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The effects of upper plate deformation on records of prehistoric Cascadia subduction zone earthquakes

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Abstract: Geophysical data from the offshore Cascadia forearc reveal many Quaternary upper-plate faults and folds. Most active structures are within the accretionary wedge, but significant deformation is also found on the continental shelf. Several faults and synclines project into adjacent coastal bays where deformation of Pleistocene marine terraces is reported. Rapidly buried marsh deposits and drowned forests in these coastal lowlands are interpreted to record coseismic deformation by prehistoric subduction zone earthquakes. The extent and amount of such coastal subsidence has been used to infer characteristic magnitudes and recurrence intervals. However, the record may incorporate both elastic strain release on the subduction zone and localized permanent upper-plate deformation. Movement on upper-plate structures may be triggered by a subduction zone earthquake, as observed in the Nankai and Alaskan forearcs. Alternatively, they may deform independently of subduction zone earthquakes. Regardless of which style of deformation predominates, the record of coseismic subsidence is likely to be affected. Crustal deformation may also contribute to the preservation of subsided marshes. Modelling of subduction zone earthquake characteristics based on coastal marsh stratigraphy is likely to be inaccurate in terms of: (a) total apparent rupture length and earthquake magnitude; (b) amount of subsidence and hence the position of the locked zone; (c) recurrence interval. Most of these shelf and coastal structures respond to N-S compression, in contrast to convergence-related northeasterly compression in the accretionary prism, but in agreement with the regional stress field. Despite low historical coastal and continental shelf seismicity, upper-plate faults may also pose an independent seismic hazard.

Geological and geophysical investigations of the Cascadia subduction zone during the last decade have increased public awareness of regional earthquake hazards from a subduction zone previously thought to be aseismic (Ando & Balazs 1979). Evidence for repeated abrupt subsidence in the last few thousand years is found in coastal bays along the active margin, in the form of buried marsh deposits and drowned forests (e.g. Atwater 1987, 1992; Darienzo & Peterson 1990; Atwater et al. 1995; Nelson et al. 1995; Yamaguchi et al. 1997). Previous workers have believed these deposits to be a result of coseismic subsidence and have attributed them to subduction zone earthquakes (e.g. Atwater 1987, 1992; Darienzo & Peterson 1990; Atwater et al. 1995; Nelson et al. 1995; Yamaguchi et al. 1997). However, the similarity of these deposits to the marsh stratigraphy of tectonically inactive coasts has led to the suggestion that many abrupt burials may be non-tectonic in origin and driven by local changes in intertidal environment (e.g. Long & Shennan 1994). A non-tectonic origin cannot be eliminated except in certain cases where the event is found to be regionally widespread, is associated with other coseismic phenomenon such as tsunami deposits or liquefaction, or where the magnitude of subsidence is too large to be explained by non-tectonic mechanisms. The coseismic v. non-seismic origin of the subsidence events will not be addressed by this paper, but remains a topic of debate.

The chronology, distribution, and amount of subsidence for individual locations and events have been used to estimate recurrence intervals and magnitudes. A significant component of subsidence recorded at these sites could be attributed to localized permanent crustal deformation (Goldfinger et al. 1992b; Nelson 1992; Goldfinger 1994; McCaffrey & Goldfinger 1995; McCrory 1996; Nelson & Personius 1996; Clague 1997). This study shows that several bays lie within actively deforming synclines or on the downthrown side of faults mapped onshore and in the contiguous offshore inner shelf. Mapping

the offshore region benefited from extensive geophysical datasets and the absence of thick vegetation, which often hinders coastal fieldwork. A question critical to the understanding of contributions by crustal structures to the subsidence record is: Do crustal and subduction zone earthquakes operate independently or together? Regardless of the answer to this question, localized upper-plate deformation calls into question calculations of recurrence intervals and earthquake magnitude obtained from records of coastal subsidence.

The objectives of this study were to consider the distribution of abruptly buried deposits in light of upper-plate Quaternary deformation in the offshore inner shelf and onshore coastal region, and to determine the effects on and limitations of this palaeeoseismic record for determination of prehistoric subduction-zone earthquake characteristics. We show that such structures reflect the predominant structural regime and stress field of the shelf and coast, and suggest that these localized structures may pose an independent seismic hazard to coastal communities.

**Methods**

Structures outlined in this paper were mapped from single-channel and multichannel seismic reflection profiles, sidescan sonar data, and submersible observations. Seismic data were collected by Oregon State University (OSU), University of Washington, the US Geological Survey, and the petroleum industry. Many of the seismic profiles extend to within a few kilometres of the coastline.

One proprietary dataset consists of a closely spaced network (between 2 and 30 km apart) of high-quality, precision-navigated, migrated multichannel profiles. This dataset covers the shelf and uppermost slope of Oregon and Washington. Acquisition details of this proprietary dataset have been are discussed by McNeill et al. (1997). Many of the structures identified in this paper were mapped from this particular dataset, with the data shown here in the form of migrated time sections.

Deep-towed sidescan sonar data were collected during several research cruises between 1992 and 1995, and were navigated using global positioning system (GPS). Sidescan sonar data on the shelf include the AMS150SIkHz and Klein systems. Details of sidescan sonar data, and sonar processing and imaging techniques have been given by Goldfinger et al. (1997b). The shallow-diving submersible, DELTA, was used to dive on fault targets selected from sidescan sonar images and seismic reflection data.

