Bulletin of the Seismological Society of America, Vol. 105, No. 2b, pp. –, May 2015, doi: 10.1785/0120140181

Tectonics and Structure of the Queen Charlotte Fault Zone, Haida Gwaii, and Large Thrust Earthquakes

by R. D. Hyndman*

Abstract This article provides a summary of the structure and tectonic history of the Queen Charlotte Transform fault zone off western Canada, as background to understanding the 2012 M_w 7.8 thrust earthquake off Haida Gwaii. There was margin subduction prior to the Eocene. The fault zone then became the mainly transcurrent Pacific–America boundary. There was mid-Tertiary oblique extension, then 15°–20° oblique convergence from ~6 Ma to the present that resulted in underthrusting and subduction initiation. The total underthrusting has been too small for Benioff–Wadati seismicity or arc volcanics but is indicated by (1) a trench, the Queen Charlotte trough, into which the oceanic plate bows downward and an offshore flexural bulge, the Oshawa rise; (2) the Queen Charlotte terrace, an accretionary sedimentary prism; (3) seismic receiver function delineation of the underthrusting Pacific plate; (4) heat flow decreasing landward as predicted for underthrusting; (5) low gravity offshore and high onshore, consistent with underthrusting; and (6) Late Tertiary uplift and erosion of the west coast of the islands. Oblique convergence is partitioned into nearly margin-normal underthrusting (i.e., M_w 7.8 event) relative to the terrace, which is moving along the margin, and margin-parallel on the Queen Charlotte strike-slip fault just off the coast that produced the 1949 M_s 8.1 earthquake. Landward on the mainland, Global Positioning System data suggest slow coast-parallel shear with no historical seismicity. The convergence rate decreases to the north of Haida Gwaii off Dixon Entrance, but large thrust earthquakes are possible. To the south, underthrusting of the Winona basin margin also could generate large earthquakes.

Introduction

This article provides a summary of the structure and tectonics of the region of the Queen Charlotte fault (QCF) zone, along the margin of western Canada and southernmost Alaska. It gives the context for the occurrence of the 2012 M_w 7.8 Haida Gwaii thrust earthquake beneath the Queen Charlotte terrace and other earthquakes of the region of Haida Gwaii. The QCF zone is the mainly transcurrent Pacific–America plate boundary western Canada in the region of Haida Gwaii and southernmost Alaska. It developed in the Eocene following the termination of margin subduction. The southern portion lies just off the west coast of the islands of Haida Gwaii (formerly Queen Charlotte Islands; Fig. 1) (e.g., Riddihough and Hyndman, 1989: Nishenko and Jacob, 1990; Hyndman and Hamilton, 1993; Rohr and Furlong, 1995; Scheidhauer, 1997; Mazzotti *et al.*, 2003; Barrie *et al.*, 2013; Walton *et al.*, 2013). Farther to the south, the Cascadia Subduction zone (CSZ) separates this transform fault zone from the transcurrent San Andreas fault zone. Between the QCF and CSZ there is a complex transition consisting of the Winona basin that may have slow margin convergence and the Dellwood Knolls and Tuzo Wilson short spreading centers with distributed deformation (e.g., Riddihough *et al.*, 1980; Davis and Riddihough, 1982; Riddihough, 1984; Carbotte *et al.*, 1989; Allan *et al.*, 1993; Rohr and Furlong, 1995; Braunmiller and Nábělek, 2002; Rohr and Tryon, 2010). To the north of Haida Gwaii the strike-slip QCF zone continues just off the coast of the Alaska Panhandle and then into the continent as the Fairweather fault.

The QCF zone is primarily transcurrent, but off Moresby Island, southern Haida Gwaii (Fig. 1), based on plate models there is current oblique convergence at 15°–20° from the margin trend (e.g., Engebretson *et al.*, 1985; Stock and Molnar, 1988; Hyndman and Hamilton, 1993; Doubrovine and Tarduno, 2008). It has been concluded that there is underthrusting beneath the islands that initiated about 6 my ago (e.g., Yorath and Hyndman, 1983), and it was recognized that there is the potential for large thrust earthquakes (e.g., Smith *et al.*, 2003; Bustin *et al.*, 2007) and their accompanying tsunamis (e.g., Leonard *et al.*, 2010). Convergence and underthrusting were confirmed by the shallow angle thrust M_w 7.8

*Also at School of Earth and Ocean Sciences, University of Victoria.

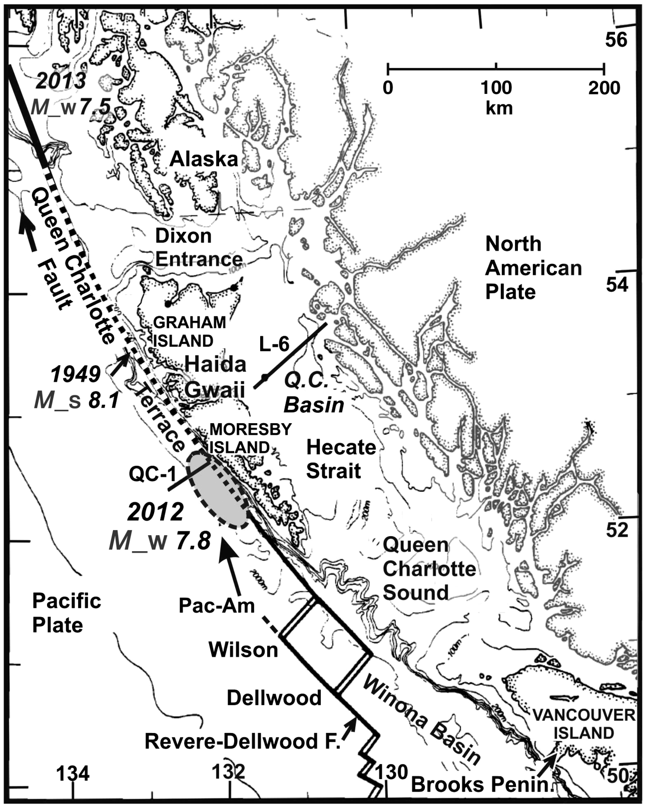

Figure 1. The Queen Charlotte fault (QCF) zone, the islands of Haida Gwaii and adjacent area, and the locations of the 2012 M_w 7.8 (ellipse), 2013 M_w 7.5 (solid line), and 1949 M_s 8.1 (dashed) earthquakes. The along margin extent of the 1949 event is not well constrained.

earthquake in 2012 off the west coast of Haida Gwaii (Szeliga, 2012; Hayes, 2013, see Data and Resources; James *et al.*, 2013; Lay *et al.*, 2013; Cassidy *et al.*, 2014), with rupture direction almost perpendicular to the coast. It generated a substantial tsunami with run-up exceeding 6 m at a number of sites (Leonard and Bednarski, 2015). The margin-parallel component of motion produced the M_s 8.1 strike-slip earthquake of 1949 just off the coast (Bostwick, 1984; Rogers, 1986; Nishenko and Jacob, 1990) and the 2013 M_w 7.5 strike-slip earthquake off the Alaska Panhandle (e.g., Lay *et al.*, 2013).

I start with a summary of the Tertiary tectonic history of the margin, including the evidence for post ~6 Ma underthrusting, then discuss how this underthrusting was initiated and developed to the current tectonics and structure that were responsible for the 2012 M_w 7.8 megathrust earthquake.

Tertiary Tectonic History and Geological Structure

The oldest exposed rocks on Haida Gwaii are the Permian to lower Middle Jurassic sedimentary and volcanic strata of the Wrangellia terrane (e.g., Sutherland Brown, 1968). The Wrangellia strata are allochthonous to northwestern North America, coming from the south, indicating that there was earlier transcurrent motion on this margin (e.g., Yorath and Chase, 1981; Irving *et al.*, 1996; Johnston, 2001, 2008). The whole western Canada margin is concluded to have been a subduction zone for some time prior to a major plate reorganization in the Eocene that resulted in the QCF zone (e.g., Engebretson *et al.*, 1985; Stock and Molnar, 1988; Atwater and Stock, 1998). Haeussler *et al.* (2003) inferred that there was another plate along this margin, the Resurrection plate, which is now completely subducted. This plate explains inferred northward terrane transport along the southeastern Alaska–British Columbia margin between 70 and 50 Ma (e.g., Pavlis *et al.*, 2004; Christeson *et al.*, 2010; McCrory and Wilson, 2013). Rusmore *et al.* (2010) described the 250 km wide region along the margin and showed evidence for a complex history that includes normal and dextral faulting since about 50 Ma. They concluded that the zone of active extension narrowed and migrated toward the margin transform through time, leading to formation of the Queen Charlotte basin about 20–30 Ma after the initiation of the transform margin.

Since about 42 Ma the Queen Charlotte margin has been the primarily right-lateral transform fault boundary between the Pacific and America plates. During this period, the Yakutat Terrane moved northward along the margin, perhaps from off Vancouver Island, to where it is now colliding with the continent in the Gulf of Alaska (Fig. 2) (e.g., Plafker *et al.*, 1994; Pavlis *et al.*, 2004, 2013; Koons *et al.*, 2010). The Yakutat Terrane may be part of an ocean plateau associated with the Crescent-Siletz terrane accreted to the Cascadia margin far to the south (e.g., Davis and Plafker, 1986; Pavlis *et al.*, 2004; Christeson *et al.*, 2010; McCrory and Wilson, 2013). In an alternative model, Perry *et al.* (2009) argued that this terrane was to the north of Haida Gwaii since ~40 Ma and may have moved farther northward recently. The recent phase of northerly Yakutat motion may have occurred after the change in plate motion at ~6 Ma that transferred the terrane from the America to Pacific plates. The Haida Gwaii margin appears to have been truncated during northward terrane motion, and some previous western portion may have been translated to Alaska on the Pacific plate along with other terranes (e.g., Johnston, 2001). During the Tertiary, the direction of Pacific–America relative motion off Haida Gwaii has varied by up to 20° with oblique extension in the mid-Tertiary and oblique convergence since about 6 Ma (e.g., Hyndman and Hamilton, 1993). Figure 3 shows the orthogonal component of the convergence vector compared to the current strike of the Haida Gwaii margin based on the plate models of Stock and Molnar (1988) and Engebretson *et al.* (1985). The latter is similar to the global plate circuit model of Doubrovine and Tarduno (2008).

There was extension in the Queen Charlotte basin, mainly in Hecate strait, that started in the early Tertiary (Rusmore *et al.*, 2010) but took place mainly in the mid-Tertiary (e.g., Rohr and Dietrich, 1992). The extension may have resulted from a period of 5°–10° very oblique extension relative to the current margin trend and caused the extensive Massett

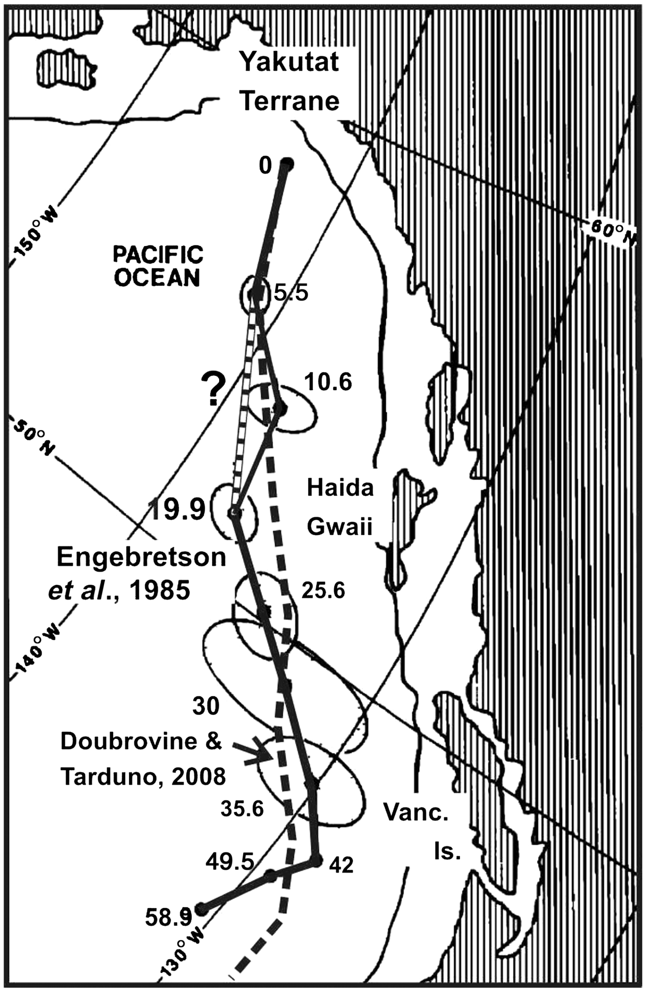

Figure 2. Models for the motion of the Yakutat terrane along the Queen Charlotte margin. The Pacific–America relative motion (a) after Stock and Molnar (1988) (solid line); the interval between 19.9 and 10.6 Ma is uncertain, and (b) from Doubrovine and Tarduno (2008) (dashed).

