at the deformation front is quite steep (>7°) compared to region 1, but then quickly flattens across strike to form a lower-slope plateau. The OAH here is defined by an anticlinal structure that bounds the seaward edge of the shelf forearc basins and roughly follows a distinct shelf break at ~200 m depth (Figs. 3, 4), similar to region 1. We note that in some locations, the OAH anticlinal structure is obscured and/or deformed by active listric normal faulting along the shelf break. In fact, the active extension occurring along the shelf edge in this region may be causing the OAH to migrate landward through time. The continental slope here displays a concave profile with a steep upper slope and a lower-slope plateau, separated by a mid-slope break that is quite distinct along both the regionally averaged (Figs. 4C, 6) and individual (Fig. S2 [footnote 1]) profiles. This part of the margin is incised by numerous submarine canyons (Fig. 2), which served as the primary conduits for sediment transport to the deep sea during the last glacial period (Barnard, 1978). Today, however, most of these canyons are currently cut off from active river systems, with the exception of Astoria and Willapa Canyons, which are thought to transport a small amount of sediment.
In contrast to region 1, the outer wedge here is characterized by a laterally expansive zone (40–50 km wide) of primarily landward-vergent folds documented in seismic data (e.g., Adam et al., 2004; Booth-Rea et al., 2008; Webb, 2017) that transition to seaward-vergent folds on the middle slope (Figs. 4B, 7B) coincident with a distinct change in profile gradient (Fig. 4C). The upper slope and outer shelf in this region are currently undergoing extension as evidenced by numerous listric normal faults imaged with seismic-reflection profiles (McNeill et al., 1997; Webb, 2017). The cause of extension and apparent margin collapse along the upper slope and outer shelf in region 2 has been related to elevated pore pressures within the Hoh mélangé that may enable downslope movement along a detachment surface between the Hoh mélangé and overlying sediments (McNeill et al., 1997). In turn, westward movement of the Hoh mélangé and extensional collapse of the overlying sediments contribute to overall shape of the mid- to upper slope in region 2. Landward of this zone of extension, deformation within the shelf forearc basins is largely compressional with deformation focused along east-west– and northwest-southeast–oriented faults and folds (McCorry et al., 2002). The width of the shelf in this region ranges from 34 to 61 km wide.

The morphotectonic differences between regions 1 and 2 correspond to dramatic changes in backstop configuration. The static backstop in region 2 is arcuate in shape and forms the seaward boundary of the Siletz-Crescent terrane (Fig. 5). The backstop...
boundary dips shallowly landward (Tabor and Cady, 1978; Parsons et al., 1999), similarly to region 1; however, here the backstop is located 100 km to >200 km from the deformation front (Fig. 5). The greater distance from the deformation front corresponds with a wider, lower-taper outer wedge here compared to adjacent regions 1 and 3 (Figs. 4C, 6). Within the outer wedge, the dynamic backstop is characterized by a "triangle zone" (Mann and Snajvel, 1984; Adam et al., 2004; Priest et al., 2009; Webb, 2017) consisting of two fault boundaries with opposing dip. The outboard dynamic backstop dips seaward, and the inboard backstop dips landward (Figs. 5, 7B). The landward-dipping dynamic backstop is marked by a distinct slope break along the mid- to lower slope (Figs. 4C, 6B) and separates older, highly deformed wedge sediments to the east from younger, less deformed wedge sediments to west (Figs. 5, 6). This material-strength boundary corresponds with the across-strike transition from landward to seaward vergence in the outer wedge and has been interpreted as an active system of splay faults that dips ~30° and extends hundreds of kilometers along the margin (Goldfinger, 1994; Goldfinger et al., 1997; Priest et al., 2009; Witter et al., 2013), although definitive evidence of continuous active faulting is lacking along much of this boundary.

Region 3—Northern and Central Oregon (43.5°–46°N)

Just south of Astoria Canyon (Fig. 2) and Astoria Fan in region 3 (profiles 20–31), the outer wedge begins to narrow (52 km) and the gradient steepens (2.5°), corresponding to an increase in average taper (10.3°; Figs. 3, 4). The outer wedge in region 3 has a steep gradient at the deformation front, similarly to region 2; however, the average profile shape becomes linear to convex seaward and is devoid of submarine canyons (Figs. 2, 4C). Second-order groupings further divide region 3 into two subregions (subregions 3a and 3b) at 45°N latitude (Figs. 4B, 5). The boundary between subregions 3a and 3b corresponds to a north-to-south increase in average outer wedge α of ~1° (Fig. 3B). In subregion 3a, the OAH runs along the steep upper slope and coincides with the seaward edge of Nehalem Bank, a bathymetric and structural high along the outer Oregon shelf (NB, Fig. 2). South of Nehalem Bank, the OAH follows a linear break in slope at ~500 m water depth known as Cascade Bench (McNeill et al., 2000), which is located seaward of and at greater depth than the modern shelf break at ~200 m depth (Figs. 2 and 4B). Cascade Bench is thought to represent an early Pleistocene lowstand paleo-shelf edge (McNeill et al., 2000) that is preserved due to subsequent subsidence within Netarts embayment (NE, Fig. 2). The OAH and Cascade Bench continue southward to Heceta Bank (HB, Fig. 2), near the southern boundary of subregion 3b.