**Tectonic setting**

The Cascadia subduction zone is located off the coast of the northwestern United States and southwestern Canada, a result of subduction of the Juan de Fuca and Gorda plates beneath the North American plate (Fig. 1 inset). Convergence is oblique to the northeast at a rate of 42 mm/year at latitude 47°30'N (NUVEL-1 plate motion model of De Mets et al. 1990). No earthquakes have been recorded on the plate boundary during the period of recorded seismicity, with the possible exception of the 1992 Petrolia earthquake (Oppenheimer et al. 1993). Crustal seismicity in the North American plate and within the subducting Juan de Fuca plate is minimal but present, with greatest recorded seismicity in Washington (Puget Sound) and in the Gorda plate off northern California (e.g. Crosson & Owens 1987; Weaver & Baker 1988; Ludwin et al. 1991). The lack of seismicity on the megathrust led early workers to infer that subduction may have ceased or that subduction is aseismic (Ando & Balazs 1979). An alternative interpretation, that strain is currently accumulating and will be released in a future large-magnitude earthquake, was supported by evidence for abrupt potentially coseismic coastal subsidence (Atwater 1987) and regularly spaced turbidites, potentially earthquake induced, in submarine channels on the Juan de Fuca abyssal plain (Adams 1990). Both lines of evidence point to several events in the last few thousand years. Coseismic subsidence in the coastal Cascadia region is predicted from elastic dislocation models of the subduction zone cycle. Figure 2 illustrates this cycle, where the land surface a certain distance from the deformation front is expected to gradually uplift during the interseismic period, followed by sudden subsidence during the seismic event. This region is expected to coincide with the coast in Cascadia. A similar elastic response was recorded accompanying the 1960 Chilean and 1964 Alaskan subduction earthquakes (Pfaffker 1969, 1972), with regions of coseismic uplift and subsidence identified. The earthquake potential of the Cascadia subduction zone based on coastal subsidence and abyssal plain turbidites was reinforced by analogies between Cascadia and Chilean-type subduction zones (Heaton & Kanamori 1984).

**Regional stratigraphy and structure**

Cenozoic strata underlying the continental shelf consist of Eocene to Quaternary bathyal to neritic forearc basin and accretionary complex sequences resting in part on early Eocene basalt (Snavely 1987; Palmer & Lingley 1989; Snavely & Wells 1996). Several regional unconformities are prominent within the basinal sequence, including...
Fig. 1. Neotectonic map of recent structures (late Miocene to Holocene anticlines and faults only), including inner shelf and coastal structures outlined in this paper, and locations of coastal subsidence. Inset shows general tectonic setting with plate convergence vector (De Mets et al. 1990). Plates: P, Pacific; JDF, Juan de Fuca; NA, North American; G, Gorda. FZ, Fracture Zone; SNF, left-lateral South Nitinat Fault. Northern California offshore structures after Clarke (1990, 1992; reproduced by kind permission of AAPG and GSA).

The lower continental slope is dominated by compressional tectonics within the active accretionary prism in response to plate convergence, with structural trends between N–S and NW–SE.

Fig. 2. Model showing the subduction earthquake cycle (after Darienzo & Peterson 1990; reproduced by kind permission of GSA). During the interseismic period (a period of hundreds of years), strain accumulates causing much of the Cascadia coastline to gradually uplift (between the two hinge lines). Coseismic strain release occurs during the subduction earthquake and causes this region to rapidly subside, resulting in drowning of the coastal area. The seaward hinge line represents the zero isobase, seaward of which coseismic uplift occurs. No vertical movement occurs at the hinge lines. Under purely elastic conditions, subsidence is completely recovered with no net subsidence or uplift through the earthquake cycle. NA, North American plate; JDF, Juan de Fuca plate.
(Fig. 1; Goldfinger et al. 1992a, 1997a). A set of WNW-trending left-lateral strike-slip faults, a result of oblique convergence, was mapped on the continental slope and locally on the outermost shelf of Oregon and Washington (Fig. 1; Goldfinger 1994; Goldfinger et al. 1992b, 1996, 1997a). On the northern Oregon and Washington shelf and upper slope, E–W extension is common in the form of listric normal faulting related to the underlying mobile mélangé and broken formation (McNeill et al. 1997). Complex fold trends on the Washington upper slope may be partially controlled by mobilization of the mélangé and broken formation (McNeill et al. 1997). Structural styles are more varied on the continental shelf, where many of the Miocene to early Pleistocene structures are no longer active (Fig. 1; Goldfinger et al. 1992a). Active fold axes on much of the inner shelf trend perpendicular to the coastline and margin (Goldfinger et al. 1992b; Goldfinger 1994). These fold trends suggest N–S compression rather than dominant northeasterly compression within the active accretionary prism.

Active crustal structures and coastal subsidence

Introduction

To date, coastal subsidence in the form of rapidly buried marshes has been identified at the following bay and river locations along the Cascadia subduction zone, from north to south: Vancouver Island sites (Ucluelet and Tofino), Neah Bay, Pysht River, Copalis River, Grays Harbor, Willapa Bay, Columbia River, Necanicum River, Nehalem Bay, Tillamook Bay, Netarts Bay, Nestucca Bay, Salmon River, Siletz Bay, Yaquina Bay, Alsea Bay, South Slough, Coquille River, Sixes River, and Humboldt Bay (Fig. 1; based on Atwater et al. 1995; Barnett 1997). Evidence of Quaternary structural downwarping is described or documented here at Grays Harbor, Willapa Bay, Nehalem Bay, Tillamook Bay, Netarts Bay, Siletz Bay, Yaquina Bay, Alsea Bay, South Slough, Coquille River, and Humboldt Bay (Table 1). At these locations, buried marshes are located within Quaternary synclinal folds or on the downthrown side of faults. Buried marsh locations where evidence of Quaternary deformation is inconclusive are also documented, along with two examples of buried marshes which may be located on the upthrown side of a Quaternary fault or on the crest of an anticline. Table 1 describes the Quaternary structures at each of these locations.

Buried marshes in areas of late Quaternary subsidence

Grays Harbor. Structure contours of the late Pliocene–Pleistocene unconformity west of Grays Harbor indicate a narrow E–W to NE–SW trending depression (Fig. 3). This active syncline lies due west of Grays Harbor, which may be structurally controlled. Buried marshes have been found throughout the bay (Fig. 3) and may not all be associated with downwarping of this particular syncline. This depression is tectonically controlled and not the simple result of backfilling of a Pleistocene lowstand channel.