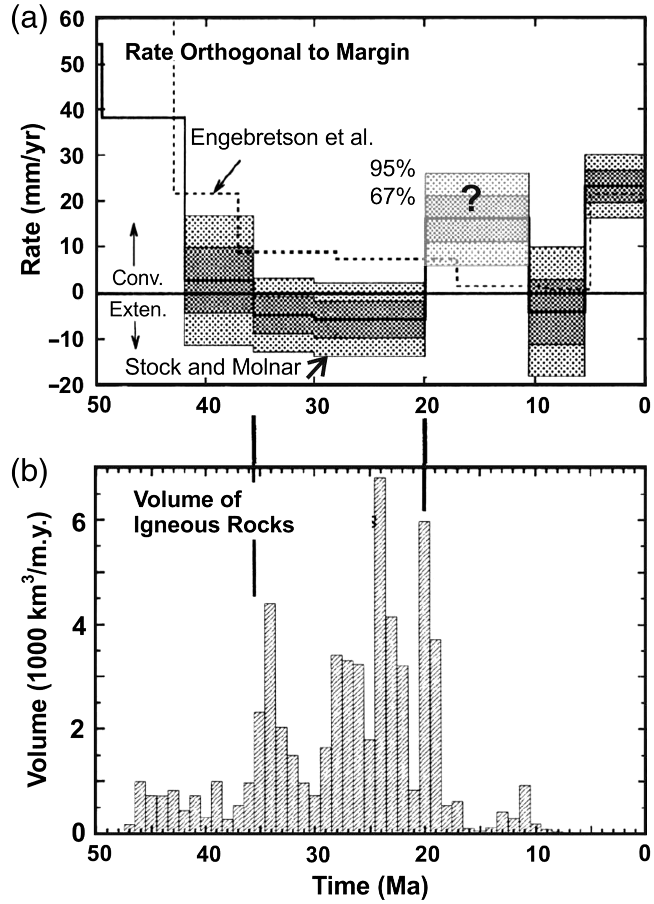

Figure 3. (a) Orthogonal component of Pacific motion across the margin of Haida Gwaii for the past 50 million years from the plate models of Stock and Molnar (1988) and Engebretson *et al.*, (1985; similar to Doubrovine and Tarduno, 2008). The latter models do not show the mid-Tertiary oblique extension, but this is the period of most Massett igneous volumes and of Queen Charlotte basin extension. (b) The volumes of mid-Tertiary Massett igneous rocks (after Hyndman and Hamilton, 1993).

volcanic and intrusive rocks found mainly on Haida Gwaii (e.g., Hickson, 1991; Hamilton and Dostal, 1993; Hyndman and Hamilton, 1993; Rusmore *et al.*, 2010) and crustal thinning in the Queen Charlotte basin (Fig. 4) (Spence and Asudeh, 1993; Lowe and Dehler, 1995). The syn-tectonic sediments associated with extensional subsidence were followed by the late Tertiary Skonun sediments of the thermally subsiding Queen Charlotte basin (e.g., Lewis *et al.*, 1991; Rohr and Dietrich, 1992; Hyndman and Hamilton, 1993; Dietrich *et al.*, 2009). Hyndman and Hamilton (1993) give a compilation of paleontology subsidence data, and Dehler *et al.* (1997) developed subsidence models following cessation of extension at ~20 Ma. The amount of extension is uncertain because the orientation of the margin relative to the Pacific–America plate motion at this time is not well determined in the plate models. As well, the Haida Gwaii Islands were moving seaward as a consequence of extension in the Queen Charlotte basin to the east, and may have rotated (e.g., Irving *et al.*, 1992, 2000).

To the north of Haida Gwaii prior to 6 Ma, the Pacific–America strike-slip boundary likely had a more westerly trend, seaward and 15°–20° oblique to the present northern Queen Charlotte–Fairweather fault. That more westerly trend fault was approximately parallel to the Pacific–America motion at that time, which was different from the present (Fig. 5). There is no evidence for the alternative of extensional faulting in the adjacent continent to accommodate the extensional component of relative plate motion. One interpretation is that the formation at 6 Ma of the Queen Charlotte–Fairweather fault in its current orientation transferred the Yakutat terrane from the North American to the Pacific plate, initiating the current phase of terrane collision. The pre-6 Ma Pacific–America transform fault may have become the Transition fault that currently marks the southern boundary of the Yakutat terrane. This fault may have rotated with the Yakutat terrane as a result of the collision, so it is no longer aligned with the pre-6 Ma Pacific–America relative motion. Ongoing

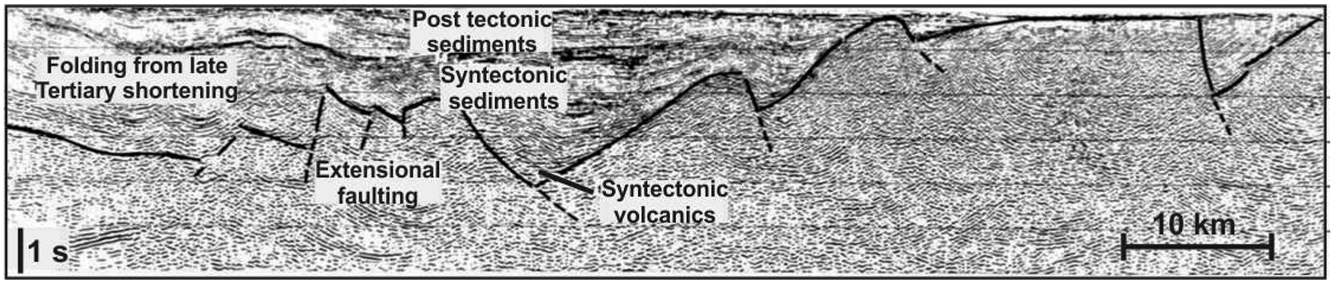

Figure 4. Multichannel seismic section across the Queen Charlotte basin to the east of Haida Gwaii, illustrating the substantial mid-Tertiary basement extensional normal faulting, and the small amount of subsequent shortening in the upper sediments (after Rohr and Dietrich, 1992). The location of the line, L-6, is shown in Figure 1.

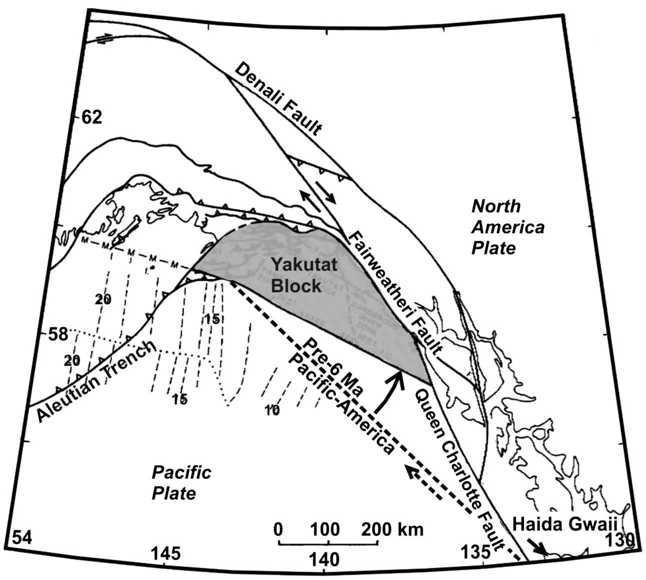

Figure 5. The pre-6 Ma Pacific–America relative motion and suggested rotation of the Yakutat terrane and its southern boundary, the Transition fault (modified from Bruns, 1983).

rotation of the Yakutat terrane is indicated by Global Positioning System (GPS) data (e.g., Fletcher and Freymueller, 1999).

Queen Charlotte Fault Zone and Terrace off Haida Gwaii

This section describes the margin structure off the Islands of Haida Gwaii where there is oblique convergence and gives the evidence that this convergence is accommodated by underthrusting starting at about 6 Ma. Immediately offshore of the islands of Haida Gwaii lies the Queen Charlotte terrace, an irregular terrace of deformed sediments at a water depth of about 1000 m (e.g., Chase *et al.*, 1975). It is interpreted to be a sedimentary prism made up primarily from sediments scraped off the underthrusting oceanic plate (e.g., Hyndman and Hamilton, 1993). It could contain slivers of volcanic rocks from the incoming crust, but as noted below it has the velocity structure of the Cascadia accretionary prism. The Queen Charlotte trough and the terrace extend irregularly from near the south end of the islands, to their north end (Fig. 6). There is some complexity off Graham Island where there is an incoming seamount just seaward of the terrace (Walton *et al.*, 2013), which is perhaps part of the Kodiak–Bowie seamount chain. The morphology of the Queen Charlotte terrace changes to the northwest from an irregular but generally linear terrace west of the Haida Gwaii islands to a broader zone of discontinuous blocks off Dixon Entrance (e.g., Scheidhauer, 1997). Farther to the north the margin trend turns gradually to be parallel to the relative plate direction such that there is no component of convergence and no accretionary sedimentary prism. To the south, the terrace terminates quite abruptly in the region of the Tuzo Wilson and Dellwood Knolls volcanic centers where new seafloor has been created in the last several million years (e.g., Riddihough *et al.*, 1980; Carbotte *et al.*, 1989; Rohr and Furlong, 1995).

Seismic reflection data show folding and shortening and landward thickening of the terrace sedimentary prism (Fig. 7), which is smaller but very similar to that of the Cascadia subduction zone to the south off Vancouver Island (e.g., Davis and Hyndman, 1989). The velocity–depth structure of the terrace is very similar to that of the Cascadia accretionary sedimentary prism off Vancouver Island (Smith *et al.*, 2003). The area of sediment in a cross section is approximately equivalent to the area formed by scraping off most of the incoming sediment section for the estimated convergence period of 6 my (Hyndman and Hamilton, 1993) so most of the incoming several kilometer thick sediment section appears to be scraped off and is incorporated into the Queen Charlotte terrace. The seismic reflection data do not resolve the subduction thrust well, but it appears to be at or near the base of the prism.

The terrace may be decoupled from the North American plate by the strike-slip QCF (e.g., Smith ~~et al.~~, 1999) and 2
hence moves along the margin at nearly the margin-parallel component of Pacific–America motion (e.g., Hyndman and Weichert, 1983). As a consequence, the Pacific plate underthrusts the terrace nearly orthogonally. As described by Scheidhauer (1997) and Walton *et al.* (2013), as the terrace migrates northwestward along the margin and reaches the eastward bend of the QCF west of Graham Island, the convergence component decreases and convergence and underthrusting follow pre-existing fractures and thrusts within the base of

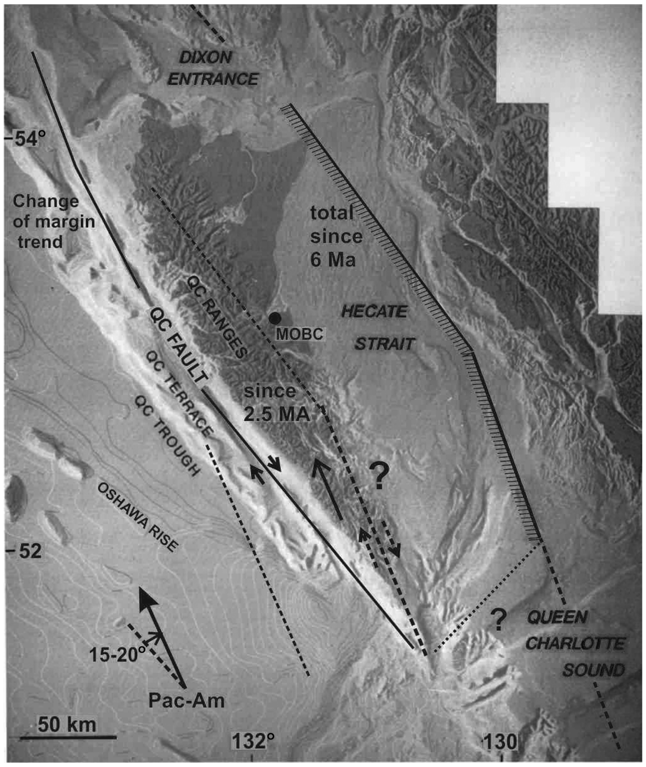

Figure 6. The Oshawa rise, Queen Charlotte trough (trench), Queen Charlotte terrace (accretionary sedimentary prism), and Queen Charlotte ranges (uplifted edge of continent). The dashed lines show the model extent of the underthrust plate for 2.5 and 6 million years, that is, for the triple junction at Brooks Peninsula (from 6 Ma) and at the Wilson Knolls (from 2.5 Ma). It is assumed that there has been no significant crustal shortening in these estimates.

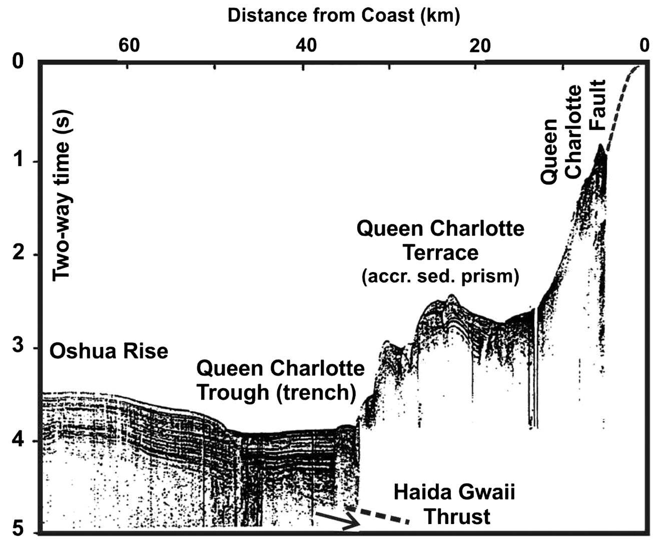

Figure 7. Seismic reflection profile across the Queen Charlotte trough and terrace in the area of the M_w 7.8 Haida Gwaii thrust earthquake (after Davis and Seemann, 1981). The location of this seismic line QC-1 is shown in Figure 1.

the terrace. A repetitive pattern of broken up terrace slivers results, and the outer edge of the terrace forms an angle of about 30° with the QCF, as expected by the transpressional strain ellipse for compressional structures. Compression and shear along faults and folds within these slivers control recent sediment dispersal as seen in ridges at the outer edges of the terrace blocks (Scheidhauer, 1997).

At the south end of the terrace the morphology is most subdued and Carbotte *et al.* (1989), Braunmiller and Nábělek (2002), and Rohr (2015) showed from geomorphology, seismicity, side-scan data, and seismic reflection profiles the complex transition zone to the Tuzo Wilson and Dellwood Knolls. Rohr (2015) points out that a northward extension of the Revere–Dellwood fault zone overlaps the southernmost part of the QCF zone in this region. The southern end of the QCF appears to have a series of offset steps. The northernmost end of the Revere–Dellwood fault zone extension is close to the southernmost extent of the 2012 M_w 7.8 thrust earthquake, and this may be the southern limit of the main underthrusting.