While the outer wedge α does not increase dramatically from region 2 to subregion 3a, the vergence within the outer wedge of subregion 3a changes from predominantly landward vergence to mixed vergence (Figs. 4B, 7C), resulting in more-symmetrical fold structures and steeper seafloor gradients along the fold limbs evident in seismic data (Booth-Dea et al., 2008; Webb, 2017; Phrampus et al., 2017) and seafloor imagery (Ryan et al., 2009). The seafloor gradient (α) of the mixed-verge zone in subregion 3a increases abruptly across strike along the lower to mid-slope in a similar fashion to region 2 (Fig. 4B; Fig. S2 [footnote 1]) The wide “triangle zone” that exists to the north in region 2 is not evident in subregion 3a. The area of low α and mixed vergence narrows from north to south in subregion 3a from ~40 km wide at profile 20 to zero at profile 26 (Figs. 3, 4B). Notably, the transition from predominantly landward vergence in region 2 to mixed vergence in subregion 3a corresponds to the boundary between the Nitnat and Astoria canyon and fan complexes, suggesting that perhaps the material properties of the sediments within these two systems are unique and result in variable deformation styles.

The boundary between subregions 3a and 3b is marked by an abrupt along-strike transition from mixed to predominantly seaward vergence in the outer wedge at 45°N (Figs. 4B, 7D). Subregion 3b is characterized by a north-to-south increase in average outer wedge α (from 1.9° to 3.1°) and a subtle increase in slope concavity (Fig. 3; Fig. S2 [footnote 1]). Subregion 3b straddles Heceta Bank, a prominent bathymetric and structural high along the outer shelf where seamounts are inferred to be subducting (Fig. 5), causing localized wedge uplift (Tréhu et al., 2012). The uplift associated with sea-mount subduction may be partly responsible for the increase in outer wedge gradient within subregion 3b. Deformation within the inner wedge of region 3 is dominated by east-northeast-directed compression and dextral translation as suggested by the orientation of structures mapped on the continental shelf (Clarke et al., 1985; Snajvel, 1987; Goldfinger et al., 1992; McNeill et al., 2000).

The transition from region 2 to region 3 again corresponds to an along-strike change in backstop geology and geometry. The static backstop in region 3 is composed of thickened basement rocks of the Siletz terrane (Tréhu et al., 1994) and is located much closer (within 50–110 km) to the deformation front compared to region 2 to the north (Fig. 5). Here the static backstop is vertical to steeply seaward dipping (~90°–70°) based on seismic-reflection, refraction, and potential-field data (Fleming and Tréhu, 1999; Gerdom et al., 2000; Tréhu et al., 2012). The vertical to steep seaward dip of the static backstop here, combined with localized uplift of the accretionary wedge associated with subduction of seamounts (Tréhu et al., 2012), may explain the slightly more convex shape of the outer wedge here compared to region 1 (Fig. 4C) with a similar width, α, and taper. The dynamic backstop in region 3 dips landward, in opposition to the static backstop (Snajvel, 1987), and the width of the Plio-Pleistocene wedge gradually decreases to the south. Again, the dynamic backstop in region 3 is commonly associated with a morphologic break in slope, which we used to delineate its location offshore southern Oregon and northern California, where seismic imagery is lacking.

Region 4—Southern Oregon and Northern California (40.5°–43.5°N)

Profiles 32–46 define region 4 in southern Cascadia where the outer wedge narrows to an average width of 29 km and steepens to an average gradient of 4.2°, corresponding to an increased average taper of 14° (Figs. 3, 4). The average gradient across
the lower slope in region 4 is >10°, and the shape of the outer wedge is notably convex (Fig. 4C) with a number of submarine canyons cutting across the steep lower slope, including Rogue, Trinidad, and Eel Canyons (Fig. 2). Second-order groupings divide the area into two subregions (subregions 4a and 4b) at ~41.5°N latitude (Figs. 3, 4B). In subregion 4a, the OAH runs along the Klamath Plateau (Kulm and Fowler, 1974) to Coquille Bank (Figs. 2, 3). Between Coquille Bank and the head of Rogue Canyon, the OAH roughly follows the shelf break along the 200 m isobath (Fig. 4B). South of Rogue Canyon, the OAH diverges from the shelf break and follows the seaward edge of Eel Basin. The subregion boundary at 41.5°N marks a decrease in average outer wedge width from 35 km to 20 km that is accompanied by an increase in average linear gradient, subduction dip, and taper (Fig. 3). While the outer wedge width remains constant in subregion 4b, the outer wedge gradient varies considerably (between 2.5° and 6.5°).