Willapa Bay. Buried marshes and drowned forests provide evidence of rapid subsidence throughout Willapa Bay, southern Washington coast (Figs 1 and 3; e.g. Atwater 1987; Atwater & Hemphill-Haley 1996; Yamaguchi et al. 1997). Multichannel seismic reflection profiles 10–30 km west of the bay reveal a broad syncline 40 km in length, extending from the northern to the southern end of the bay (Figs 3 and 4). The synclinal axis trends NW–SE to E–W and projects into the centre of Willapa Bay. To the north and south of the syncline, two bounding anticlines may be underlain by faults with reverse separation (strike-slip component unknown) (Fig. 3). The northern N-dipping reverse fault may be blind (A1, Fig. 4). Alternatively, this fault may be the southeastern projection of a left-lateral fault which deforms Holocene sediments on the continental slope (South Nitinat Fault of Goldfinger (1994) and Goldfinger et al. (1997a); SNF in Fig. 1). The Willapa Bay structures deform Eocene to late Miocene mélange and broken formation and overlying late Miocene to Quaternary basinal sediments, and gently deform the seafloor. The northern fault was investigated during a submersible and sidescan sonar cruise in 1995. A submersible dive found no evidence of seafloor offset, possibly as a result of high-energy wave action in shallow water or because the fault is blind. However, evidence of carbonate cementation and algal mats was observed, suggesting fluid venting accompanying active faulting, as observed elsewhere on the margin (Kulm & Suess 1990; Goldfinger et al. 1997a). The southern anticline (A2) and underlying fault may be inactive in the Quaternary (minimal deformation of the Pliocene–Pleistocene unconformity in Fig. 4). Structure contours of the late Pliocene–Pleistocene unconformity also outline the position of the Willapa syncline (Fig. 3). The different positions of the late Pliocene (dashed line), the early Pleistocene (unconformity
Table 1. Quaternary deformation at marsh burial locations.

<table>
<thead>
<tr>
<th>Site</th>
<th>Local structures</th>
<th>Quaternary</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tofino</td>
<td>unknown</td>
<td>unknown</td>
<td>Snively 1987 (Calawah fault)</td>
</tr>
<tr>
<td>Ucluelet</td>
<td>unknown</td>
<td>unknown</td>
<td>Gower 1960; Tabor &amp; Cady 1978; Wagner et al. 1987</td>
</tr>
<tr>
<td>Neah Bay</td>
<td>inconclusive</td>
<td>inconclusive</td>
<td>McCrory 1996</td>
</tr>
<tr>
<td>Psht River</td>
<td>yes</td>
<td>unproven</td>
<td>This study; Grim &amp; Bennett 1969; Palmer &amp; Lingley 1989</td>
</tr>
<tr>
<td>Copalis River</td>
<td>yes, Langley anticline</td>
<td>yes</td>
<td>McCrory 1996</td>
</tr>
<tr>
<td>Grays Harbor</td>
<td>yes</td>
<td>yes</td>
<td>This study</td>
</tr>
<tr>
<td>Willapa Bay</td>
<td>yes, Fern Hill fault, Youngs Bay syncline</td>
<td>yes</td>
<td>This study; Niem &amp; Niem 1985; Ryan &amp; Stevenson 1995</td>
</tr>
<tr>
<td>Columbia River</td>
<td>yes, Youngs Bay syncline</td>
<td>unproven</td>
<td>This study; Niem &amp; Niem 1985; Wells et al. 1992; Goldfinger et al. 1992; Goldfinger 1994</td>
</tr>
<tr>
<td>Necanicum River</td>
<td>inconclusive</td>
<td>inconclusive</td>
<td>This study; Parker 1990; Goldfinger et al. 1992; Wells et al. 1992, 1994; Goldfinger 1994</td>
</tr>
<tr>
<td>Nehalem Bay</td>
<td>yes, Cape Falcon fault</td>
<td>unproven</td>
<td>This study; Goldfinger et al. 1992</td>
</tr>
<tr>
<td>Tillamook Bay</td>
<td>yes, Tillamook Bay fault</td>
<td>unproven</td>
<td>This study; Goldfinger et al. 1992</td>
</tr>
<tr>
<td>Netarts Bay</td>
<td>yes, Nehalem Bank and Happy Camp faults</td>
<td>yes</td>
<td>This study; Goldfinger et al. 1992; Goldfinger 1994</td>
</tr>
<tr>
<td>Nestuca Bay</td>
<td>yes</td>
<td>no</td>
<td>This study; Goldfinger et al. 1992; Goldfinger 1994</td>
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<tr>
<td>Salmon River</td>
<td>inconclusive</td>
<td>inconclusive</td>
<td>Kelsey et al. 1990; Goldfinger et al. 1992; Kelsey et al. 1990</td>
</tr>
<tr>
<td>Siletz Bay</td>
<td>yes</td>
<td>yes</td>
<td>McNelly &amp; Kelsey 1990; Goldfinger et al. 1992; Goldfinger 1994</td>
</tr>
<tr>
<td>Yaquina Bay</td>
<td>yes</td>
<td>yes</td>
<td>This study; McNelly &amp; Kelsey 1990</td>
</tr>
<tr>
<td>Alsea Bay</td>
<td>yes</td>
<td>yes</td>
<td>Kelsey 1990</td>
</tr>
<tr>
<td>South Slough</td>
<td>yes</td>
<td>yes</td>
<td>Clarke &amp; Carver 1992</td>
</tr>
<tr>
<td>Coquille River</td>
<td>yes, anticline</td>
<td>yes</td>
<td>Kelsey 1990</td>
</tr>
<tr>
<td>Sixes River</td>
<td>yes, Little Salmon and Mad River faults, Freshwater syncline</td>
<td>yes</td>
<td>Kelsey 1990</td>
</tr>
<tr>
<td>Humboldt Bay</td>
<td>yes</td>
<td>yes</td>
<td>Kelsey 1990</td>
</tr>
</tbody>
</table>

Presence or absence of localized Quaternary deformation at marsh burial locations documented here and in published literature. Local structures are synclines or downthrown sides of faults indicating subsidence, except for those shown in italics, which show evidence of uplift. (See Fig. 1 for locations.)

structure contours), and the most recently active synclinal axis (continuous line) indicate that the fold axis has migrated north from Pliocene time to the present. Early Pliocene growth strata thicken towards the southern anticline (Fig. 4 A2), whereas overlying late Pliocene strata thin towards the anticline, indicating that growth of this anticline occurred during the late Pliocene. Before this time, the syncline was formerly a larger structure extending to the south. The position of synclinal folding with respect to Willapa Bay strongly suggests that the bay is structurally controlled.