Margin Convergence and Underthrusting

Plate models show current 15°–20° oblique convergence on the Haida Gwaii margin that initiated in the Pliocene (Fig. 3) (Engebretson *et al.*, 1985; Stock and Molnar, 1988; DeMets and Dixon, 1999; Doubrovine and Tarduno, 2008; DeMets *et al.*, 2010; Austermann *et al.*, 2011). The convergence could be taken up either by shortening of the oceanic or continental plates (e.g., Rohr *et al.*, 2000) or by underthrusting. In support of at least some crustal shortening accommodation is the recent shortening evident in the Hecate strait Queen Charlotte basin sediment section (Rohr and Dietrich, 1992). However, the most recent sediments in the basin do not appear to be deformed, and most of the convergence appears to be accommodated by underthrusting of the oceanic lithosphere. As noted below, models for the initiation of subduction predict some crustal shortening at the initiation, so the continental crustal shortening may have been mainly at the time of initial development of underthrusting. Transpression initiated and the oceanic lithosphere was flexed downward at the plate boundary eventually reaching a steady state configuration of an underthrusting slab, marking the initiation of subduction.

The oblique convergence appears to be the result of a change in Pacific–America relative plate motion at about 6 Ma. One explanation for the altered Pacific motion is the initiation of west Pacific island arcs (e.g., Stern and Bloomer, 1992). The change in Pacific–America motion has been variously estimated as occurring between 3.9 and 8 Ma (e.g., Atwater and Stock, 1998) with recent analyses constraining the time to be close to 6 Ma (e.g., Austermann *et al.*, 2011), with an uncertainty of about 5–8 Ma. There undoubtedly was some duration to the shift in relative plate motion, and there may have been some delay after the plate motion change before the formation of significant margin deformation structures. A date of 6 Ma is used in the discussions below. The change in relative Pacific–America motion was probably responsible for

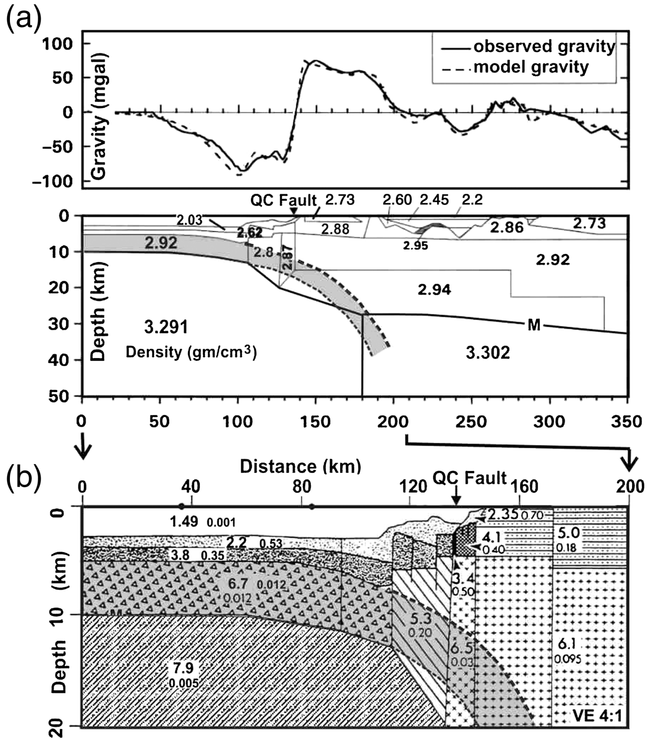

Figure 8. (a) Gravity and model density section across Haida Gwaii and adjacent regions, and (b) *P*-wave seismic velocity section and across the west coast margin of southern Haida Gwaii (adapted from Spence and Long, 1995). Density in grams per cubic centimeter and *P*-wave velocities in kilometers per second with gradient in kilometers per second per kilometer. The dashed lines are the down-dip extension of the top of the incoming oceanic crust to beneath the islands from the receiver function analyses.

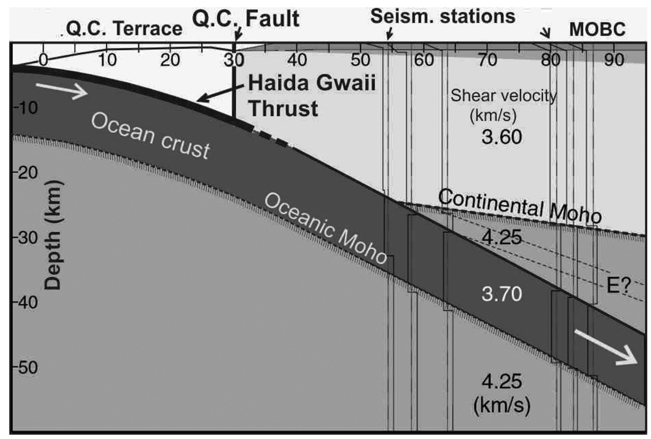

Figure 9. The underthrusting oceanic plate defined by receiver functions at Haida Gwaii stations (after Bustin *et al.*, 2007). See Gosselin *et al.* (2015), for an additional study.

many characteristics of late Tertiary deformation in western United States (e.g., Atwater and Stock, 1998; Argus and Gordon, 2013), extension in the Gulf of California (e.g., Oskin and Stock, 2003), and initiation of collision or increased convergence of the Yakutat terrane in southern Alaska (e.g., Enkelmann *et al.*, 2010; Gulick *et al.*, 2013; Pavlis *et al.*, 2013).

The current oblique convergence is concluded to be accommodated mainly by subduction-type underthrusting beneath Haida Gwaii, based on eight factors:

1. There is an offshore ocean plate forebulge, the Oshawa rise (e.g., Chase, 1975; Von Huene *et al.*, 1979) that is characteristic of subduction underthrusting. Offshore sedimentary horizons, which were likely deposited horizontally, dip landward toward the Queen Charlotte trough and the Queen Charlotte terrace (see Harris and Chapman, 1994, and Prims *et al.*, 1997, for flexure models).
2. The Queen Charlotte trough has the characteristics of a subduction trench into which the converging oceanic plate bows downward (e.g., Yorath and Hyndman, 1983; Hyndman and Hamilton, 1993; Harris and Chapman, 1994; Prims *et al.*, 1997).
3. The Queen Charlotte terrace appears to be a compressive sedimentary fold and thrust belt as characteristic of subduction zone accretionary sedimentary prisms (e.g., Hyndman and Hamilton, 1993; Spence and Long, 1995; Scheidhauer, 1997). Although there may be some crustal material incorporated, the velocity structure of the terrace is very similar to that of the Cascadia accretionary sedimentary prism off Vancouver Island (Smith, 1999; Smith *et al.*, 2003).
4. The gravity pattern across the margin is characteristic of subduction underthrusting, with a pronounced linear low over the Queen Charlotte trough and terrace and large gravity high over the west coast of the islands (Fig. 8) (Dehler *et al.*, 1988; Spence and Long, 1997). This gravity pattern cannot be explained by local isostasy and seems to require flexural support. Although not modeled with a dipping slab by Horn *et al.* (1984), Mackie *et al.* (1989), and Spence and Long (1997), the seismic structure across the margin also is consistent with underthrusting (Fig. 8).
5. The pattern of heat flow across the margin is consistent with underthrusting (Smith, 1999; Smith *et al.*, 2003; Wang *et al.*, 2015). The offshore high heat flow decreases landward under the terrace and the western islands where the subducting oceanic plate provides a cool heat sink (Hyndman *et al.*, 1982).
6. Receiver function seismic structure studies have delineated the landward-dipping boundaries that are interpreted to delineate the underthrust Pacific plate beneath the islands to a depth of over 40 km (Fig. 9) (Smith *et al.*, 2003; Bustin *et al.*, 2007; Gosselin *et al.*, 2015).
7. The GPS derived displacement vectors on the islands are oriented 10°–30° clockwise from the margin trend (Smith, 1999; Mazzotti *et al.*, 2003; Hippchen, 2011) (Fig. 10) indicating convergence and consistent with an underthrusting model.
8. The initiation of margin underthrusting appears to have uplifted the west coast of the islands in the Queen Charlotte ranges (Fig. 6). These mountains have a steep linear western side and slope gently to the east. The geological

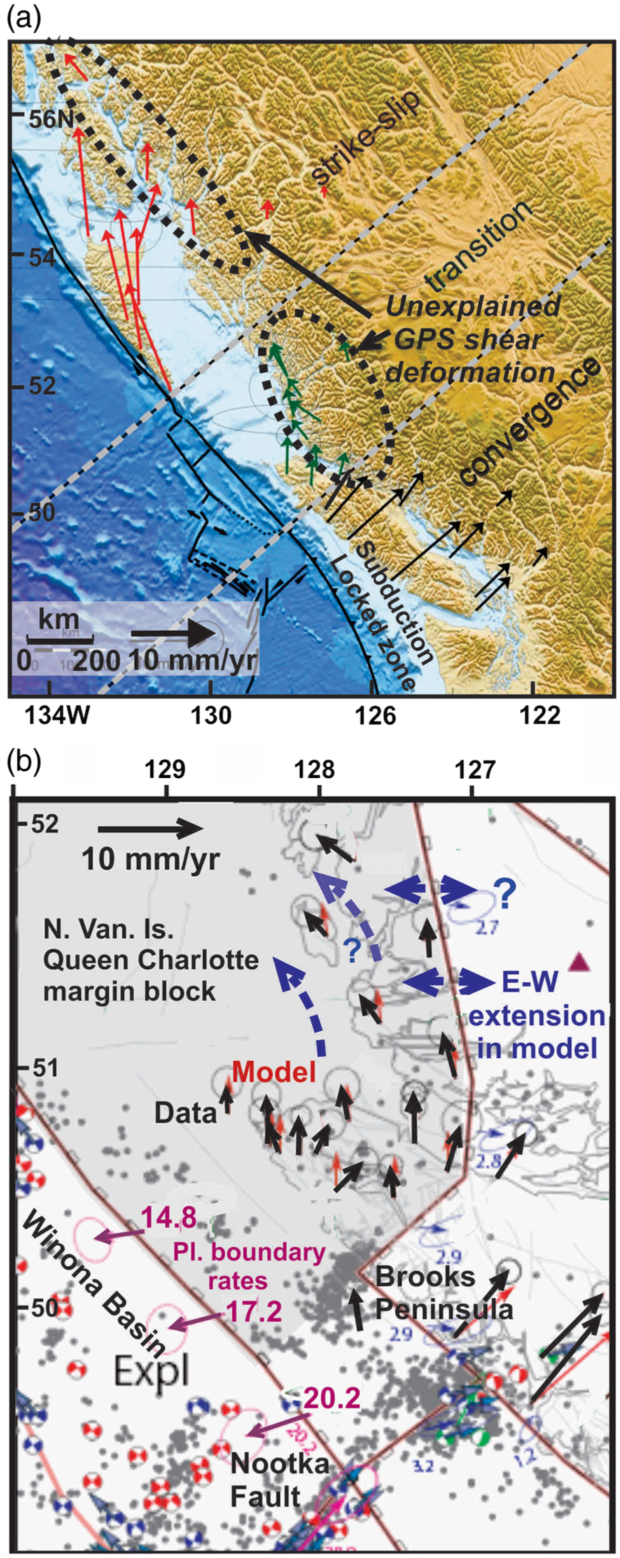

Figure 10. (a) Global Positioning System (GPS) motion vectors on Haida Gwaii and adjacent areas to the southeast prior to the 2012 thrust earthquake (summary after Hippchen, 2011). (b) GPS vectors and block model for the local area of northern Vancouver Island and adjacent mainland, showing northerly motion of the mainland block (adapted from McCaffrey *et al.*, 2007). Black arrows are measured vectors; lighter arrows are model vectors at the measured sites. The color version of this figure is available only in the electronic edition.

sections of Sutherland Brown (1968) for the Haida Gwaii Massett volcanics indicate gentle eastward tilting, suggesting uplift of the west side of the islands, assuming there is little depositional tilting (Hyndman and Hamilton, 1993). Also, erosion of the uplifted west coast of the islands has thinned the crust and exposed mid-Tertiary Massett igneous rocks that were likely emplaced at considerable depth.

Based on these constraints it is highly likely that there is ongoing underthrusting beneath the Queen Charlotte terrace and Haida Gwaii. Receiver function analysis at site MOBC on northeast Moresby Island (Bustin *et al.*, 2007; Gosselin *et al.*, 2015) (Fig. 9) delineates the underthrust oceanic plate extending at least 50 km inland. The predicted landward extent from relative plate motion for 6 my of underthrusting is shown in Figure 6. The model shown is for oblique underthrusting from 6 to 2.5 Ma that extended northwest from Brooks Peninsula on Vancouver Island (Fig. 1). After 2.5 Ma the underthrusting was concluded to take place only to the northwest of the Dellwood–Wilson spreading centers off Haida Gwaii, which is the location of the current complex plate triple junction. In this simple model the post 2.5 Ma underthrusting has carried the previous slab in the same direction (the “total since 6 Ma” on Fig. 6), but the post 2.5 Ma slab may have become detached from the older larger slab. This model does not allow for shortening deformation in the overlying continent or in the underthrust slab. Shortening is evident in the Queen Charlotte basin of Hecate strait after the extensional phase of deformation (e.g., Lewis *et al.*, 1991; Rohr and Dietrich, 1992; Rohr *et al.*, 2000). The motion of the down-dip edge of the underthrust plate beneath the islands may be responsible for the concentration of crustal seismicity in the northeastern Haida Gwaii (Bérubé, 1989; 5
Bird, 1997). Prims *et al.* (1997) estimate only 10–15 km of underthrusting along the northernmost Queen Charlotte Islands using 2D elastic modeling for northernmost Haida Gwaii where the predicted convergence is much smaller than to the south. At the southernmost end of Haida Gwaii, the oceanic lithosphere has been very young and weak since ~2.5 Ma, and the oceanic plate probably has been deforming rather than underthrusting the margin (southeast of dotted line on Fig. 6). This also is indicated by the lack of an uplifted ridge along the margin of Queen Charlotte sound south of the islands in contrast to that along the margin of Haida Gwaii to the north (Fig. 6).

Subduction Initiation

The margin of Haida Gwaii provides an example of subduction initiation from a prior transform fault ocean-continent margin (e.g., Gurnis *et al.*, 2004). The west coast of the islands appears to have been uplifted at ~6 Ma. Allowing for subsequent erosion, the elevation at that time must have been somewhat greater than the current average elevation of about 1000 m in the Queen Charlotte ranges (highest peak is Mount Moresby, 1164 m) (Fig. 6). The mid-Tertiary extension of the continental margin that is evident in the Queen Charlotte basin, Hecate strait, may have substantially heated and weakened the region of Haida Gwaii making underthrusting initiation possible with smaller than normal compression forces.