The availability of existing data to characterize outer wedge vergence varies along strike within region 4 (see Fig. S5 [footnote 1]). North of Rogue Canyon, the structural framework of the outer wedge remains somewhat ambiguous primarily because of a lack of publicly available seismic data crossing the outer wedge here (Fig. 4B). In addition, super-scale seafloor slope failures inferred from proprietary seismic data (Goldfinger et al., 2000) make imaging wedge structure along this part of the margin challenging. Deformation within the outer wedge south of Rogue Canyon consists of a mixture of landward- and seaward-vergent structures along the deformation front (Figs. 4B, 7E). Landward vergence dominates within only a narrow band along the deformation front, with seaward vergence evident inboard (Gulick et al., 1998). Deformation within the Eel forearc basin is primarily transpressional and characterized by northwest-southeast– and WNW-ESE–oriented faults, including the Little Salmon and Mad River fault zones (Clarke and Carver, 1992; Gulick and Meltzer, 2002; McCrory, 2000).

There are three distinct landward-dipping backstop boundaries identified in region 4 (Figs. 5, 6), and again, the backstop configuration here is quite different than in region 3 to the north. The inboard static backstop forms the eastward-dipping fault boundary between the Klamath terrane on the east and various terranes of the Franciscan Complex to the west (Blake et al., 1985; McLaughlin et al., 1994; McCrory, 2000). Farther west within the Franciscan Complex, there is a distinct material-strength boundary between the older (Cretaceous and Jurassic) Central belt rocks and the younger (Paleogene and Cretaceous) Coastal belt rocks and the younger (Paleogene and Cretaceous) Coastal belt rocks that we classify as another static backstop located 50 to 75 km from the deformation front. This backstop boundary, called the Coastal Belt fault (Clarke, 1992), forms the eastern boundary of the Eel Basin south of Cape Blanco and dips ~30° to the north-east based on a generalized geologic cross-section (McCrory, 2000) in southern Cascadia. The offshore extent of this backstop and fault farther to the north is inferred from aeromagnetic, seismic, and borehole data (Clarke, 1992; Snively and Wells, 1996). The location of this boundary was modified north of Cape Blanco in this study to coincide with a nearby (within 5–13 km) morphologic break in slope just seaward of that previously inferred by Snively and Wells (1996). Notably, our inferred backstop boundary intersects a seafloor seep site just south of Coquille Bank (yellow hexagon in Fig. 5), characterized by gas bubbles containing mantle-derived helium suspected to be sourced from the subducting Gorda plate (Baumberger et al., 2018). This spatial correlation suggests that the backstop fault here may be physically and potentially kinematically linked to the subducting plate, providing a pathway for mantle-derived fluids sourced in the subducting plate to vent at the seafloor.

The dynamic backstop in region 4 separates the narrow Plio-Pleistocene wedge from the older accretionary complex rocks (Fig. 5). North of Rogue Canyon, this boundary is delineated based on limited seismic data (e.g., Snively, 1987) and subtle, somewhat discontinuous morphologic breaks in slope visible on individual topo-bathymetric profiles (Fig. S2 [footnote 1]). South of Rogue Canyon, the dynamic backstop marks the boundary between the landward- and seaward-vergent fold domains along the frontal thrust (Fig. 7E). To the south in subregion 4b, the Plio-Pleistocene wedge is so steep and narrow that it becomes difficult to discern from the deformation front at the margin scale (Figs. 5, 6).

## DISCUSSION

### Controls on Regional Morphotectonic Segmentation

Here we compare along-strike variability in marine forearc character, as described by our morphotectonic analysis, with various trends in the subducting and overriding plates along the Cascadia margin (Table 1). The following paragraphs discuss how the geologic framework of the plates may influence the first-order morphotectonic variability of the outer wedge in Cascadia. Qualitative spatial correlations suggest that the first-order morphotectonic heterogeneity in Cascadia is the result of complex interactions among tectonic plates primarily controlled by upper-plate backstop configuration, glacial sediment loading, and subducting plate hydration, with second-order variability modulated by subducting-plate age and structure (Table 1; Fig. 6). We suggest that these parameters define the strength heterogeneity of the wedge at length scales of 50–100 km or greater that controls the morphotectonic response reflected in outer wedge taper and vergence documented here.

### Backstop Configuration

The curvature of the subduction interface (McCrory et al., 2012) is one factor that influences the stress regime and strain pattern in the overriding plate (Bonnardot et al., 2008). The 3-D variations in material strength and relative motions between the subducting and overriding plates play a fundamental role in defining the 3-D geometry (curvature) of this interface and thus should strongly influence the major differences in the stress regime and deformation patterns within the overriding plate. Given that the relative motions between the subducting and overriding plates do not vary considerably south of 48°N latitude at the boundary between regions 1 and 2, the 3-D variations in material strength likely...
play a more important role in controlling morpho-
tectonic changes along the margin south of 48°N. 
This is precisely what is reflected in our analysis and 
is described in detail in the Outer Wedge Segmen-
tation section above. Each of the morphotectonic 
regions identified in this study has unique backstop 
configurations that define the material strength of 
the overriding plate.