Exposures of Pleistocene marine terraces and underlying sediments are widespread and continuous throughout the central region of Willapa Bay. Amino acid racemization of shell deposits within the lowest altitude marine terrace sediments of the central bay suggest conflicting ages. Kvenolden et al. (1979) identified two separate highstands, with absolute ages of 120 ka BP and 190 ka ± 40 ka BP. Amino acid ages with supporting faunal correlations and palaeoecology from Kennedy (1978) and Kennedy et al. (1982) suggest a more probable and precise age of 80–85 ka BP (West 1986). Despite the unresolved terrace chronology, relative altitudes of continuous and correlative terraces have been documented. The lowest terrace surface is reported to be at a constant altitude of c. 13–15 m above sea level in the central region of the bay (West 1986). However, reconnaissance investigation of terrace elevations for this study found that the terrace actually reaches a minimum elevation of c. 9–10 m at Sandy Point (Sa, Fig. 3). There are no signs of landslides or slumping in this region,
and therefore terrace elevations are assumed to be undisturbed. In the northern bay, the lowest terrace, which West (1986) assumed to be the continuation of the central bay's lowest terrace (80–85 ka BP?), increases in altitude to 24 m (West 1986). South of Nemah (N, Fig. 3), limited exposures of terraces of unknown age range in altitude from 12 to 18 m. These altitude changes, although only loosely constrained by age control, are in agreement with a synclinal axis close to the centre of the bay. The stratigraphy of the estuarine sediments underlying the lowest and youngest central bay terrace shows no clear evidence of deformation along a N–S transect (Kvenvolden et al. 1979; Clifton 1983, 1994; H. E. Clifton, pers. comm., 1997). However, a unit of older (early Pleistocene or late Pliocene?) well-indurated estuarine terrace sediments including a sequence of palaeosols, locally exposed within the youngest terrace, may be deformed (Fig. 5; Clifton 1994; H. E. Clifton, pers. comm., 1997). Between Stony Point and the Bone River (Fig. 3), these surfaces dip gently to the south (Fig. 5; Clifton 1994). If these surfaces were originally approximately horizontal, they are now deformed in agreement with the projection of the offshore projected synclinal axis.

The WNW trend of Quaternary structures offshore contrasts slightly with NW- to NNW-trending structures which deform Miocene and older formations mapped east of Willapa Bay (Fig. 3; Walsh et al. 1987). This change in trend may reflect a change in compressive stress field from the early to late Tertiary, or non-linearity of mapped structures. One of two sub-parallel NW-trending synclinal axes southeast of the bay (Fig. 3; Walsh et al. 1987) could be the landward projection of the offshore structure.

Atwater & Hemphill-Haley (1996) addressed the possibility of deformation by upper-plate structures in Willapa Bay by studying the marsh stratigraphy on and off the N–S trending South Bend antiform (Fig. 3). They found no difference in the amount of subsidence between these sites, although marshes are buried less deeply below the current marsh surface on, rather than off the antiform (Atwater & Hemphill-Haley 1996). However, this antiform deforms sediments no younger than Miocene and may not be currently active. An E–W trending structure, such as the Willapa Bay syncline, should show no measurable difference in amount of subsidence along an approximately E–W transect, and therefore the results of Atwater & Hemphill-Haley (1996) are not unexpected.

The offshore northern bounding anticline and thrust fault project landward towards one location of marsh burial in the northernmost bay (Fig. 3). Evidence for subsidence at this site (Cedar River (CR in Fig. 3) of Atwater & Yamaguchi (1991)) is a western red cedar snag which died within a few months of AD 1700. This apparent contradiction can be explained if the crustal structure was not activated during the AD 1700 event, or if subsidence caused by the subduction event exceeded coseismic uplift on the crustal structure with resulting net subsidence, or if the fault and anticline changestrike onshore to project to the north of this particular
Fig. 4. N–S proprietary migrated multichannel seismic reflection profile 10 km west of Willapa Bay. The lower interpreted section shows late Miocene to Quaternary sediments and underlying melange are deformed by an active syncline and associated structures. To the north, Pleistocene sediments are deformed by an active N-dipping blind reverse fault. Regional angular unconformities: M, P, late Miocene–Pliocene; Plio–Pleist, Pliocene–Pleistocene.

subsidence site. In addition, the absolute vertical motion may differ from the observed relative vertical motion across this anticline.

Nehalem and Tillamook Bays. Seismic reflection profiles west of Nehalem Bay reveal synclinal deformation opposite the bay and uplift on the north side of the bay which may be equivalent to the Tertiary Cape Falcon fault of Niem & Niem (1985) (Fig. 6). The middle Miocene Columbia River Basalt (CRB in Fig. 7) is deformed by these structures and exposed at Cape Falcon (Niem & Niem 1985).

The northern margin of Tillamook Bay is uplifted by the WNW-trending high-angle Tillamook Bay fault, which offsets Columbia River Basalt onshore (Wells et al. 1992). Offset on this

Fig. 5. Early Pleistocene terrace sediments between Stony Point (St) and Bone River (BR) in Willapa Bay showing S-dipping palaeosols (after Clifton 1994; reproduced by kind permission of GSA), in agreement with the projection of the offshore synclinal axis to a position south of this location. (See Fig. 3 for locations.)

Fig. 6. Location map of northern Oregon coast and shelf, including active structures near Nehalem, Tillamook, and Netarts Bays. The offshore Nehalem Bank fault and associated syncline projects onshore as the Happy Camp fault (HCF) on the north side of Netarts Bay (Parker 1990; Wells et al. 1992, 1994). Tillamook and Nehalem Bay may also lie within active synclines or on the downthrown side of active faults (Tillamook Bay fault of Wells et al. (1992); Cape Falcon fault of Niem & Niem (1985)). Onshore structures after Parker (1990), Wells et al. (1992, 1994), and Niem & Niem (1985). Bold line represents N–S seismic reflection profile shown in Fig. 7.
Fig. 7. N–S proprietary multichannel seismic reflection profile between Netarts Bay and Nehalem Bay, 10 km west of the coastline. The Netarts Bay syncline and Nehalem Bank fault project into Netarts Bay. This section reveals sea-floor deformation by the southeasterly segment of the Nehalem Bank fault and synclinal deformation on the downthrown side of the fault. The middle Miocene Columbia River Basalt Group (CRB, bright reflector) is noticeably deformed. Sediments (assumed mostly late Miocene) show little or no thickening within the syncline, suggesting that deformation on this structure is largely post-late Miocene. Synclinal folds are also found opposite Nehalem and Tillamook Bays, although the Tillamook structure is complicated by anticlinal deformation within the broad low. Uplifted regions north of Tillamook and Nehalem Bays may be equivalent to the onshore Tillamook Bay fault and Cape Falcon fault, respectively.
fault is up to the north and may have a strike-slip component, but Quaternary deformation is unconfirmed. Deformation opposite Tillamook Bay is more complex than at Nehalem Bay, but a broad low between the Happy Camp fault at Cape Meares and an uplifted region at Twin Rocks can be seen in Fig. 7. This low is interrupted by smaller fold axes. The uplifted region north of the bay may be related to the Tillamook Bay fault (Figs 6 and 7).