Toth and Gurnis (1998) discuss such initiation and the plate force required. They concluded that subduction initiation requires a weak zone such as a strike-slip fault as existed off Haida Gwaii (e.g., Gurnis *et al.*, 2004). Their models provide an explanation for the compression inferred in Queen Charlotte basin beneath Hecate strait at the start of underthrusting. Geodynamic models generally predict dynamic uplift during the first few million years after initiation of convergence (e.g., Mueller and Phillips, 1991; Shemenda, 1992; Toth and Gurnis, 1998), consistent with the inference of uplift of the west coast of Haida Gwaii. As a consequence of the underthrusting dynamic support, a strong positive gravity anomaly develops on the overriding plate as the crust is progressively uplifted, in agreement with the linear positive anomaly just inland of the coast of Haida Gwaii (Fig. 8) of +70 mGal regionally, rising from −60 offshore over the terrace and trough (+80 to −90 mGal locally; e.g., Spence and Long, 1995).

In a similar tectonic situation, House *et al.* (2002), Reyners and Webb (2002), and Lebrun *et al.* (2003) describe subduction initiation from a strike-slip boundary in Fiordland of South Island, New Zealand, that is very similar to that off Haida Gwaii. The important difference for a discussion of the forces involved in underthrusting beneath Haida Gwaii compared to many other areas is that the Haida Gwaii margin represents only a small section of the margin of the very large Pacific plate. Forces on the ~250 km long Haida Gwaii margin are unlikely to affect the total plate force balance of the large Pacific plate significantly. Also, the margin was heated and weakened by prior extension so less force was probably required to rupture and underthrust the margin than for older cold lithospheres. The change in Pacific–America plate relative motion at ~6 Ma is likely the result of a change in a larger boundary; such as the west Pacific island arc boundary reorganization suggested by Stern and Bloomer (1992).

The amount of uplift and erosion of the west coast of Haida Gwaii has not been estimated quantitatively through fission track, potassium argon, etc., but there are several qualitative indications. The first is the widespread mid-Tertiary Massett igneous rocks along the west coast Queen Charlotte ranges. Although locally these deep seated rocks may have been exposed by tectonic faulting, they likely came from significant depth and were mostly exposed by erosion. The second is the current crustal thickness at the west coast is approximately 24 km (see Fig. 8) compared to a regional crustal thickness of about 32 km, suggesting approximately 8 km erosional thinning. The third is by comparison with the analogous thrust initiation in southwest New Zealand Fiordland where there has been a detailed study. House *et al.* (2002) estimated about 7 km of exhumation resulting from initiation of underthrusting. A total of 5–10 km of erosion and exhumation resulting from underthrusting and subduction initiation is therefore reasonable for the west coast of Haida Gwaii (Fig. 11).

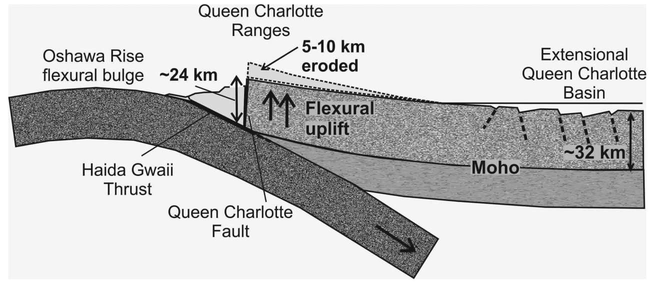

Figure 11. Model cross section of the Haida Gwaii margin illustrating uplifted and eroded west coast. The regional crustal thickness of ~32 km has been reduced to the current west coast thickness of ~24 km, suggesting about 8 km erosional thinning exposing Masset plutons. The initial thrust rupture may have been located approximately at the continental Moho.

Considering the evidence described for underthrusting, it is surprising that near to and parallel to the coast, the linear strike-slip trace of the QCF is clear in: (1) the seafloor bathymetry and morphology (Fig. 6). Recent very high-resolution bathymetry and acoustic imaging shows that this fault is nearly purely transcurrent at the seafloor (Fig. 12) (Barrie *et al.*, 2013); (2) the M_s 8.1 earthquake of 1949 that ruptured over 200 km from the central islands to the north off southern Alaska, had a dominantly strike-slip mechanism (Chandra, 1974; Rogers, 1986; Bérubé *et al.*, 1989; Nishenko *et al.*, 1990). Also, local seismicity located with temporary land and ocean-bottom seismograph stations was mainly beneath the surface trace of the QCF, most at a depth of 15–20 km (Hyndman and Ellis, 1981; Bérubé *et al.*, 1989; Bird, 1997) so the main fault must be nearly vertical. The models that fit the seismic structure and gravity data across the margin have assumed a near-vertical QCF although the dip is not well constrained (e.g., Spence and Long, 1995). I will refer to this fault that is close to and parallel to the Haida Gwaii margin as the QCF and the thrust at the base of the accretionary sedimentary prism as the Haida Gwaii thrust fault.

Very few of the mechanisms for the smaller earthquakes on the southern Haida Gwaii margin before the 2012 thrust earthquake were strike slip (Ristau *et al.*, 2007) so either there are only infrequent large strike-slip earthquakes with smaller local events of variable mechanism on or near the fault, or the main transcurrent fault in that area is creeping aseismically (e.g., Lay *et al.*, 2013). The large GPS vectors on the islands (Fig. 10) seems to preclude significant aseismic creep and the mechanisms of smaller earthquakes in the region are oblique to the margin, so we conclude that most of the motion on the main strike-slip fault occurs in large strike-slip earthquakes such as the 1949 event, rather than aseismically. Also, the statistics of the ongoing seismic moment rate from earthquakes on or near the QCF zone approximately equals that expected for the Pacific–America relative motion rate (Hyndman and Weichert, 1983).

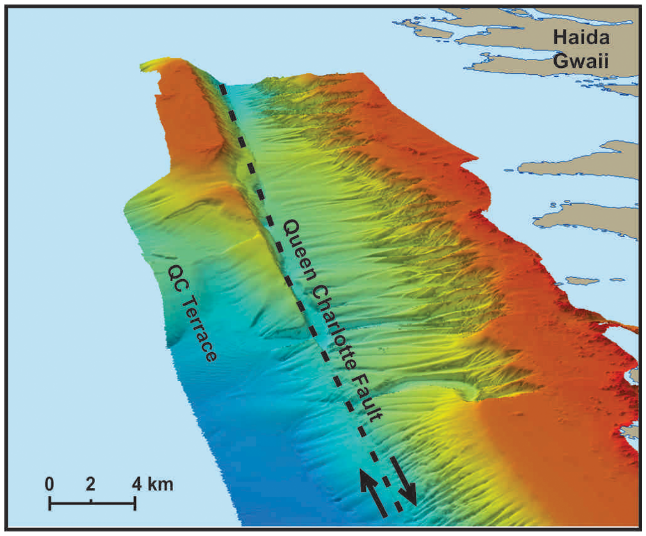

Figure 12. High-resolution seafloor bathymetry for a portion of the QCF near the thrust rupture area, showing the nearly linear trace of the fault. There is little sediment deformation, indicating nearly
12 pure strike-slip motion (after Barrie *et al.*, 2012). The color version of this figure is available only in the electronic edition.

This near-vertical strike-slip fault appears to define a fore-arc sliver that includes just the accreted sediments of the Queen Charlotte terrace (Yorath and Hyndman, 1983; Hyndman and Hamilton, 1993; Smith *et al.*, 2003). One of the alternative models suggested earlier (e.g., Smith *et al.*, 2003) was that this near-vertical fault cuts down through the underthrusting plate, but recent analyses including the 2012 large thrust event indicate that this model is unlikely (e.g., James *et al.*, 2013; Cassidy *et al.*, 2014). Fore-arc slivers and fore-arc strike-slip faulting above obliquely subducting plates have been extensively studied (e.g., Fitch, 1972; Beck, 1983; Jarrard, 1986; McCaffrey, 1992). Such strike-slip faulting commonly lies well landward of the coast and as far inland as
6 the volcanic arc, such as Sumatra (e.g., Fitch, 1986). Haida Gwaii is unusual because the main fore-arc sliver appears to include just the accretionary sedimentary prism. North of Haida Gwaii where the accretionary terrace disappears and there is no inferred underthrusting, the vertical fault must go from being above the subducting slab to a strike-slip plate boundary cutting through the whole ocean-continent lithosphere. This is a similar situation to New Zealand where there is strike-slip faulting (with some crustal shortening) in the Alpine fault zone in the northern part of the South Island that changes southwestward to the underthrusting of the Puysegur subduction zone (e.g., Lebrun *et al.*, 2003). A schematic for the tectonic regime of the Haida Gwaii margin is described in a later section.

Winona Basin

The Winona basin (Fig. 1) is a 150 km long and 50 km wide sedimentary basin lying between the Paul Revere ridge and the margin (Fig. 1). The seaward portion has formerly flat lying bedding that is now tilted landward (e.g., Davis and Riddihough, 1982; Rohr and Furlong, 1995), and the landward part of the basin shows shortening deformation. Prior to ~6 Ma the Pacific–America–Juan de Fuca/Explorer triple junction was located off Brooks Peninsula, Vancouver Island (Fig. 1), and strike-slip transform faulting extended south to that location. After 6 Ma the margin off Queen Charlotte sound to the south of Haida Gwaii was even more oblique to the Pacific–America relative motion than the Haida Gwaii margin so there must have been a significant convergence (e.g., Davis and Riddihough, 1982), at least ~20 mm/yr from, 6 Ma to about 2.5 Ma, that is, ~70 km. Seismic structure data shows an indication of a dipping reflector beneath southern Queen Charlotte sound (Yuan *et al.*, 1992) that may be the old underthrust slab that is now isolated by the ridge jump to the north.

At between 1.8 and 2.5 Ma the triple junction jumped to the current tectonically complex zone of the Dellwood and Tuzo Wilson Knolls just south of Haida Gwaii (e.g., Davis and Riddihough, 1982; Riddihough, 1984; Carbotte *et al.*, 1989; Rohr and Furlong, 1995; Braunmiller and Nábělek, 2002; Rohr and Tryon, 2010). Swath bathymetry and acoustic imaging show what may be an active volcanic seamount even farther north off the southernmost Haida Gwaii, so the northward migration of the triple junction may be continuing (Barrie *et al.*, 2013). The Dellwood and Wilson spreading centers have produced an area of very young oceanic crust at the northwest end of the Winona basin (e.g., Chase, 1977; Riddihough *et al.*, 1980). Between 1.8 and 2.5 Ma the Winona basin crustal block was separated from the Pacific plate by the formation of the Paul Revere ridge (e.g., Davis and Riddihough, 1982; Rohr, 2015). The ridge may have formed from rupture of an underthrusting flexural bulge similar to the situation of the Zenisu ridge off southwest Japan (Lallemant *et al.*, 1989; Malservisi *et al.*, 2003). Up to 8 km of turbidite sediments have accumulated in the basin with a landward increase in sediment thickness and sediment deformation that Davis and Riddihough (1982) interpreted to have resulted from scraping off on a margin thrust fault of the initially horizontal sediment section.

The current Revere–Dellwood fault zone (Fig. 1) is oblique to the Pacific–America vector so continuing Pacific–America convergence must be accommodated either by slow ongoing margin underthrusting or by local deformation (e.g., Davis and Riddihough, 1982; Rohr and Furlong, 1995; Rohr and Tryon, 2010; Rohr, 2015). The GPS vectors on the north end of Vancouver Island (summary by Hippchen, 2011; Fig. 10) are approximately in the direction of the Riddihough (1984) plate model motion, more northerly than for Explorer convergence. They have rates of 5–10 mm/yr decreasing to the northwest, rates that are about half the margin-normal rate of the Explorer plate to the south at about the same distance inland from the edge of the shelf. It is not clear if the GPS motions are associated with elastic strain buildup on a locked subduction thrust, by ongoing deformation, or by northeasterly motion of the adjacent continent. If there is a seismogenic thrust fault, this convergence could be accommodated by a $\sim M$ 7.5 earthquake about every 1000 years, that is about half the convergence rate of the Haida Gwaii margin (Leonard *et al.*, 2010).

Large Northward Moving Coastal Block

In addition to the narrow fore-arc sliver of the Queen Charlotte terrace that resulted from the Pacific–North America oblique convergence and that moves at nearly the Pacific–North America relative rate, there is some but not conclusive evidence for a band of the coastal mainland that is moving slowly to the northwest parallel to the margin. This may be another form of slip partitioning of the plate boundary. To define this larger scale mainland coastal motion using GPS data from the mainland region adjacent to Haida Gwaii, it is necessary to remove the interseismic elastic deformation effect of the locked Haida Gwaii thrust and the strike-slip QCF (Mazzotti *et al.*, 2003). Estimates of this GPS signal are model dependent, but for reasonable models the residual GPS indicates that the mainland coast is moving to the northeast at ∼5 mm/yr (Mazzotti *et al.*, 2003). Also, to the north of Haida Gwaii in the Alaska Panhandle there are GPS vectors on the mainland that seem too large for reasonable models of the locked Queen Charlotte strike-slip fault in that area (Mazzotti *et al.*, 2003). To the south in the region of Queen Charlotte sound, the Queen Charlotte strike-slip fault and the Haida Gwaii thrust are well to the north and should have minimal effect on the GPS vectors of northernmost Vancouver Island and adjacent mainland. These vectors therefore should reflect the few millimeters per year motion of this large coastal block (e.g., Hippchen, 2011) (Fig. 10). McCaffrey *et al.* (2007) concluded from analysis of GPS data on Vancouver Island and the adjacent mainland, that there is current extension in northern Vancouver Island north of Brooks Peninsula that may be accommodating the northwestward motion of this coastal block (Fig. 10). This northern portion of the island has been shown to have high heat flow, to have distinct crustal structure compared to the rest of Vancouver Island, and likely has had recent extension (Lewis *et al.*, 1997), which supports the conclusion of a slow northwesterly moving coastal block.