In particular, the geology, geometry, and loca-
tion (distance from the trench) of the static backstop 
varies between each of the morphologically defined 
regions (Figs. 5–7; Table 1). The boundary between 
regions 1 and 2 corresponds to the southern limit 
of the Wrangellia terrane that underlies much of 
Vancouver Island. In region 2, the static backstop 
is composed of the Siletz-Crescent terrane and is 
located much farther inland than in regions 1 or 3. 
This variability across regions 1–3 is likely related 
 to the >500-km-long concave-outboard bend in the 
trench of the northern Cascadia margin and associ-
ated orocline bending documented in the Olympic 
Peninsula of Washington (Finley et al., 2019). The 
oroclinal bending has resulted in along-strike vari-
ations in static backstop geometry and distance from 
the trench that also correspond to changes in ter-
rane geology among regions 1, 2, and 3 (Fig. 5). The 
boundary between regions 3 and 4 is also associ-
ated with changes in static backstop geology (from 
thickened Siletz to Klamath-Franciscan), dip (from 
steeply seaward dipping to landward dipping), 
and distance to the trench (increasing southward). 
These north-to-south changes in southern Cascadia 
are likely due to a combination of transpression and 
crustal shortening in the overriding plate and 
complex convergence with an internally deforming 
and relatively young Gorda plate.

Glacial Sediment Loading

The relationship between sedimentation and 
outer wedge morphotectonics can be explained by 
critical taper theory. In order to maintain a sta-
 ble taper angle, the geometry of the wedge must 
evolve through time to compensate for changes in 
accretionary flux or erosion of the wedge. There-
fore, it has been suggested that large-scale shifts in 
sediment flux to the wedge, as commonly occurs 
during deglaciation, can influence the morphology 
and structure of the wedge through changes in pore 
pressure and effective stress along the base of the 
wedge (e.g., Seely, 1977; MacKay, 1995; MacKay 
et al., 1992; Brandon, 2004). We propose that mor-
photectonic boundaries between regions 1, 2, and 
3 are, in part, related to regional changes in glacial 
 sediment loading along the margin.

Source-to-sink glacial-age sediment flux varies 
along the Cascadia margin (Adam et al., 2004, and 
references therein), with the largest influx of sedi-
ment occurring offshore Washington and northern 
Oregon in region 2 and subregion 3a (hatched area 
in Fig. 5), which are characterized by a wide outer 
 wedge, low taper, and predominately landward 
vergence. In contrast, areas with limited or more 
gradual input of glacial-age sediment, such as in 
regions 1 and 4, are characterized by relatively 
narrow and steep outer wedges exhibiting predom-
inantly seaward vergence. In addition, the transition 
from predominately landward vergence in region 2 
to mixed vergence in subregion 3a (cross-hatched 
area in Fig. 4b) corresponds to the boundary 
between the Nitnat and Astoria canyon and 
complexes, suggesting perhaps that the material 
properties of the sediments within these two sys-
tems are unique and result in variable deformation 
styles. Increased glacial-age sediment loading of 
the trench and outer wedge in region 2 and subregion 
3a is thought to be responsible for both rapid Qua-
ternary seaward growth of the accretionary prism 
(Silver, 1972; Barnard, 1973) and development of a 
broad area of landward and mixed vergence along 
this part of the margin (MacKay et al., 1992; MacKay, 
1995; Fisher et al., 1998; Fluhr et al., 1998; Gutscher 
et al., 2001; Adam et al., 2004).

While the formation mechanisms of atypical 
landward vergence remain unclear, most hypothe-
ses require low basal shear stress and/or a relatively 
strong wedge. For example, Adam et al. (2004) pre-
sented kinematic and mechanical analysis of seismic 
data and proposed a mechanism for maintaining a 
broad area of landward vergence in which rapid gla-
cial sedimentation resulted in dramatic thickening 
of the wedge, increased pore pressure along its base, 
and development of a detachment fault ~500 m 
above the oceanic basement surface. More recent 
seismic imaging in this area during the 2012 Ridge-
to-Trench (Carbotte et al., 2015) and 2012 Cascadia 
Open-Access Seismic Transects (COAST; Holbrook 
et al., 2012) seismic experiments suggests that the 
megathrust interface is at the top of the oceanic 
crust (Han et al., 2017; Peterson and Keranen, 2019) 
with no evidence for the low-velocity detachment 
suggested by Adam et al. (2004) to explain the for-
mation of landward vergence. Instead, these studies 
find evidence for a consolidated, well-drained, and 
strong outer wedge in region 2. Further character-
ization of wedge physical properties will be key to 
reconciling these observations and understanding 
the mechanism(s) responsible for landward ver-
gence in Cascadia and elsewhere.

Subducting Plate Structure, Hydration, and Age

Subducting plate structure or roughness has 
been linked to along-strike segmentation of a num-
ber of margins worldwide, including Costa Rica 
(Morgan and Bangs, 2017), Sumatra (McNeill and 
Henstock, 2014), and Chile (Tréhu et al., 2019). In all 
of these settings, the subducting plate feature, whether 
it is a propagator wake, fracture zone, or seamount, 
interrupts the along-strike distribution of sediment 
thickness and properties entering the trench and/or 
provides a pathway for fluids that results in local-
ized or more regional changes in pore fluid pressure 
along the plate interface. These changes in pore fluid 
pressure, in turn, alter the mechanics and geo-
metry of the subduction wedge. Spatial correlations 
between morphotectonic boundaries and subduct-
 ing plate structures along Cascadia suggest a causal 
relationship, perhaps dependent on the hydration 
state of the upper mantle and lower crust.