Owing to the thin Quaternary section on the innermost shelf, offshore Quaternary deformation cannot be confirmed, but the onshore Tillamook Bay fault suggests Quaternary activity. Rapid marsh burial has been identified at both bays (Atwater et al. 1995; Barnett 1997).

Netarts Bay. Netarts Bay, on the northern Oregon coast (Fig. 1), is bounded to the north by WNW-trending, NE-dipping high-angle reverse faults which deform coastal sediments, similar in style to Tillamook Bay to the north (Wells et al. 1992, 1994). Faults thrust middle Miocene Columbia River Basalt Group (15 Ma) over late Pleistocene gravels, and may also offset the youngest Pleistocene marine terrace surface (Wells et al. 1994). The fault is known as the Happy Camp fault onshore (Parker 1990; Wells et al. 1992, 1994) and is the southernmost extension of the prominent Nehalem Bank fault zone, which deforms Miocene to Holocene sediments offshore (Fig. 6; Niem et al. 1990; Goldfinger 1994). This complex zone of deformation trends roughly N-S on the outer shelf and upper slope off the northern Oregon coast (25 km west of the coastline), but changes to the southeast at its southern end to project onshore just north of Netarts Bay (Fig. 6). No evidence of major strike-slip offset has been found along the northern segment of the fault, but its orientation (suggesting margin-parallel right-lateral offset where oblique convergence is partitioned into a compressional and strike-slip component), the presence of minor strike-slip faulting, and the linearity of the fault, which truncates bedding planes in AMS 150 kHz side-scan images, support a strike-slip component. However, the fault also shows significant vertical offset (both along its N-S and SE-trending segments) and we interpret it as a reverse fault system downthrown to the west and south. As the fault trends southeasterly, the zone of deformation becomes less complex, being characterized by a N-dipping (possibly blind) reverse fault, with Miocene sediments uplifted and exposed at the sea floor, and an asymmetrical syncline to the south (south end of Fig. 7), which lies immediately opposite Netarts Bay. Sidescan images show bedrock within the hanging-wall anticline exposed on the sea floor and offset by minor N- to NNE-trending right-lateral faults. The vertical motion on the northern segment of the fault may be a flower structure or transpression deformation.

The nearshore Nehalem Bank fault clearly deforms and offsets the middle Miocene Columbia River Basalt Group (highly reflective in seismic reflection data, south end of Fig. 7) and overlying sediments. Absence of structural growth of strata within the syncline in Fig. 7 suggests this fault and associated fold post-date late Miocene sedimentation. Investigation of other seismic reflection data across both the southern and northern sections of the fault shows minimal thinning in late Miocene sediments and some thinning of Pliocene sediments across the fault-controlled anticline. This indicates that the fault was active as early as the late Miocene, but the bulk of deformation has taken place during the Pliocene and Quaternary. Vertical seafloor offset of 10–20 m across the

Fig. 8. Location map of Siletz Bay on the central Oregon coast showing structures mapped offshore from single-channel and multichannel seismic reflection profiles and onshore from Pleistocene marine terrace deformation (structures in Fig. 9 cross-section). Bold line indicates position of single-channel line in Fig. 10. Onshore and offshore deformation suggests that Siletz Bay is structurally downwarped within a syncline or on the downthrown side of a fault. Generalized locations of subsided marshes after Darienzo et al. (1994).
Fig. 9. Cross-section of beach exposure of Pleistocene sediments underlying the youngest marine terrace at Siletz Bay. Marker horizons, including a clay horizon, gravel beds, and the wavecut platform were used to determine deformation of Pleistocene sediments and the location of Quaternary structures. The laterally continuous clay horizon is apparently offset across the bay (up to the north). Faults also offset the wavecut platform at Fishing Rock and Fogarty Creek. The trends of these structures are poorly defined.

fault zone is estimated from sidescan and seismic records, presumed to post-date late Pleistocene lowstand erosion on the shelf (Goldfinger 1994). Cooper (1981) and Parker (1990) also suggested that a west-plunging Miocene syncline is centred about Netarts Bay.

*Siletz Bay.* Structures deforming the underlying sediments and wavecut platform of Pleistocene marine terraces (presumed 80 ka BP Whiskey Run terrace, West & McCrumb 1988; Kelsey 1990) have been identified in the Siletz Bay region of the central Oregon coast (Figs 1, 8 and 9). The wavecut platform and a locally continuous carbonaceous clay horizon, interpreted as a lagoonal deposit or Palaeosols and assumed to be initially sub-horizontal, were shown to be deformed. Variations in elevation of the wavecut platform and the clay horizon/palaeosol may alternatively be controlled by existing topography at the time of formation, and not by deformation. Variations in altitude of these marker horizons indicate faulting, with vertical offsets of 5–30 m and broad folding, with a wavelength of 8–12 km (Figs 8 and 9), assuming the terrace is the same age throughout. Beach exposures alone indicate trends between NNW and SSW, but offshore data (see below) provide more precise trends. The clay horizon of the youngest terrace (80 ka BP) dips gently north between 5 km and 3 km south of the Siletz River mouth, where it is below beach level and projected below sea level (Fig. 9). This clay horizon is exposed again c.10 m above beach level just north of the river mouth, where the wavecut platform is at c.2 m elevation (Fig. 9). The platform is presumed below sea level south of the river, inferred from the elevation and dip of the clay horizon. Projection of the clay horizon below beach level suggests maximum fault offset across the Siletz River mouth of c.30 m up to the north (offset could be a combination of folding and faulting). If this horizon is assumed to be the same age as the terrace (80 ka BP), this produces a late Quaternary vertical slip or subsidence rate relative to terrace levels across the river mouth of 0.4 mm/year. This is a maximum slip rate as sediments