This mainland margin coastal zone has had very little historical seismicity and no currently active faults have been identified that could be accommodating this current margin-parallel motion. The one major margin-parallel structure that could be important is the very prominent Coast Shear zone, which is parallel to and about 100 km inland of the mainland coast, and ∼250 km inland of the QCF (Fig. 13). It is a major transcurrent fault structure that may have carried terranes to the northwest, but there is no indication of significant motion since the Eocene (e.g., Rusmore *et al.*, 2001). Also, this fault appears to have been deformed from a likely original nearly linear trend by mid-Tertiary extension (e.g., Rusmore *et al.*, 2010). Although no recent motion has been detected (summary by Hippchen, 2011), this is a difficult area in which to resolve active faulting, and no detailed study has been carried out specifically directed at recent motion.

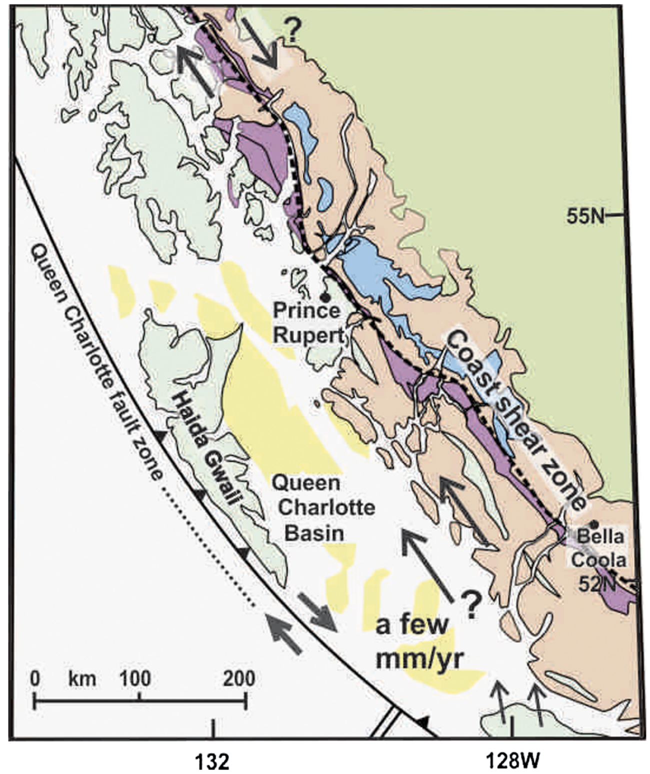

Figure 13. Slowly northwesterly moving mainland coastal block (modified after Rusmore *et al.*, 2010) inferred from GPS data. The prominent Coast Shear zone may define the landward boundary of the moving block, but no recent motion has been identified on it. The color version of this figure is available only in the electronic edition.

Oblique Convergence, Underthrusting, and Earthquakes on the Haida Gwaii Margin

The largest historical earthquake in Canada, the M_s 8.1 strike-slip earthquake in 1949, occurred on the QCF zone extending from mid-Haida Gwaii north to about the southern limit of the 2013 M_w 7.5 earthquake off the Alaska Panhandle, (Chandra, 1974; Rogers, 1986; Bérubé *et al.*, 1989; Nishenko *et al.*, 1990). This earthquake occurred before any local stations, and the rupture limits are poorly constrained by a limited number of aftershocks. The rupture could have extended through the 2013 rupture area. From Haida Gwaii north, most earthquakes in the catalog are in a narrow band with mechanisms indicating nearly pure strike-slip rupture on a nearly vertical fault, parallel to the Pacific–America motion in this region. The Craig, Alaska, M_w 7.5 event on 5 January 2013 at the southernmost end of the Alaska Panhandle was nearly pure strike slip (Hayes, 2013, see Data and Resources; Lay *et al.*, 2013). It occurred four months after the M_w 7.8 Haida Gwaii thrust event and may have been in response to the stress change from the Haida Gwaii earthquake. A Haida Gwaii tectonic model must reconcile the strike-slip margin parallel

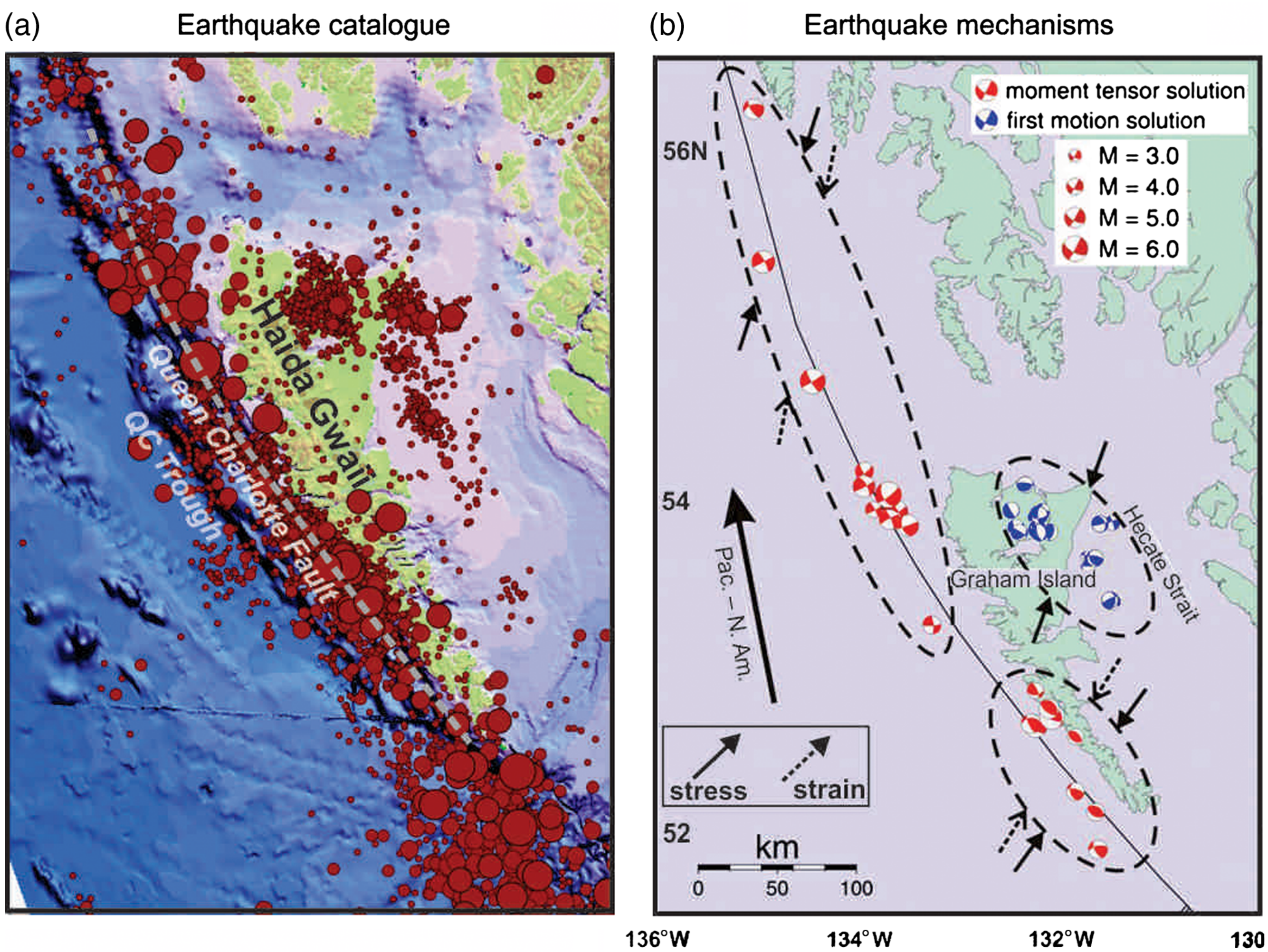

Figure 14. (a) Geological Survey of Canada catalog seismicity for the Haida Gwaii margin (after Bird, 1997). Only the recent events are well located. (b) Mechanisms for the larger events with composite stress and strain directions (after Ristau *et al.*, 2007). The color version of this figure is available only in the electronic edition.

earthquake off Haida Gwaii, and the evidence for underthrusting beneath the margin and the M_w 7.8 nearly margin-normal thrust event.

The smaller historical earthquakes along the southern Haida Gwaii margin are more dispersed and the mechanisms are complex, including many thrust events. The better located events using temporary local seismograph arrays including ocean-bottom seismographs were mainly beneath the surface trace of the QCF (Fig. 14; Hyndman and Ellis, 1981; Bérubé *et al.*, 1989; Bird, 1997). They appear to be concentrated in the area of the intersection of the subduction thrust and the near-vertical QCF near the coast. From moment tensor analyses, Ristau *et al.* (2007) showed that the maximum compressive stress for the earthquakes to the west of the islands is at a small angle to the Pacific–America relative motion (Fig. 14). The crustal earthquakes farther east under the northeast part of the islands have a similar maximum compressive stress direction oblique to the margin. Off southern Haida Gwaii, Ristau *et al.* (2007) and Kao *et al.* (2014) found very few strike-slip or shallow thrust mechanisms. Complex shortening faulting within the accretionary sedimentary prism near the margin and adjacent crustal rocks is suggested for this area as well as infrequent large shallow-dip thrust events at or near the base of the terrace accretionary prism.

Tectonic Origin of The 2012 Haida Gwaii Thrust Earthquake

The 2012 M_w 7.8 thrust earthquake has been studied in detail, including aftershock analyses using data from temporary land stations and ocean-bottom seismographs (Szeliga, 2012; James *et al.*, 2013; Lay *et al.*, 2013; and articles in this issue). I discuss only the characteristics that are important to understanding its tectonic origin. Most of the aftershocks of the 2012 thrust earthquake appear not to be on its rupture plane but rather to bound the rupture area (Fig. 15). Many of the aftershocks have extensional mechanisms (Kao *et al.*, 2015) and include few thrust events, both seaward and landward of the main rupture. The Coloumb stress change due to

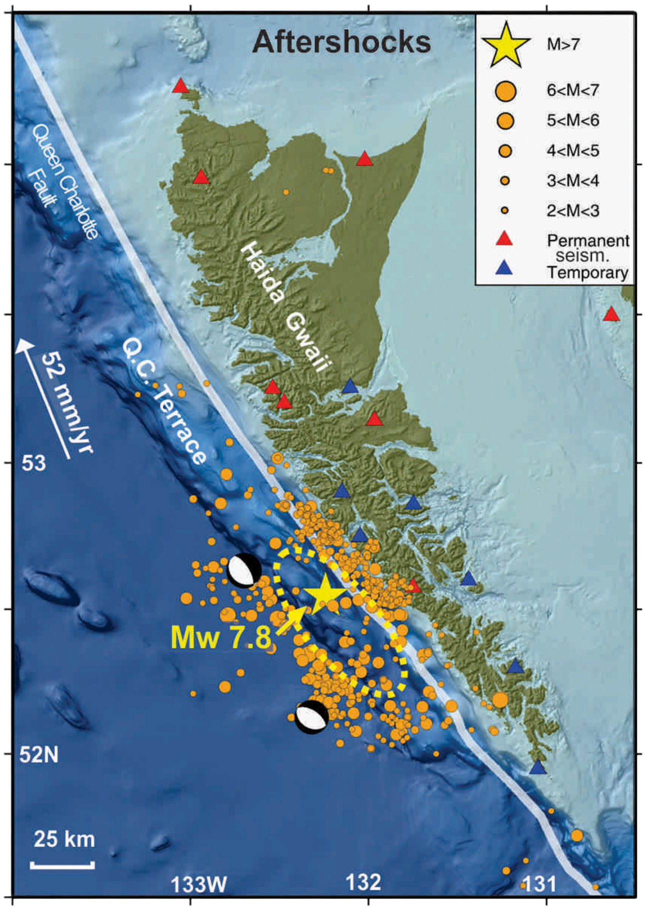

Figure 15. Aftershocks of the 2012 M_w 7.8 Haida Gwaii thrust
13 earthquake (after Cassidy *et al.*, 2013). They approximately define the rupture area. The normal-faulting mechanisms for two of the larger aftershocks are also shown. Many of the aftershocks are within the incoming oceanic plate and within the overriding continental plate rather than on the thrust rupture plane. The color version of this figure is available only in the electronic edition.

the large thrust earthquake is extension in both the offshore oceanic plate and in the onshore overriding continental plate, so these aftershock mechanisms are consistent with the transient stress change from the thrust earthquake.

The important aspect for understanding the tectonics of the area is the direction of thrust rupture, which is nearly but not quite orthogonal to the margin. A number of estimates have given similar rupture directions (e.g., Hayes, 2013, see Data and Resources; Lay *et al.*, 2012; Hobbs *et al.*, 2015; Kao *et al.*, 2015; Nykolaishen *et al.*, 2015), about 14° from orthogonal to the margin (Figs. 16 and 17). This is a little smaller but in general agreement with a number of other oblique convergence subduction zones (McCaffrey, 1992 and references therein).

For very oblique convergence, McCaffrey (1992) showed from fault stress analyses that, for increasingly oblique convergence directions, rupture angles relative to the margin should reach a limit. For a number of subduction zones, he found that this value ranges from 20° to 35° from orthogonal

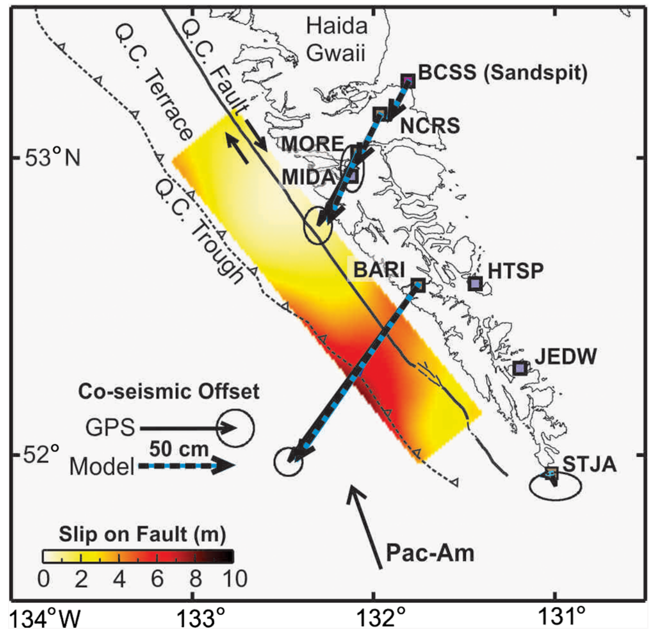

Figure 16. GPS-measured earthquake displacement vectors (black arrows) and modeled displacements (dashed arrows) on Haida Gwaii near the rupture area (after Nykolaishen *et al.*, 2015), showing rupture nearly orthogonal to the margin. The color version of this figure is available only in the electronic edition.