There are a number of structural asperities on the 
subducting plate in Cascadia (Figs. 2, 5), includ-
ing propagator wakes (Wilson, 2002), seamounts 
(Tréhu et al., 2012), WNW-trending strike-slip faults 
(Goldfinger et al., 1997), and bending faults (Han et 
et al., 2018). Interpretation of active-source geophysical 
imaging, seismicity, and potential-field data along 
the margin suggests that these inherited structures 
can provide important fluid pathways within the
subducting wedge that both locally and regionally influence pore fluid pressure and associated deformation (Tobin et al., 1993; Tréhu et al., 2012; Canales et al., 2017; Han et al., 2018; Stone et al., 2018). We see evidence of this interaction particularly within regions 2 and 3 offshore Washington and Oregon. The morphotectonic boundary between regions 2 and 3 at 46°N marks a north-to-south increase in the hydration of the lower crust and upper mantle (Canales et al., 2017) that is attributed primarily to changes in slab curvature and bend-faulting orientation of the subducting plate (Han et al., 2018). The increased hydration is also related to seamount subduction and localized fluid release along propagator wakes (pw-3 and pw-4, Fig. 5) and the WNW-trending Daisy Bank fault zone (Goldfinger et al., 1997; Canales et al., 2017; Han et al., 2018). Propagator wake pw-4 (Fig. 5) at the southern boundary of region 3 is a ~75-km-wide feature that delineates an area of the subducting plate that underwent significant shear prior to subduction (Wilson, 2002) and may be a significant pathway for fluids within the subducting plate.

There are several other areas along the margin where propagator wakes and WNW-trending subducting-plate strike-slip faults intersect the deformation front (Fig. 5). Only two of the five intersecting propagator wakes (pw-3 and pw-4) and two of the eight or more strike-slip faults (Daisy Bank and Coos Basin faults, Fig. 5) are co-located with morphotectonic boundaries identified here. Perhaps individual faults and narrow propagator wakes only locally affect pore fluid pressures along the megathrust and deformation within the outer wedge, resulting in a limited effect on the first-order along-strike morphologic character of the margin. Expanded investigations of the along-strike variation in the hydration state of the subducting plate and the consolidation state of the incoming sediment package are needed to improve our understanding of fluid circulation patterns and their influence on the morphology and structure of the outer wedge.

Notably, the ~5 m.y. change in subducting plate age between the Juan de Fuca and Gorda plates along the Blanco fracture zone intersects the margin within morphotectonic subregion 4a identified in this study. Both onshore (Kelsey et al., 1994; Burgette et al., 2009) and offshore evidence (this study) suggest that the unique morphotectonics of subregion 4a is due, at least in part, to the subduction of young, buoyant oceanic crust of the Gorda plate. The boundary between region 3 and subregion 4a corresponds to a north-to-south increase in Coast Range topography and long-term coastal uplift rates determined from marine terraces (Kelsey et al., 1994), as well as an increase in interseismic uplift rates determined from tide gauge and leveling records (Burgette et al., 2009). Offshore we observe a ~120-km-long embayment in the trench axis (blue dashed line, Fig. 5), coincident with both a shallowing of the subduction angle (β) and an increase in wedge gradient (α) within subregion 4a (Fig. 3B). We also note that there are no shelf forearc basins along this stretch of the margin near Cape Blanco and suggest that this may be the result of long-term shortening and uplift of the inner wedge in this region. Together, these observations are consistent with the upper-plate response to subduction of young and buoyant oceanic crust. Similarly, subduction of the Cocos Ridge in Costa Rica has been shown to influence trench and wedge geometry (Vannucci et al., 2013) and induce uplift (Morell, 2016; Edwards et al., 2018).

The unique character of subregion 4b is likely related to the complex interactions between the Gorda, Pacific, and North American plates in proximity to the Mendocino triple junction. For example, Gulick and Meltzer (2002) documented a clear rotation of offshore structural trends in subregion 4b that is associated with the northward migration of the Mendocino fracture zone. The very steep coastal topography within subregion 4b is thought to be the result of northward propagation of crustal thickening associated with the subduction of the Mendocino triple junction (Furlong and Schwartz, 2004; Balco et al., 2013). The increased relative convergence and associated crustal thickening may explain the overall steeper, narrower, and more convex outer wedge within region 4.

Links to Megathrust Behavior

Based on the first-order morphotectonic analysis and synthesis presented above, we observe four outer wedge regions that each respond differently to long-term convergence along the Cascadia subduction zone. Here we discuss the potential linkages between the morphotectonic response of the upper plate and shallow megathrust behavior. As discussed above, outer wedge morphotectonics in Cascadia is likely controlled by a variety of factors that, in turn, modulate both megathrust friction and the strength of the wedge. Although this study does not directly address critical taper mechanics, the margin-wide characterization of variable taper, vergence, and inferred variations in wedge strength provides a set of initial conditions that can be tested in future modeling studies. In particular, future efforts should be focused on teasing apart the relative contribution of fault strength and wedge strength to offshore morphotectonics through more in-depth examination and quantification of wedge taper and strength heterogeneity in three dimensions, similar to studies by Han et al. (2017) in Cascadia and Kirkpatrick et al. (2020) in Costa Rica.