Fig. 10. Line drawing of N–S trending OSU single-channel sparker profile, 6 km offshore Siletz Bay (bold line in Fig. 8). The MP (late Miocene–early Pliocene) unconformity is projected to the sea floor at the southern end of the profile. Profile shows synclinal deformation of presumed late Miocene strata off Siletz Bay, and sea floor offset by possible flexural-slip faults within an active synclinal fold west of Fishing Rock and Fogarty Creek. Fault dips are poorly constrained by seismic data.
underlying the terrace are somewhat older than 80 ka BP. Siletz Bay may lie in a Quaternary syncline controlled by a fault at the northern end of the bay, similar to the structure observed at Netarts Bay. Fault offset (downthrown to the north) of the wavecut platform and marine terrace was also documented at Fishing Rock and Fogarty Creek (Fig. 8; Priest et al. 1994). Orientation of these two faults is poorly defined, but previously mapped faults onshore are oriented NW–SE and NE–SW. Poor exposure prevents the determination of any strike-slip component on onshore faults.

The possible correlative of a syncline at Siletz Bay is traced on N–S trending seismic reflection profiles 4–17 km offshore. Figure 10 is a line drawing of a N–S single-channel seismic profile 6 km west of Siletz Bay, which clearly shows synclinal deformation opposite the bay. The late Miocene–early Pliocene (MP) unconformity is truncated at the sea floor and therefore the age of the youngest strata is late Miocene. The sea floor indicates no synclinal deformation, therefore offshore Quaternary deformation at this scale is unconfirmed. The trend of the syncline across several profiles is between E–W and ESE–WNE. Deformed synclinal sediments on middle- to outer-shelf profiles are truncated by the MP unconformity, indicating little or no activity since the late Miocene or early Pliocene time to the west; however, this unconformity is deformed by the southern bounding anticline which may project into the Gleneden Beach area. Onshore deformation of late Pleistocene marine terraces by the syncline points to the recent activity of these structures. Possible flexural-slip faults north of the river mouth (Fig. 8) were poorly imaged in the single-channel sparker profile and therefore have uncertain offset or dip. Faults with similar offset to those at Fishing Rock and Fogarty Creek (Fig. 9), identified in single-channel sparker lines between 2 and 6 km offshore, may also be flexural-slip faults (Fig. 10).

**Yaquina and Alsea Bays.** A flight of uplifted Pleistocene marine terraces is preserved north and south of Yaquina Bay (Figs 1 and 11) on the central Oregon coast. These terraces have been differentiated by age using amino acid enantiomeric (D:L) ratios in conjunction with the palaeoecology of fossil shells (Kennedy 1978; Kennedy et al. 1982) and a soil chronosequence (Ticknor et al. 1992; Ticknor 1993; Kelsey et al. 1996). These techniques indicate offset of marine terraces on the inferred Yaquina Bay fault of 75 m down to the south (Ticknor 1993; Kelsey et al. 1996). The fault juxtaposes the 80 ka BP terrace (Qn) north of the bay against the 125 ka BP (Qy) terrace south of the bay (Fig. 11; Kelsey et al. 1996). This offset yields a slip rate of 0.6 mm/year. The continuation of the Yaquina Bay fault to the east was mapped by Snively (1976), giving an ENE fault orientation (Kelsey et al. 1996). All core locations of buried marshes identified by Peterson & Priest (1995) are located on the south or downthrown side of the Yaquina Bay fault.

Similar studies at Alsea Bay (Ticknor et al. 1992; Ticknor 1993; Kelsey et al. 1996) show that Quaternary faults strike generally N–S. The N–S striking Waldport fault zone vertically
displaces terrace platforms down to the east (Fig. 11), with cumulative offset apparently greatest at Alsea Bay, suggesting a structural origin for this embayment (Kelsey et al. 1996). All rapid subsidence sites are on the downthrown side of the Waldport fault zone.

Kelsey et al. (1996) concluded, from the evidence of Pleistocene terrace deformation, that both Yaquina and Alsea Bays are downwarped and structurally controlled by faults. Offshore data neither support nor refute the onshore terrace evidence.

South Slough. Many active structures with N–S trends on the southern Oregon coast and shelf, where the deformation front is closer to the coastline, are interpreted to be part of accretionary prism-related deformation (Fig. 12). One example is the South Slough syncline which deforms Quaternary sediments southwest of Coos Bay on the southern Oregon coast (Fig. 12; Nelson 1987; Peterson & Darienzo 1989; Kelsey 1990; McInelly & Kelsey 1990) and may have produced multiple buried peats as an independent local structure (e.g. Nelson & Personius 1996). The syncline has been traced onto the shelf on seismic reflection profiles (Fig. 12; Goldfinger et al. 1992a; Goldfinger 1994). Both offshore and onshore deformation suggests that many faults are flexural-slip faults bounding active folds, such as the South Slough syncline, with fault slip parallel to bedding planes (McInelly & Kelsey 1990; Goldfinger 1994).

Coquille River. The Coquille fault (Fig. 12; Clarke et al. 1985; Goldfinger 1994) comes onshore just south of the Coquille River mouth, where it deforms Pleistocene marine terraces (McInelly & Kelsey 1990). The Whisky Run (80 ka BP) platform descends from an altitude of 35 m at Cape Arago to sea level just north of the Coquille River, deformed by the Pioneer anticline (Fig. 12; McInelly & Kelsey 1990). The terrace abruptly gains altitude to 18 m above sea level just south of the Coquille River at Coquille Point (Fig. 12). This altitude change is accompanied by a change in dip of platforms from southwest north of the river to west or seaward south of the Coquille River (Fig. 12; McInelly & Kelsey 1990). The Whisky Run platform is tilted slightly landward at Coquille Point, possibly resulting from deformation by the Coquille fault. The terrace elevation descends once again south of Coquille Point to reach sea level c. 10 km to the south (McInelly & Kelsey 1990). Fold trends in Tertiary and Mesozoic formations underlying the Whisky Run wavecut plat-
et al. (1985) suggested the fault downdrops Pleistocene sediments to the northeast on the innermost shelf as observed in the deformed onshore marine terraces, but other seismic data across the fault indicate a fairly symmetrical ridge.