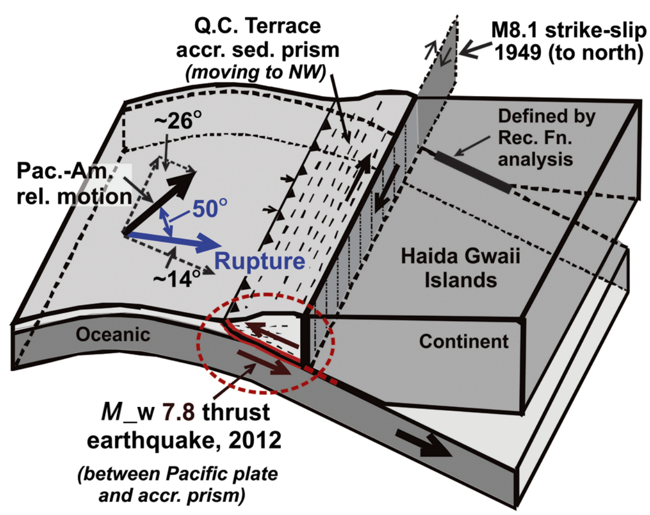

Figure 17. Model for the 2012 M_w 7.8 earthquake rupture and the partitioning of oblique convergence into margin parallel motion on the Queen Charlotte transcurrent fault and nearly orthogonal thrust convergence on the Haida Gwaii thrust fault. The color version of this figure is available only in the electronic edition.

to the margin strike. From minimum energy and force balance conditions the horizontal shear force on the two faults are about equal (Beck, 1983; McCaffrey, 1992). For Haida Gwaii the vertical extents of the thrust and strike-slip faults are similar, which is not commonly the case. The Haida Gwaii rupture angle of about 14° from orthogonal to the margin is slightly more orthogonal than the average for the subduction zones studied by McCaffrey (1992), which may mean that the QCF is somewhat weaker than the fore-arc sliver strike-slip faults that he studied. This may be because of its high temperatures or because the QCF is bounded by relatively weak sediments of the terrace accretionary prism.

One example with numerous similarities is the transition between the transcurrent Alpine fault and the Puysegur subduction zone where the Australian plate is obliquely subducted beneath southwestern New Zealand. A 60 km wide accretionary prism has formed between the Alpine fault and the trench (e.g., Barnes *et al.*, 2002; Reyners and Webb, 2002; Lebrun *et al.*, 2003). The oblique underthrusting generated the mainly oblique thrust M_w 7.6 Dusky sound earthquake of 2009. As discussed below (and Smith *et al.*, 2003; Wang *et al.*, 2015), the narrow width of the terrace sliver may be the result of the unusually high temperatures in the underthrust slab. The high temperatures limit the coupling to a shallow down-dip distance, about to the coast, which is approximately the rupture limit of the 2012 large thrust earthquake (e.g., James *et al.*, 2013; Lay *et al.*, 2013; Wang *et al.*, 2015). The Dusky sound earthquake penetrated substantially deeper because of the much older and colder incoming plate. Most of the rupture was down-dip landward of the transcurrent Alpine fault and where the relative motion was oblique thrust between the underthrust plate and the overlying continental plate. For the Dusky sound earthquake, only the uppermost part of the rupture seaward of the strike-slip fault, had rupture nearly orthogonal to the margin. In contrast, the Haida Gwaii thrust rupture was mainly between the underthrust plate and the terrace, which itself is moving along the margin. Therefore, for most of the rupture plane the thrust direction was nearly orthogonal to the terrace. Only the small most landward portion of the rupture beneath the coast had oblique rupture relative to the overlying fore-arc crust.

Thermal Limit to Down-Dip Rupture

Rupture models and the aftershock distribution indicate that this earthquake ruptured mainly offshore, extending not much farther landward than under the coast, with landward decreasing displacement. The reason for the shallow rupture down-dip limit is probably the young incoming plate age that results in very high thrust temperatures. The thermal seismogenic limit for subduction thrust faults is commonly about 350° C with a tapering displacement transition to about
15 450° C (e.g., Hyndman and Wang, 1993; Hyndman *et al.*, 1997). Finite-element thermal models for three 2D cross sections across the margin of Haida Gwaii were carried out by Smith *et al.* (2003), and an improved model in the area of the 2012 thrust earthquake is presented by Wang *et al.* (2015), using updated detailed structural, sediment deposition and plate age constraints. Wang *et al.* (2015) included the thermal effect of hydrothermal circulation in the incoming oceanic crust and a more detailed structural model. They found somewhat cooler temperatures on the thrust. The hydrothermal circulation was found necessary to explain the observed high heat flow in the ocean basin seaward of the terrace that was not modeled by Smith *et al.* (2003).

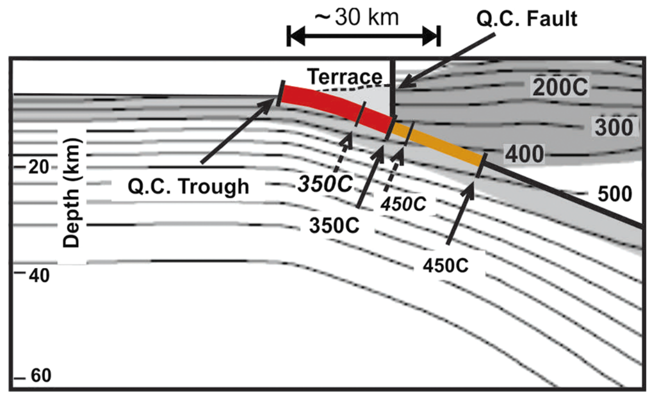

Figure 18. Model temperatures on the Haida Gwaii thrust in the area of the M_w 7.8 earthquake by Smith *et al.* (2003, dashed arrows) and Wang *et al.* (2015, solid arrows). The seismogenic thermal limits of 350° C and 450° C are near the coast in agreement with the limit of the 2012 M_w 7.8 rupture. Inclusion of the effect of hydrothermal cooling of the incoming oceanic crust in the Wang *et al.* model results in the critical temperatures extending deeper beneath the margin. The color version of this figure is available only in the electronic edition.

The critical 350° C temperature is reached near the coast (Fig. 18) in agreement with the down-dip limit of rupture in the 2012 earthquake. Smith *et al.* (2003) showed that a transcurrent margin structure with no underthrusting could give the general pattern of observed heat flow but with a poorer fit to their thrust model. It also is a poorer fit than the Wang *et al.* (2015) model. The limit of seismic behavior on the thrust being about at the coast may represent a transition between significant shear strength on the thrust to less shear strength at greater depth. This shallow transition in shear strength may be the reason the fore-arc sliver is limited by the Queen Charlotte transcurrent fault and includes only the accretionary sedimentary prism rather than a wider fore-arc crustal band as is common elsewhere.

Conclusions

At the continental margin of Haida Gwaii, the Pacific–North America relative motion was nearly strike slip from the Eocene to about 6 Ma with a period of very oblique mid-Tertiary extension that is associated with the extensive Massett volcanism and extensional subsidence in the Queen Charlotte basin. At 6 Ma the relative motion changed to the current orientation that is 15°–20° oblique to the margin. The oblique convergence is resolved into nearly margin-orthogonal underthrusting on the Haida Gwaii thrust fault beneath the Queen Charlotte terrace, which itself is moving along the margin to the northwest, and strike-slip motion just off the coast on the nearly vertical QCF that overlies the thrust. These two components respectively resulted in the 2012 M_w 7.8 thrust earthquake off Haida Gwaii, and the M_s 8.1 strike-slip earthquake of 1949 that extended from off Haida Gwaii to the

Alaska Panhandle. The 2013 M_w 7.5 strike-slip earthquake off the Alaska Panhandle may be the northward extension of the latter rupture. To the north of Haida Gwaii, the QCF trend changes to be nearly parallel to the Pacific–America relative motion and mainly strike-slip earthquakes are observed, including the 2013 event. Haida Gwaii convergence and underthrusting has not continued long enough to initiate arc volcanism or to develop a mature subducting slab to host Wadati–Benioff seismicity. In addition to the Queen Charlotte terrace fore-arc sliver, there also is a poorly resolved margin-parallel shear on the mainland that defines a very slowly northwestward moving larger block that could produce infrequent but large strike-slip earthquakes.

The initiation of convergence and underthrusting on the Haida Gwaii margin at about 6 Ma represents a form of subduction initiation. The start of underthrusting resulted in an offshore flexural bulge, downbowing of the incoming oceanic lithosphere at the margin, and uplift and erosion of the west coast of the islands. Substantial margin erosion followed, exposing mid-Tertiary Massett plutonic rocks. This recent underthrusting initiation is very similar to that for the SW New Zealand Puysegur subduction zone.

Underthrusting on the margin of the Winona basin to the south off northern Vancouver Island may also have initiated at ~6 Ma. The presence of a seismogenic thrust fault beneath this part of the margin is very uncertain, but up to about M_w 7.5 events are possible. Since ~2.5 Ma the Dellwood and Tuzo Wilson short spreading centers have developed off Queen Charlotte sound. They represent a very weak zone and this oceanic region likely is deforming rather than exhibiting ongoing underthrusting.

The present convergence and underthrusting beneath the Haida Gwaii margin is indicated by many consequences, including the M_w 7.8 thrust earthquake of 2012, which was similar to subduction zone megathrust earthquakes elsewhere. The evidence for subduction-type underthrusting include (1) an offshore ocean plate forebulge, the Oshawa rise; the Queen Charlotte trough trench; the Queen Charlotte terrace accretionary sedimentary prism; a low-high gravity pair across the margin characteristic of subduction; heat flow across the margin that is consistent with underthrusting; receiver function seismic structure studies that delineate the underthrust oceanic plate; and GPS vectors on the islands and mechanisms for small earthquakes that both indicate oblique convergence. Also, the initiation of margin underthrusting is concluded to have uplifted the west coast of the islands in the Queen Charlotte ranges.

The 2012 M_w 7.8 megathrust ruptured about half the length of the thrust margin. Rupture of the whole convergent margin could have a maximum magnitude of about M 8 and generate a substantial tsunami. If the convergence is accommodated mainly by characteristic M_w 7.8 events similar to that in 2012, they may occur with recurrence times of a few hundred years, assuming that the thrust is fully seismic with no creep accommodation of the plate convergence (e.g., Leonard *et al.*, 2014). Future large strike-slip earthquakes are expected on the near-vertical QCF near the coast of Haida Gwaii, similar to the 1949 event. Infrequent thrust events are possible to the north off Dixon Entrance where there is convergence but at a lower rate than off Haida Gwaii.

The 2012 thrust event will have reduced the normal stress on the strike-slip fault such that there may be a higher than average probability of a large strike-slip earthquake on the QCF occurring in the near future. Although located some 400 km to the northwest and occurring a few months later, the 2013 M_w 7.5 strike-slip Craig event off southernmost Alaska may have been triggered by stress changes produced by the M_w 7.8 thrust earthquake.

Data and Resources

Global Centroid Moment Tensor (CMT) (2014). The Global CMT Webpage, Global CMT Project database, http://www.globalcmt.org/ (last accessed June 2014). *Updated finite-fault results for the 28 October 2012 M_w 7.8 141 km south of Masset, Canada, Earthquake (Version 2)* by Hayes, G. (2013) were obtained at *U.S. Geological Survey Earthquakes* website: http://comcat.cr.usgs.gov/earthquakes/eventpage/pde20121028030408820_14#scientific_moment-tensor (last accessed June 2014). All other data used in this article came from published sources listed in the references.

Acknowledgments

The author wishes to thank the many collaborators and colleagues who have contributed to the science and to the information included in this article. Reviewer Terry Pavlis and two anonymous reviewers made very helpful comments and revisions. Guest editor Tom James is especially recognized for his detailed comments, revisions, and editing, which resulted in a much improved article.

References

Allan, J. F., R. L. Chase, B. Cousens, P. J. Michael, M. P. Gorton, and S. D. Scott (1993). The Tuzo Wilson Volcanic field, NE Pacific: Alkaline volcanism at a complex, diffuse, transform-trench-ridge triple junction, *J. Geophys. Res.* **98,** 22,367–22.387.

Argus, D. F., and R. G. Gordon (2013). Present tectonic motion across the Coast Ranges and San Andreas fault system in central California, *Geol. Soc. Am. Bull.* **113,** 1580–1592, doi: 10.1130/0016-7606(2001)113<1580:PTMATC>2.0.CO;2.

Atwater, T., and J. Stock (1998). Pacific-North America plate tectonics of the Neogene southwestern United States: An update, *Int. Geol. Rev.* **40,** 375–402.

Austermann, J., Z. Ben-Avraham, P. Bird, O. Heidbach, G. Schubert, and J. M. Stock (2011). Quantifying the forces needed for the rapid change of Pacific plate motion at 6 Ma, *Earth Planet. Sci. Lett.* **307,** 289–297, doi: 10.1016/j.epsl.2011.04.043.

Barnes, P., B. Davy, R. Sutherland, and J. Delteil (2002). Frontal accretion and thrust wedge evolution under very oblique plate convergence: Fiordland basin, New Zealand, *Basin Res.* 14**,** 439–466.

Barrie, J. V., K. W. Conway, and P. T. Harris (2013). The Queen Charlotte Fault, British Columbia: Seafloor anatomy of a transform fault and its influence on sediment processes, *Geo. Mar. Lett.* **33,** 311–318.

Beck, M. E. (1983). On the mechanism of tectonic transport in zones of oblique subduction, *Tectonophysics* **93,** 1–11.

Bérubé, J., G. C. Rogers, R. M. Ellis, and F. O. Hasselgren (1989). A microseismicity survey of the Queen Charlotte Islands region, *Can. J. Earth Sci.* **26,** 2556–2566.

Bird, A. L. (1997). Earthquakes in the Queen Charlotte Islands region: 1982–1996, *M.Sc. Thesis*, University of Victoria, Victoria, British Columbia 123 p.