While it is commonly assumed that mechanically strong faults rupture seismically and weak faults rupture aseismically (e.g., Kanamori, 1986; Scholz, 1992), recent laboratory experiments (Ikari and Kopf, 2017), numerical modeling (Noda and Lapusta, 2013; Cubas et al., 2018), and observations of shallow slip-to-the-trench in the Tohoku-Oki earthquake (Fujiwara et al., 2011) suggest that weak faults can host seismic-slip behavior. While relationships between fault strength and slip behavior may be complex, spatial variations in morphotectonics in Cascadia suggest that wedge strength and megathrust fault friction, and possibly slip behavior, may vary spatially. For example, estimates of the width of the seismogenic zone based on temperature (e.g., Hyndman and Wang, 1995; Cozzens and Spinelli, 2012), GPS velocities (e.g., Schmalzle et al., 2014), and geologic structure (e.g., Priest et al., 2009) vary widely in Cascadia; however, all models exhibit some degree of along-strike variability in the width of the locked zone. In particular, the width of locking determined from the thermal model of Hyndman and Wang (1995) varies somewhat along strike and mimics the location of the OAH (Figs. 5, 6), suggesting a possible thermomechanical link...
between morphotectonic variability and fault locking. In addition, numerical modeling results suggest that the width of the seismogenic zone in subduction zones may control the occurrence of supercycles and the recurrence interval between the largest events (Herrendörfer et al., 2015). The apparently coupled spatial variability in morphotectonics of the outer wedge and seismogenic zone width supports some level of interaction that may influence the earthquake cycle in Cascadia.

To explore these interactions further, we qualitatively compare first-order and second-order morphotectonic boundaries from this study (vertical dashed lines, Fig. 8) with megathrust segmentation boundaries (gray ovals) inferred from previous work. In general, spatial correspondence between all segmentation boundaries is somewhat variable along the margin, with more alignment in northern Cascadia (north of 45°N latitude) than southern Cascadia. Notably, the first-order morphotectonic boundary at ~46°N latitude between regions 2 and 3 is common to all the studies listed in Figure 8, suggesting interactions between processes occurring at multiple depths within the subduction system. Morphotectonic boundaries are most closely aligned with along-strike changes in oceanic asthenospheric velocities beneath the subducting plate revealed through onshore-offshore teleseismic P-wave mantle imaging (Bodmer et al., 2018). Distinct low-velocity anomalies beneath northern and southern Cascadia are interpreted to reflect areas of sub-slab buoyancy that may influence plate coupling (Bodmer et al., 2018). In addition, we note correspondence between morphotectonic boundaries and those inferred based on changes in the timing (phase) and recurrence of episodic tremor and slow slip derived from GPS records (Brudzinski and Allen, 2007), the degree of plate locking derived from models of GPS data (Schmalzle et al., 2014), and the distribution of seismicity near the plate interface (Stone et al., 2018). In addition, the morphotectonic boundaries identified in this study are consistently located in the middle of basin-centered gravity lows (Wells et al., 2003) where slip has been suggested to be concentrated in megathrust earthquakes. While the spatial correlations described herein do not necessarily equate to causation, together they depict a high degree of along-strike variability in margin structure and physical processes occurring along the megathrust and in the adjacent plates that may reflect patterns of megathrust earthquake behavior over multiple seismic cycles.

Seismological data provide evidence of depth-dependent megathrust rupture behavior (Lay et al., 2012), while geological and geophysical studies in Cascadia (Brudzinski and Allen, 2007; Li and Liu, 2017; Wells et al., 2017) and around the globe (Sallarès and Ranero, 2019; Shi et al., 2020) have suggested a link between upper-plate terrane composition and changes in megathrust frictional behavior both along strike and downdip. Similarly, we speculate that the correspondence among our morphotectonic boundaries and previously inferred megathrust segmentation reflects the influence of along-strike variation in terrane composition and interface effective stress on the wedge taper and perhaps the frictional behavior of the megathrust. As described earlier, the morphotectonic boundaries identified here also generally correspond with along-strike variation in the composition and configuration of upper-plate terranes (Figs. 5–8). We equate this compositional heterogeneity to strength differences in the upper plate that directly

**Figure 9.** Plot comparing along-strike morphotectonic boundaries identified in this study with subduction zone segmentation boundaries (colored ovals) inferred from previous studies. See Figure 4 for explanation of vertical line style and color. The latitudinal location of each boundary on the figure represents the margin-normal linear extension of that segmentation boundary to the deformation front. Upper plate terranes are denoted by name and color (see Fig. 5 for map extents). Previously inferred along-strike segmentation boundaries are based on (from top to bottom) changes in oceanic asthenospheric velocities (Bodmer et al., 2018); episodic tremor and slip (ETS) recurrence and phase patterns derived from continuous GPS records (Brudzinski and Allen, 2007); the degree of plate locking derived from models of GPS data (Schmalzle et al., 2014); the distribution of near-plate-interface seismicity (Stone et al., 2018); areas of low moment release in three-dimensional elastic dislocation models of the A.D. 1700 earthquake (Wang et al., 2013); correlation and dating of marine turbidites and records of coastal subsidence from Goldberg et al. (2017) and Leonard et al. (2010), respectively; and basin-centered gravity lows from Wells et al. (2003). Note the consistent correspondence of segmentation boundaries at 46°N latitude, as well as the comparative mismatch (gray shaded region) where there are numerous oblique structures on the subducting plate.
impact the distribution of effective stress along the megathrust interface.