Humboldt Bay. Two large thrust fault systems, the Little Salmon fault and the Mad River fault zone, deform Holocene sediments in the Humboldt Bay region of northern California (Fig. 1), and are interpreted to be part of the onshore expression of the southern Cascadia accretionary prism (Clarke & Carver 1992). The Freshwater syncline lies between these two fault systems within Humboldt Bay and Mad River Slough, and has produced Holocene subsidence resulting in stacked buried marsh and forest horizons (Clarke & Carver 1992). The most recent subsidence event is dated at 250–300 radiocarbon a BP, contemporaneous with the most recent event recorded throughout much of the subduction zone.

Locations with inconclusive evidence for Quaternary subsidence

Other abrupt subsidence sites are characterized by inconclusive evidence or no evidence of Quaternary deformation, in the form of crustal downwarping. Marsh burials have been found in Neah Bay (Waatch River) and the Pysht River on the northernmost Olympic Peninsula (Fig. 1; Atwater 1992). No active faults have been mapped and directly linked to evidence of rapid subsidence in Neah Bay. Buried marshes at Pysht River (Atwater 1992) lie on the downthrown side of a high-angle fault through Tertiary formations (Gower 1960; Tabor & Cady 1978). This fault is mapped parallel to the river (NE), downthrown to the west, and projects offshore to a similar fault in the Strait of Juan de Fuca which deforms an acoustic unit of Holocene age (Wagner & Tomson 1987). A second fault trends parallel to the coast (WNW) and projects into the bay. Buried marshes may lie on the upthrown side of this fault, but there is no evidence that it is recently active. High-resolution seismic profiles along the lower reaches of the Columbia River (Ryan & Stevenson 1995) indicate possible evidence of Quaternary faulting, including the NE-trending Fern Hill fault, which offsets Miocene Astoria Formation onshore (Niem & Niem 1985). These faults have not yet been linked directly to locations of rapid marsh burial or liquefaction in the Columbia River (Atwater 1992, 1994; Obermeier 1995). Niem & Niem (1985) also mapped a WNW-trending syncline through Youngs Bay (Fig. 1), south of Astoria, where a drowned forest and rapidly buried marshes have been identified (Peterson et al. 1997, C. D. Peterson, pers. comm., 1997). Quaternary deformation across this structure remains unproven. An older syncline (deforming late Miocene and possibly Pliocene strata but with no conclusive evidence of Quaternary deformation) has been mapped on the outer shelf opposite Nestucca Bay (Fig. 1). No evidence of Pleistocene deformation has been reported onshore, but it has been suggested that Cape Kiwanda, the headland to the north of the bay which is composed of Miocene Astoria Formation and Smugglers Cove Formation, may be a structural high (Parker 1990). Evidence of Quaternary downwarping at Vancouver Island sites, the Necanicum River, and Salmon River also remains inconclusive, to judge from the available data.

Marsh burial located near structural uplifts

Two possible exceptions to the hypothesis that buried marshes lie within tectonic downwarps are the Sixes River, southern Oregon, and the Copalis River, central Washington (Fig. 1). Kelsey (1990) mapped an E–W trending anticline, which deforms Pleistocene terraces, just north of Cape Blanco and coincident with the lower reaches of the Sixes River (Fig. 1). Buried marshes and tsunami sands have been identified on the southern limb of this anticline (Kelsey et al. 1993, 1998) in a cutoff meander of the Sixes River. Further examination of marshes in a N–S transect across the southern flank of the anticline may reveal differential subsidence (H. M. Kelsey, pers. comm., 1997). This apparent anomaly of local uplift and regional subsidence could be explained by subsidence being the net result of local uplift and regional subsidence, by the anticline not being triggered by every subduction zone earthquake, or by the fact that the anticline was active in Pleistocene but not during Holocene time. A second possible exception is an ENE-trending ridge (Langley ridge) located 5 km south of the Copalis River. Deformation is interpreted as anticlinal folding and diffuse faulting above a possibly N-dipping blind thrust fault (McCrory 1996), with buried marshes at the Copalis River on the upthrown or north side of this fault. In addition to buried marshes, Atwater (1992)
identified liquefaction evidence from 900–1300 years ago at Copalis River with no indication of accompanying tectonic subsidence. This event could be attributed to movement on a local crustal thrust fault such as that underlying Langley Ridge. The dip of a blind fault is often difficult to determine from geomorphology and surface faulting, and this thrust fault may, in fact, dip to the south, with the Copalis River marshes lying on the downthrown side of a fault. Alternatively, a syncline to the north of this ridge may coincide with the Copalis River. This is supported by N–S single-channel and multichannel seismic profiles which indicate a series of closely spaced approximately E–W trending ridges and intervening synclines on the continental shelf. The Copalis River buried marshes may in fact lie within one of these synclines rather than being associated with the Langley Ridge structure 5 km to the south.

**Discussion**

**Implications for the Cascadia subduction zone earthquake record**

The record of prehistoric subduction earthquakes on the Cascadia subduction zone, in the form of rapidly buried marshes, documents sudden submergence, inundation of coastal lowlands, and burial of the former land surface. Correlation of coseismic events between coastal bays, based on radiocarbon ages and dendrochronology, has allowed rupture lengths, magnitudes, and recurrence intervals of prehistoric Cascadia earthquakes to be proposed. In addition, estimates of amounts of coastal subsidence can be used to approximate the position of the rupture zone and earthquake magnitude using elastic dislocation modelling. The possible non-tectonic origin of some submergence events should, however, be considered when assessing potential earthquake hazards.

In this study, it has been demonstrated that many abruptly buried marsh locations can be linked to Quaternary structures (synclines and downdropped side of faults) which produce downwarping. The influence of upper-plate crustal deformation on the prehistoric earthquake record may lead to inaccuracies in calculations of magnitudes and recurrence intervals if based on the Holocene stratigraphy of coastal bays. Evidence of Quaternary deformation off-shore is equivocal in some cases, but the use of offshore datasets and coastal exposures together has increased the probability of identifying recent activity. Only two possible examples of rapid subsidence coincident with crustal uplift were identified. However, these apparent anomalies could be explained by active upper-plate structures not deforming during every subduction event.