9 Bostwick, T. K. (1984). A re-examination of the August 22, 1949 Queen Charlotte earthquake, *M.Sc. Thesis*, University of British Columbia, Vancouver, British Columbia, available at http://hdl.handle.net/2429/24553.

Braunmiller, J., and J. Nábelek (2002). Seismotectonics of the Explorer region, *J. Geophys. Res.* **107,** doi: 10.1029/2001JB000220.

Bruns, T. R. (1983). Model for the origin of the Yakutat block, an accreting terrane in the northern Gulf of Alaska, *Geology* **11,** 718–721.

Bustin, A. M. M., R. D. Hyndman, H. Kao, and J. F. Cassidy (2007). Evidence for underthrusting beneath the Queen Charlotte Margin, British Columbia, from teleseismic receiver function analysis, *Geophys. J. Int.* **171,** 1198–1211.

Carbotte, S. M., J. M. Dixon, E. Farrar, E. E. Davis, and R. P. Riddihough (1989). Geological and geophysical characteristics of the Tuzo Wilson Seamounts: Implications for plate geometry in the vicinity of the Pacific-North America-Explorer triple junction, *Can. J. Earth Sci.* **18,** 2365–2384.

Cassidy, J., G. C. Rogers, and R. D. Hyndman (2014). An Overview of the 28 October 2012 M_w 7.7 Earthquake in Haida Gwaii, Canada: A tsunamigenic thrust event along a predominantly strike-slip margin, *Pure Appl. Geophys.* doi: 10.1007/s00024-014-0775-1.

17 ~~Chamot-Rooke, N., and X. LePichon (1989). Zenisu Ridge: Mechanical model of formation, *Tectonophysics* **160,** 175–193.~~

Chandra, U. (1974). Seismicity, earthquake mechanisms, and tectonics along the western coast of North America, from 42° to 61° N, *Bull. Seismol. Soc. Am.* **64,** 1529–1549.

Chase, R. L. (1977). J. Tuzo Wilson Knolls: Canadian hotspot, *Nature* **266,** 344–346.

Chase, R. L., D. L. Tiffin, and J. W. Murray (1975). The western Canadian continental margin, in *Canada's Continental Margins and Offshore Petroleum Exploration*, AAPG Memoir, Vol. **4,** 701–721.

Christeson, G. L., S. P. Gulick, H. J. van Avendonk, L. L. Worthington, R. S. Reece, and T. L. Pavlis (2010). The Yakutat terrane: Dramatic change in crustal thickness across the Transition fault, Alaska, *Geology* **38,** 895–898.

Davis, A. S., and G. Plafker (1986). Eocene basalts from the Yakutat terrane: Evidence for the origin of an accreting terrane in southern Alaska, *Geology* **14,** 963–966.

18 ~~Davis, E. E., and R. M. Clowes (1986). High velocities and seismic anisotropy in Pleistocene turbidites off western Canada, *Geophys. J. Roy. Astron. Soc.* **84,** 381–399.~~

Davis, E. E., and R. D. Hyndman (1989). Accretion and recent deformation of sediments along the northern Cascadia subduction zone, *Geol. Soc. Am. Bull.* **101,** 1465–1480.

Davis, E. E., and R. P. Riddihough (1982). The Winona Basin: Structure and tectonics, *Can. J. Earth Sci.* **19,** 767–788.

Davis, E. E., and D. A. Seemann (1981). Compilation of seismic reflection profiles across the continental margin of western Canada, *Geol. Surv. Can. Open-File Rept. 751*, Ottawa, Canada.

Dehler, S. A., and R. M. Clowes (1988). The Queen Charlotte Islands refraction project project. Part 1. The Queen Charlotte Fault Zone, *Can. J. Earth Sci.* **25,** 1857–1870.

Dehler, S. A., C. E. Keen, and K. M. M. Rohr (1997). Tectonic and thermal evolution of Queen Charlotte Basin: Lithospheric deformation and subsidence models, *Basin Res.* **9,** 243–261, doi: 10.1046/j.1365-2117.1997.00043.x.

DeMets, C., and T. H. Dixon (1999). New kinematic models for Pacific-North America motion from 3 Ma to present, I: Evidence for steady motion and biases in the NUVEL-1A Model, *Geophys. Res. Lett.* **26,** 1921–1924, doi: 10.1029/1999GL900405.

DeMets, C., R. G. Gordon, and D. F. Argus (2010). Geologically current plate motions, *Geophys. J. Int.* **181,** 1–80, doi: 10.1111/j.1365-246X.2009.04491.x.

Dietrich, J. R., R. Higgs, K. M. Rohr, and J. M. White (2009). The Tertiary Queen Charlotte Basin: A strike-slip basin on the western Canadian Continental margin, in *Tectonic Controls and Signatures in Sedimentary Successions: Special Publication 20 of the Inter. Assoc, Sed.* Vol. **40,** 161–169.

Doubrovine, P. V., and J. A. Tarduno (2008). A revised kinematic model for the relative motion between Pacific oceanic plates and North America since the Late Cretaceous, *J. Geophys. Res.* **113,** doi: 10.1029/2008JB005585.

Engebretson, D. C., A. Cox, and G. A. Thompson (1985). Relative motions between oceanic and continental plates in the Pacific basin, *Spec. Pap. Geol. Soc. Am.* **206,** 59 pp.

Enkelmann, E., P. K. Zeitler, J. I. Garver, T. L. Pavlis, and B. P. Hooks (2010). The thermochronological record of tectonic and surface process interaction at the Yakutat-North American collision zone in southeast Alaska, *Am. J. Sci.* **310,** 231–260, doi: 10.2475/04.2010.01.

Evenchick, C. A., M. L. Crawford, V. J. McNicoll, L. D. Currie, and P. B. 10
O'Sullivan (1999). Early Miocene and younger normal faults and other Tertiary structures in west Nass River map area, northwest British Columbia, and adjacent parts of Alaska, *Curr. Res.* 1999-A, 1–11.

Fitch, T. J. (1972). Plate convergence, transcurrent faults and internal deformation adjacent to southeast Asia and the western Pacific, *J. Geophys. Res.* **77,** 4432–4460.

Fletcher, H. J., and J. T. Freymueller (1999). New GPS Constraints on the Motion of the Yakutat Block, *Geophys. Res. Lett.* **26,** 3029–3032.

Gosselin, J. M., J. F. Cassidy, and S. E. Dosso (2015). Shear wave velocity structure in the vicinity of the 2012 M_w 7.8 Haida Gwaii earthquake from receiver function inversions, *Bull. Seismol. Soc.* **105,** no. 2B, doi: 10.1785/0120140171.

Gulick, S. P. S., R. S. Reece, G. L. Christeson, H. van Avendonk, L. L. Worthington, and T. L. Pavlis (2013). Seismic images of the Transition fault and the unstable Yakutat–Pacific–North American triple junction, *Geology* **41,** 571–574, doi: 10.1130/G33900.1.

Gurnis, M., C. Hall, and L. Lavier (2004). Evolving force balance during incipient subduction, *Geochem. Geophys. Geosys.* **5,** doi: 10.1029/2003GC000681.

Haeussler, P. J., D. C. Bradley, R. E. Wells, and M. L. Miller (2003). Life and death of the Resurrection plate: Evidence for its existence and subduction in the northeastern Pacific in Paleocene-Eocene time, *Geol. Soc. Am. Bull.* **115,** 867–880.

Hamilton, T. S., and J. Dostal (1993). Geology, geochemistry and petrogenesis of middle Tertiary volcanic rocks of the Queen Charlotte Islands, British Columbia (Canada), *J. Volcanol. Geoth. Res.* **59,** 77–99.

Harris, R. N., and D. S. Chapman (1994). A comparison of mechanical thickness estimates from trough and seamount loading in the southeastern Gulf of Alaska, *J. Geophys. Res.* **99,** 9297–9317.

Hickson, C. J. (1991). The Masset Formation on Graham Island, Queen Charlotte Islands, BC, in *Evolution and Hydrocarbon Potential of the Queen Charlotte Basin, BC*, G. J. Woodsworth (Editor), Geol. Surv. Can. Paper 90-10.

Hippchen, S. (2011). Slip partitioning, crustal tectonics and deformation of the Queen Charlotte margin and northern Vancouver Island, *Ph.D. Thesis*, School of Earth and Ocean Sciences, University of Victoria, Victoria, B.C., 191 pp.

Hobbs, T. E., J. F. Cassidy, S. E. Dosso, and C. Brillon (2015). Coulomb stress changes following the M_w 7.8 2012 Haida Gwaii, Canada earthquake: Implications for seismic hazard, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140158.

Horn, J. R., R. M. Clowes, R. M. Ellis, and D. N. Bird (1984). The seismic structure across an active oceanic/continental transform fault zone, *J. Geophys. Res.* **89,** 3107–3120.

House, M. A., M. Gurnis, P. J. J. Kamp, and R. Sutherland (2002). Uplift in the Fiordland region, New Zealand: Implications for incipient subduction, *Science* **297,** doi: 10.1126/science.1075328.

Hyndman, R. D., and R. M. Ellis (1981). Queen Charlotte fault zone: Microearthquakes from a temporary array of land stations and ocean bottom seismographs, *Can. J. Earth Sci.* **18,** 776–788.

Hyndman, R. D., and T. S. Hamilton (1993). Queen Charlotte area Cenozoic tectonics and volcanism and their association with relative plate motions along the northeastern Pacific margin, *J. Geophys. Res.* **98,** 14,257–14,277.

Hyndman, R. D., and D. H. Weichert (1983). Seismicity and rates of relative motion on the plate boundaries of western North America, *Geophys. J. Int.* **72,** 59–82.

Hyndman, R. D., T. J. Lewis, J. A. Wright, M. Burgess, D. S. Chapman, and M. Yamano (1982). Queen Charlotte fault zone: Heat flow measurements, *Can. J. Earth Sci.* **19,** 1657–1669.

Hyndman, R. D., M. Yamano, and D. A. Oleskevich (1997). The seismogenic zone of subduction thrust faults, *Island Arc* **6,** 244–260.

Irving, E., J. Baker, P. J. Wynne, T. S. Hamilton, and M. T. D. Wingate (2000). Evolution of the Queen Charlotte Basin; further paleomagnetic evidence of Tertiary extension and tilting, *Tectonophysics* **326,** 1–22.

Irving, E., J. G. Souther, and J. Baker (1992). Tertiary extension and tilting in the Queen Charlotte Islands, evidence from dyke swarms and their paleomagnetism, *Can. J. Earth Sci.* **29,** 1878–1898.

Irving, E., P. J. Wynne, D. J. Thorkelson, and P. Schiarizza (1996). Large (1000 to 4000 km) northward movements of tectonic domains in the northern Cordillera, 83 to 45 Ma, *J. Geophys. Res.* **101,** 17,901–17,916.

James, T., G. Rogers, J. Cassidy, H. Dragert, R. Hyndman, L. Leonard, L. Nykolaishen, M. Riedel, M. Schmidt, and K. Wang (2013). Field studies target 2012 Haida Gwaii earthquake, *Eos Trans. AGU* **94,** 197–198, doi: 10.1002/2013EO220002.

Jarrard, R. D. (1986). Terrane motion by strike-slip faulting of forearc slivers, *Geology* **14,** 780–783.

Johnston, S. T. (2001). The Great Alaskan Terrane Wreck: Reconciliation of paleomagnetic and geological data in the northern Cordillera, *Earth Planet. Sci. Lett.* **193,** 259–272.

Johnston, S. T. (2008). The Cordilleran ribbon continent of North America, *Ann. Rev. Earth Planet. Sci.* **36,** 495–530.

Kao, H., S.-J. Shan, A. Bent, C. Woodgold, G. Rogers, J. F. Cassidy, and J. Ristau (2014). Regional Centroid-Moment-Tensor analysis for earthquakes in Canada and adjacent regions: An update, *Seismol. Res. Lett.* **83,** 505–515, doi: 10.1785/gssrl.83.3.505.

Kao, H., S.-J. Shan, and A. M. Farahbod (2015). Source characteristics of the 2012 Haida Gwaii earthquake sequence, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140165.

Koons, P. O., B. P. Hooks, T. Pavlis, P. Upton, and A. D. Barker (2010). Three-dimensional mechanics of Yakutat convergence in the southern Alaskan plate corner, *Tectonics* **29,** doi: 10.1029/2009TC002463.

Lallemant, S., N. Chamot-Rooke, X. Le Pichon, and C. Rangin (1989). Zenisu Ridge: A deep intraoceanic thrust related to subduction, off southwest Japan, *Tectonophysics* **160,** 151–174, doi: 10.1016/0040-1951(89)90389-2.

Lay, T., L. Ye, H. Kanamori, Y. Yamazaki, K. F. Cheung, K. Kwong, and K. D. Koper (2013). The October 28, 2012 M_w 7.8 Haida Gwaii underthrusting earthquake and tsunami: Slip partitioning along the Queen Charlotte Fault transpressional boundary, *Earth Planet. Sci. Lett.* **375,** 57–70.

Lebrun, J.-F., G. Lamarche, and J.-Y. Collot (2003). Subduction initiation at a strike-slip plate boundary: The Cenozoic Pacific-Australian plate boundary, south of New Zealand, *J. Geophys. Res.* **108,** doi: 10.1029/2002JB002041.

Leonard, L. J., and J. M. Bednarski (2014). Field survey following the 28 October 2012 Haida Gwaii tsunami, *Pure Appl. Geophys.* **171,** 3467–3482, doi: 10.1007/s00024-014-0792-0.

Leonard, L. J., and J. M. Bednarski (2015). The preservation potential of coastal coseismic and tsunami evidence observed following the 2012 M_w 7.8 Haida Gwaii thrust earthquake, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140193.

Leonard, L. J., G. C. Rogers, and R. D. Hyndman (2010). Annotated bibliography of references relevant to tsunami hazard in Canada, *U.S. Geol. Surv. Can. Open-File Rept. 6552*, 269 pp.

Leonard, L. J., G. C. Rogers, and S. Mazzotti (2014). Tsunami hazard assessment of Canada, *Nat. Hazards* **70,** 237–274. 19

Lewis, P. D., J. W. Haggart, R. G. Anderson, C. J. Hickson, R. I. Thompson, J. R. Dietrich, and K. M. M. Rohr (1991). Triassic to Neogene geologic evolution of the Queen Charlotte region, *Can. J. Earth Sci.* **28,** 854–869, doi: 10.1139/e91-078.