We acknowledge that this qualitative comparison may be limited by data gaps, interpretation uncertainty, and the complex structural obliquity of the upper and lower plates. For example, records of coseismic subsidence are lacking for much of the margin between 46°N and 49°N latitude, and marine turbidite records do not extend north of the Olympic Peninsula (48°N latitude), highlighting an important data gap across the boundary between regions 1 and 2. In addition, there are commonly large uncertainties associated with interpreting rupture extent from the correlation and dating of both offshore marine turbidites (Goldfinger et al., 2017) and coseismic subsidence events onshore (Shennan et al., 2016). The morphotectonic segmentation of the margin inferred from this study is broadly defined, in that reality these boundaries are likely geometrically complex involving interactions between structures below, above, and along the megathrust that are oriented obliquely to the deformation front. This may partially explain the mismatch in spatial alignment between the morphotectonic boundary between regions 3 and 4 and those inferred from previous work. As explained previously, there are several structural features on the lower plate that are subducting obliquely beneath regions 3 and 4 that may influence the morphotectonics of the wedge and skew the spatial alignment of inferred segmentation boundaries. From north to south, these features include a northeast-southwest-trending series of seamounts subducting beneath Heceta Bank (Figs. 2 and 5), a similarly oriented ~75-km-wide propagator wake (pw-4, Fig. 5), the WNW-ESE-trending Blanco fracture zone, and a northeast-southwest-trending propagator wake subducting beneath Rogue Canyon (pw-5, Fig. 5). We consider the boundaries identified here to be gradational, representing the long-term upper plate response to subduction processes.

Global Context

To understand the variability of outer wedge geometry in Cascadia in a global context, we compare calculations of outer wedge taper (α + β) in this study to a global compilation by Hu and Wang (2008) in Figure 9. Notably, the variability in outer wedge taper along the Cascadia margin is greater than at all other margins within the global catalog. This suggests either that Cascadia is an anomaly, or that observations of outer wedge taper in the global catalog do not accurately represent the range of heterogeneity within individual margins. While Cascadia is considered an end-member subduction zone in terms of its relative seismic quiescence, young and warm subducting plate, and thick sediment accumulation along the trench (Wang and Tréhu, 2016), recent margin-wide morphotectonic characterizations worldwide (e.g., Fujie et al., 2013; McNeill and Henstock, 2014; Saillard et al., 2017) suggest that this level of variability is likely more common than suggested in Figure 9. Rather than suggesting Cascadia is a global anomaly in terms of wedge morphotectonics, we argue that observations of outer wedge taper in the global catalog are not accurately representing the range of heterogeneity within individual margins.

To understand the links between Cascadia’s morphotectonic variability and its seismogenic behavior requires a systematic look at the spatial and temporal variability of the entire system. As with similar systematic studies of heterogeneity along plate-boundary faults (e.g., Ponce et al., 2010; Cubas et al., 2013; Johnson and Watt, 2012; Watt et al., 2014; McNeill and Henstock, 2014; Bassett et al., 2016; Brothers et al., 2020), detailed observations of the spatial and temporal variability within the system enable investigations of the links between fault-zone character and earthquake behavior. We suggest that Cascadia’s along-strike morphotectonic variation is a key feature of all subduction zones that may reflect differences in megathrust behavior and associated geologic hazards.

Tsunami Hazard Implications

The margin-wide characterization of fault-bounded backstop configurations and outer wedge vergence presented here (Figs. 4–7) highlights...
CONCLUSIONS

Tsunamis resulting from ruptures along splay faults (rather than distant sources) is dependent primarily on rupture details (Geist and Yoshioka, 1996; Wang and Tréhu, 2016) and the heterogeneity of crustal materials (Tung and Masterlark, 2018). The tsunami wave amplitude and waveform have the greatest influence on the tsunami runup at the coastline (Geist and Yoshioka, 1996). When megathrust earthquakes rupture to the seafloor in the overlying wedge, resulting in localized vertical deformation of the seafloor, the resulting tsunami wave amplitude is much greater than that for a buried rupture. Depending on the resulting waveform, these tsunamis can produce greater coastal runup as well. Tsunamis resulting from ruptures along splay faults or other prominent structures in the upper plate are particularly hazardous because the tsunami source is located closer to the coastline, resulting in significantly faster arrival times and limited time for coastal evacuation.

Clear evidence of either widespread trench-breaching or splay fault rupture in Cascadia is lacking. To date, only one of the dynamic backstops identified in this study has been studied in detail and considered as a potential tsunamigenic source (Priest et al., 2008; Witter et al., 2013; Gao et al., 2018). There is growing evidence from seismic-reflection imagery for the existence of a prominent active upper-plate fault system along an arcuate landward-dipping dynamic backstop extending through regions 1–3 (Figs. 5, 7B, 7D), coincident with the across-strike change from landward to seaward vergence (Mann and Snively, 1984; Goldfinger, 1994; Adam et al., 2004; Priest et al., 2009). Based on its length, apparent continuity, and mechanical role as a backstop, this fault is a likely candidate for a megathrust splay fault. This fault system has been incorporated into tsunami inundation scenarios developed for Canon Beach, Oregon (Priest et al., 2009), showing that inclusion of splay fault rupture amplifies tsunami water levels in open ocean by 6%–31% and coastal inundation by 2%–20%. To date, however, there is no definitive evidence that this fault ruptures in concert with the megathrust. This study highlights several other dynamic and static backstop faults that should be reexamined with modern multibeam bathymetry and high-resolution seismic-reflection imaging to examine active faulting along the distinct mid-slope breaks and backstops delineated in Figures 5–7.

Devastating historic tsunamis in Japan (Moore et al., 2007) and Alaska (Haeussler et al., 2015; von Huene et al., 2016) resulted from rupture along megasplay faults in the accretionary wedge. In each of these cases, the proposed tsunamigenic megasplay fault was associated with a distinct mid-slope break and/or defined a mechanical boundary, or fault-bounded backstop, separating older accretionary or country rocks from younger accretionary rocks. Variable backstop fault configurations and associated changes in slip mode and distance from the coastline (Fig. 5) suggest the tsunami hazard from megasplay may vary significantly along strike, requiring construction of tsunami source scenarios for the entire Cascadia margin.

Geophysical observations and mechanical modeling experiments in Cascadia and elsewhere suggest that landward vergence within the outer wedge may be more prone to shallow megathrust or trench-breaching rupture than regions characterized by seaward vergence. For example, suspected slip-to-the-trench in the 2004 Mw 9.2 Sumatra-Andaman earthquake is associated with landward-vertgent thrust faults observed in seismic-reflection imagery (Gulick et al., 2011). Mechanical analysis suggests that landward vergence in accretionary prisms is indicative of past seafloor frontal ruptures and tsunami genesis through dynamic weakening of the outer wedge and preferential activation of landward- versus seaward-vertgent structures (Cubas et al., 2016). In addition, recent work suggests that overconsolidation of the landward-vertgent portion of the outer wedge offshore Washington and northern Oregon is favorable for strain accumulation on the megathrust and rupture to the trench (Han et al., 2017). The discovery of a ~60-m-high fault scarp cutting an inferred Late Pleistocene erosional feature along a landward-vertgent frontal thrust (Beeson et al., 2017) provides additional geologic evidence of rupture to the trench along this portion of the margin. While the mechanisms of landward vergence and its relation to shallow megathrust rupture remain ambiguous, further investigation of a possible link between landward vergence within a wedge and trench-breaching megathrust rupture is needed. The work of Han et al. (2017) also highlights the need for better constraints on the plate boundary properties associated with shallow megathrust slip.

CONCLUSIONS

Globally observed correlations between marine forearc morphology and structure and megathrust earthquake slip, magnitude, and rupture length along subduction margins suggest possible correlations between seismic slip behavior and morphotectonic variability. Systematic margin-wide analysis of first-order morphotectonic variation along the Cascadia margin highlights four regions with distinct along-strike changes in outer wedge taper and structural vergence occurring at spatial scales (tens to hundreds of kilometers) roughly equivalent to rupture lengths of large earthquakes (≥M 6.7). Each region is characterized by unique seafloor morphology and wedge taper and displays distinct patterns of structural vergence. Interactions between upper-plate backstop configuration, glacial sediment flux, and subducting plate hydration, structure, and age all appear to play a role in controlling this variability.

Our margin-wide synthesis of backstop configurations and wedge geometry provides potential constraints on wedge strength, which may in turn lend insight into the slip behavior and tsunami potential offshore. This work delineates possible megasplay fault structures along backstop boundaries and documents areas of landward vergence, both of which represent regions theoretically associated with shallow megathrust rupture and amplification of tsunami waves (Cubas et al., 2016; Han et al., 2017). Detailed high-resolution imaging combined with targeted geologic sampling of these areas to understand recent deformation history will be key to improving future hazard assessments.
Geologically controlled variability in outer wedge morphology and structure is broadly associated with along-strike changes in oceanic asthenospheric velocities (Bodmer et al., 2018), episodic tremor and slip timing (Brudzinski and Allen, 2007), the degree of plate locking (Schmalzle et al., 2014), and the distribution of seismicity near the plate interface (Stone et al., 2018). We argue that the morphotectonic variability documented here reflects both geometric and rheological variations within the wedge, as well as frictional properties and fluid pressure on the plate interface. The along-strike morphotectonic boundaries identified here likely signify changes in the response of the upper plate to subduction processes over at least the last few million years that could influence the rupture process during the next major earthquake along the megathrust interface. Variability in outer wedge taper along the Cascadia margin compared with global compilations highlights the need for detailed margin-wide morphotectonic analyses of subduction zones worldwide.

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