Lewis, T. J., C. Lowe, and T. S. Hamilton (1997). Continental signature of a ridge-trench-triple junction: Northern Vancouver Island, *J. Geophys. Res.* **102,** 7767–7781.

Lowe, C., and S. A. Dehler (1995). Crustal thickness beneath the Queen Charlotte Basin, Canada: Results of a seismically constrained gravity inversion, *J. Geophys. Res.* **100,** 24,331–24,345.

Mackie, D. J., R. M. Clowes, S. A. Dehler, R. M. Ellis, and P. Morel-À-l'Huissier (1989). The Queen Charlotte Islands refraction project. part II. Structural model for transition from Pacific plate to North American plate, *Can. J. Earth Sci.* **26,** 1713–1725.

Malservisi, R., K. P. Furlong, and H. Anderson (2003). Dynamic uplift in a transpressional regime: Numerical model of the subduction area of Fiordland, New Zealand, *Earth Planet. Sci. Lett.* **206,** 349–364.

Mazzotti, S., R. D. Hyndman, P. Flück, A. J. Smith, and M. Schmidt (2003). Distribution of the Pacific/North America motion in the Queen Charlotte Islands—S. Alaska plate boundary zone, *Geophys.Res.Lett.* **30,** doi: 10.1029/2003GL017586.

McCaffrey, R. (1992). Oblique plate convergence, slip vectors, and forearc deformation, *J. Geophys. Res.* **97,** 8905–8915.

McCaffrey, R., A. I. Qamar, R. W. King, R. Wells, G. Khazaradze, C. A. Williams, C. W. Stevens, J. J. Vollick, and P. C. Zwick (2007). Fault locking, block rotation and crustal deformation in the Pacific Northwest, *Geophys. J. Int.* **169,** 1315–1340, doi: 10.1111/j.1365-246X.2007.03371.x.

McCrory, P. A., and D. S. Wilson (2013). A kinematic model for the formation of the Siletz-Crescent forearc terrane by capture of coherent fragments of the Farallon and Resurrection plates, *Tectonics* **32,** 718–736, doi: 10.1002/tect.20045.

Mueller, S., and R. J. Phillips (1991). On the initiation of subduction, *J. Geophys. Res.* **96,** 651–665, doi: 10.1029/90JB02237.

Nishenko, S. P., and K. H. Jacob (1990). Seismic potential of the Queen Charlotte-Alaska-Aleutian Seismic Zone, *J. Geophys. Res.* **95,** 2511–2532.

Nykolaishen, L., H. Dragert, K. Wang, T. James, and M. Schmidt (2015). GPS Observations of Crustal Deformation Associated with the M_w 7.8 Haida Gwaii Earthquake, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140177.

Oskin, M., and J. Stock (2003). Pacific–North America plate motion and opening of the Upper Delfín basin, northern Gulf of California, Mexico, *Geol. Soc. Am. Bull.* **225,** 1173–1190, doi: ~~10.1016/j.epsl.2011.04.043~~. 11

Pavlis, T. L., J. B. Chapman, R. L. Bruhn, K. Ridgway, L. L. Worthington, S. P. S. Gulick, and J. Spotila (2013). Structure of the actively deforming fold-thrust belt of the St. Elias orogen with implications for glacial exhumation and three-dimensional tectonic processes, *Geosphere* **8,** 991–101, doi: 10.1130/GES00753.1.

Pavlis, T. L., C. Picornell, L. Serpa, R. L. Bruhn, and G. Plafker (2004). Tectonic processes during oblique collision: Insights from the St. Elias orogen, northern North American Cordillera, *Tectonics* **23,** doi: 10.1029/2003TC001557.

Perry, S. E., J. I. Garver, and K. D. Ridgeway (2009). Transport of the Yakutat Terrane, Southern Alaska: Evidence from sediment petrology and Detrital Zircon Fission-Track and U/Pb double dating, *J. Geol.* **117,** 156–173.

Plafker, G., J. C. Moore, and G. R. Winkler (1994). Geology of the southern Alaska margin, in *The geology of Alaska, Geology of North America*, G. Plafker and H. C. Berg (Editors), Vol. G-1, Geol. Soc. Am., Boulder, Colorado, 389–449.

Prims, J., K. P. Furlong, K. M. M. Rohr, and R. Govers (1997). Lithospheric structure along the Queen Charlotte margin in western Canada: Constraints from flexural modeling, *Geo Mar. Lett.* **17,** 94–99.

Reyners, M., and T. Webb (2002). Large earthquakes near Doubtful Sound, New Zealand, 1989-1993, *New Zealand J. Geol. Geophys.* **45,** 109–120.

Riddihough, R. P. (1984). Recent movements of the Juan de Fuca plate system, *J. Geophys. Res.* **89,** 6980–6994.

Riddihough, R. P., and R. D. Hyndman (1989). Queen Charlotte Islands margin, in *The Eastern Pacific Ocean and Hawaii*, E. L. Winterer, D.M. Hussong, and R.W. Decker, The Geological Society of America, Boulder, Colorado, Vol. N, 403–411.

Riddihough, R. P., R. G. Currie, and R. D. Hyndman (1980). The Dellwood knolls and their role in triple junction tectonics off northern Vancouver Island, *Can. J. Earth Sci.* **17,** 577–593.

Ristau, J., G. C. Rogers, and J. F. Cassidy (2007). Stress in western Canada from regional moment tensor analysis, *Can. J. Earth Sci.* **44,** 127–148.

Rogers, G. C. (1986). Seismic gaps along the Queen Charlotte fault, *Earthq. Predic. Res.* **4,** 1–11.

Rohr, K. M. M. (2015). Plate boundary adjustments of the southernmost Queen Charlotte fault, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140162.

Rohr, K. M. M., and J. R. Dietrich (1992). Strike-slip tectonics and development of the Tertiary Queen Charlotte Basin, offshore western Canada: Evidence from seismic reflection data, *Basin Res.* **4,** 1–20, doi: 10.1111/j.1365-2117.1992.tb00039.x.

Rohr, K. M. M., and K. P. Furlong (1995). Ephemeral plate tectonics at the Queen Charlotte triple junction, *Geology* **23,** 1035–1038, doi: 10.1130/0091-7613(1995)023<1035:EPTATQ>2.3.CO;2.

Rohr, K. M. M., and A. J. Tryon (2010). Pacific-North America plate boundary reorganization in response to a change in relative plate motion: Offshore Canada, *Geochem. Geophys. Geosys.* **11,** doi: 10.1029/2009GC003019.

Rohr, K. M. M., M. Scheidhauer, and A. M. Trehu (2000). Transpression between two warm mafic plates: The Queen Charlotte Fault revisited, *J. Geophys. Res.* **105,** 8147–8172, doi: 10.1029/1999JB900403.

Rusmore, M. E., S. W. Bogue, K. Dodson, K. A. Farley, and G. J. Woodsworth (2010). Deformation of continental crust along a transform boundary, Coast Mountains, British Columbia, *Tectonics* **29,** doi: 10.1029/2009TC002502.

Rusmore, M. E., G. Gehrels, and G. J. Woodsworth (2001). Southern continuation of the Coast shear zone and Paleocene strain partitioning in British Columbia–southeast Alaska, *Geol. Soc. Am. Bull.* **113,** 961–975.

Scheidhauer, M. (1997). Crustal structure of the Queen Charlotte Transform Fault Zone from multichannel seismic reflection and gravity, *M.Sc. Thesis*, Oregon State University, Corvallis, Oregon, 184 pp.

Shemenda, A. I. (1992). Horizontal lithosphere compression and subduction: Constraints provided by physical modeling, *J. Geophys. Res.* **97,** 11,097–11,116.

Smith, R. A. J. (1999). Structure, deformation, and thermal regime of the Queen Charlotte Transform margin, *M.Sc. Thesis*, Univ. Victoria, Victoria, British Columbia, 179 pp.

Smith, A. J., R. D. Hyndman, J. F. Cassiday, and K. Wang (2003). Structure, seismicity, and thermal regime of the Queen Charlotte Transform Margin, *J. Geophys. Res.* **108,** doi: 10.1029/2002JB002247.

Souther, J., and A. M. Jessop (1991). Dike swarms in the Queen Charlotte Islands, British Columbia, and implications for hydrocarbon exploration, in *Evolution and Hydrocarbon Potential of the Queen Charlotte Basin, British Columbia*, G. J. Woodsworth (Editor), *Pap. Geol. Surv. Can. 90-10*, 465–487. 16

Spence, G. D., and I. Asudeh (1993). Seismic velocity structure of the Queen Charlotte basin beneath Hecate Strait, *Can. J. Earth Sci.* **30,** 787–805.

Spence, G. D., and D. T. Long (1995). Transition from oceanic to continental crustal structure: Seismic and gravity models at the Queen Charlotte transform margin, *Can. J. Earth Sci.* **32,** 699–717.

Stern, R. J., and S. H. Bloomer (1992). Subduction zone infancy: Examples from the Eocene Izu-Bonin-Mariana and Jurassic California arcs, *Geol. Soc. Am. Bull.* **104,** 1621–1636.

Stock, J. M., and P. Molnar (1988). Uncertainties and implications of the Late Cretaceous and Tertiary position of NorthAmerica relative to the Farallon, Kula, and Pacific plates, *Tectonics* **7,** 1339–1384.

Sutherland Brown, A. (1968). Geology of the Queen Charlotte Islands, *Bull. Br. Columbia Depart. Mine. Petrol. Resour.* **54,** 226 pp.

Szeliga, W. (2012). Haida Gwaii Quake: Insight Into Cascadia's Subduction Extent, *Eos Trans. AGU* **94,** 85–86, doi: 10.1002/2013EO090001.

Toth, J., and M. Gurnis (1998). Dynamics of subduction initiation at preexisting fault zones, *J. Geophys. Res.* **103,** 18,053–18,067, doi: 10.1029/98JB01076.

Von Huene, R., G. G. Shor Jr., and J. Wageman (1979). Continental margins of the eastern Gulf of Alaska and boundaries of tectonic plates, in *Geological and Geophysical Investigations of Continental Margins*, Am. Assoc. Petrol. Geol. Memoir, Vol. **29,** 273–290.

Walton, M. L., E. C. Roland, S. S. Gulick, P. J. Haeussler, G. L. Christeson, and H. J. Van Avendonk (2013). New mapping and structural constraints on the Queen Charlotte-Fairweather Fault system, southeast Alaska, *Am. Geophys. Un. Annual Meeting*, San Francisco, California, December 2013, Abstract T43A–2624.

Wang, K., J. He, F. Schulzeck, R. D. Hyndman, and M. Riedel (2015). Thermal condition of the 27 October 2012 M 7.8 Haida Gwaii subduction earthquake at the obliquely convergent Queen Charlotte margin, *Bull. Seismol. Soc. Am.* **105,** no. 2B, doi: 10.1785/0120140183.

Yorath, C. J., and R. L. Chase (1981). Tectonic history of the Queen Charlotte Islands and adjacent areas—A model, *Can. J. Earth Sci.* **18,** 1717–1739.

Yorath, C. J., and R. D. Hyndman (1983). Subsidence and thermal history of Queen Charlotte Basin, *Can. J. Earth Sci.* **20,** 135–159.

Yuan, T., G. D. Spence, and R. D. Hyndman (1992). Structure beneath Queen Charlotte Sound from seismic-refraction and gravity interpretations, *Can. J. Earth Sci.* **29,** 1509–1529.

Pacific Geoscience Centre
Geological Survey of Canada
9860 W. Saanich Road
Sidney, British Columbia
V8L4B2 Canada
rhyndman@nrcan.gc.ca

Manuscript received 21 June 2014

Queries

1. AU: Please verify author initials and affiliations. Also please provide complete postal address along with zip code.
2. AU: The citation "Smith *et al.* (1999)" does not have a corresponding Reference entry. There is a "Smith (1999)" in the References, which is not cited in the paper. Please (1) decide whether these refer to the same work and (2) indicate the required changes to the paper and the References.
3. AU: The citation "Chase (1975)" does not have a corresponding Reference entry. There is a "Chase et al. (1975)" in the References, which is not cited in the paper. Please (1) decide whether these refer to the same work and (2) indicate the required changes to the paper and the References.
4. AU: SSA avoids using quotation marks. The quotation marks have been deleted, but please provide alternative wording if desired.
5. AU: The citation "Bérubé (1989)" does not have a corresponding Reference entry. There is a "Bérubé et al. (1989)" in the References, which is not cited in the paper. Please (1) decide whether these refer to the same work and (2) indicate the required changes to the paper and the References.
6. AU: Please note that you have cited Fitch (1986) in the text but no such reference is seen in the list. Kindly provide publication details foe the same or delete the citation.
7. AU: Please indicate if the roman capital M should be changed to (1) **M** (bold) or (2) Mw.
8. AU: Please check the edit made in Data and Resources section.
9. AU: Please provide last accessed date for the url http://hdl.handle.net/2429/24553.
10. AU: Please provide intext citation for Evenchick et al. (1999) or delete the reference from the list.
11. AU: The DOI does not match reference entry. Do you intend to use "10.1130/B25154.1"?
12. AU: Please note that you have cited Barrie *et al.* (2012) in the figure captions but no such reference is seen in the list. Kindly provide publication details for the same or delete the citation.
13. AU: The citation "Cassidy *et al.* (2013)" does not have a corresponding Reference entry. There is a "Cassidy *et al.* (2014)" in the References. Please (1) decide whether these refer to the same work and (2) indicate the required changes to the paper and the References.
14. AU: Please note that you have cited Spence and Long (1997) in the text but no such reference is seen in the list. Kindly provide publication details for the same or delete the citation.
15. AU: Please note that you have cited Hyndman and Wang (1993) in the text but no such reference is seen in the list. Kindly provide publication details for the same or delete the citation.
16. AU: Please provide intext citation for Souther and Jessop (1991) or delete the reference from the list.
17. AU: Please provide intext citation for Chamot-Rooke and LePichon (1981) or delete the reference from the list.
18. AU: Please provide intext citation for Davis and Clowes (1986) or delete the reference from the list.
19. AU: Please provide intext citation for Leonard et al. (2014) or delete the reference from the list.