**Localised upper-plate deformation at other subduction zones.** Localized upper-plate deformation has been documented at subduction margins world-wide, with deformation both synchronous with and independent of subduction zone events. Examples include the Hikurangi margin of New Zealand (Berryman et al. 1989; Cashman & Kelsey 1990; Berryman 1993a, b), the Alaskan margin (Plafker 1972), the Nankai forearc of SW Japan (Maemoku 1988a, b; Maemoku & Tsubono 1990; Sugiyama 1994), and the Huon peninsula of Papua New Guinea (Pandolfi et al. 1994). Holocene terraces along the coastal Hikurangi margin off eastern North Island, New Zealand, are uplifted by movement on steep reverse faults of the onshore accretionary prism (Berryman et al. 1989), with clustering of terrace ages along the coast. Stratigraphic and ecological studies of Holocene terrace sediments on the Mahia Peninsula reveal that sedimentation was progradational between events, implying a lack of interseismic subsidence that would be expected with a subduction earthquake cycle and supporting the formation of Holocene coseismically uplifted terraces by local crustal structures (Berryman et al. 1997). Other earthquakes within the accretionary prism include the 1931 Hawkes Bay earthquake (M_s 7.8), caused by a fault cutting up from the megathrust (Hull 1990), and the 1855 Wairarapa earthquake, which may have originated on the megathrust and propagated into the upper plate along a blind thrust fault (Darby & Beanland 1992). In Alaska, significant deviations from the regional subsidence or uplift patterns during the 1964 earthquake (up to 12 m uplift across the Patton Bay fault on Montague Island, relative to a regional 2–4 m of uplift) were associated with movement on crustal faults contemporaneous with the subduction zone earthquake (Plafker 1969, 1972). Along the Nankai margin of Japan, two types of subduction earthquake have been inferred from coseismically uplifted terraces (Fig. 13; Sugiyama 1994). Subduction events where no permanent crustal deformation and therefore no uplifted terrace preservation occurred are known as Taisho type events (T, Fig. 13). Preserved uplifted terraces resulting from the triggering of crustal deformation are known as Genroku type events (G, Fig. 13).
Fig. 13. Illustration of the resulting record of Taisho and Genroku subduction earthquakes on the Nankai margin (after Sugiyama 1994; reproduced by kind permission of Geofisica International). No permanent inelastic crustal deformation occurs during Taisho events (T1–5); coseismic deformation is recovered in the interseismic period leaving no permanent record of the earthquake. Genroku events (G1, G2) involve local faulting or other inelastic crustal deformation leading to preservation of an uplifted bench. It should be noted that the coseismic and interseismic vertical motions are opposite to those expected on much of the Cascadia coastline.

Effects of local crustal deformation on subsidence records. Recorded coseismic subsidence is the net result of regional elastic strain release from a subduction zone earthquake and local crustal deformation (permanent and/or elastic), assuming a tectonic origin for subsidence. For each event, subsidence could result from strain release on the plate boundary or on local structures, or a combination of the two. The contributions of each cannot be determined for prehistoric events, although the amount and pattern of subsidence at each location may suggest a particular mechanism. The apparent rupture length (and hence magnitude), amount of subsidence, and timing of coseismic events can be specifically affected in the following ways.

1. Triggering of local crustal faults beyond (along strike) the subduction rupture zone, as the release of the elastic load on the upper plate in one area causes loading in other areas, thereby increasing the apparent rupture length and magnitude (e.g. LS2 and SZE1 in Fig. 14). Similar patterns occurred during the Landers earthquake, where a sequence of delayed ruptures occurred within the fault zone (Sieh et al. 1993; Wald & Heaton 1994; Spotila & Sieh 1995), and more distant earthquakes were also triggered following the mainshock (Hill et al. 1993), although these events lacked surface rupture and geodetic change.

2. The amount of subsidence per event at each location is dependent not only on the magnitude of the subduction zone earthquake and position of the rupture zone, but also on the amount of localized upper-plate deformation accompanying the earthquake. The amount of subsidence during a given earthquake cannot be used for elastic dislocation modelling of the locked zone.

3. Local crustal faults may move independently of subduction zone earthquakes and produce anomalous local coseismic subsidence and marsh burial (LS1 of Fig. 14). Correlation of subsidence events from site to site is dependent on age control with sufficient precision to distinguish such events. The majority of radiocarbon ages from marsh burials have large error bars, of the order of ±50 years to hundreds of years (e.g. Atwater 1992; Atwater et al. 1995), with errors often larger than the suggested recurrence intervals. More recently, AMS and high-precision radiocarbon and dendrochronology ages in some locations have significantly reduced errors to ±10–20 years or even to within a year or season (Nelson et al. 1995; Jacoby et al. 1997; Yamaguchi et al. 1997), but suitable material for such precise dating techniques is often unavailable (Nelson et al. 1996b). The most abundant high-precision data are available for the most recent subsidence event, which is dated within a few decades of AD 1700, and is consistent with evidence for a remote tsunami in Japan at that time (Satake et al. 1996). Older
events are less accurately dated. Where error bars are large, it is impossible to distinguish between regional subsidence events and locally anomalous events which may have resulted from independent deformation on crustal structures. These anomalous events may be wrongly correlated with regional coseismic subsidence events, producing an inaccurate picture of the earthquake rupture zone. In addition, subduction earthquake recurrence intervals may be underestimated if independent local subsidence events contribute to the marsh stratigraphy.

Synchronous v. independent movement of local structures and subduction earthquakes. It is unknown, in the absence of historic subduction zone earthquakes and little upper-plate seismicity, whether local crustal structures in Cascadia are triggered by subduction zone earthquakes or operate independently, or both. If movement on crustal structures is always synchronous with and triggered by subduction events, estimates of the subduction earthquake recurrence interval will be unaffected but magnitude calculations may be inaccurate. If these structures operate independently, deforming both during and between subduction events, both the magnitude and recurrence interval of subduction zone events will be affected.

If patterns of strain release are similar to those of the Alaskan and Nankai subduction zones, we might expect crustal structures to be triggered by slip on the megathrust (Plafker 1969, 1972; Sugiyama 1994). Minimal historic seismicity in the coastal and shelf region supports the hypothesis that these structures are predominantly triggered by subduction zone earthquakes, which also lack seismicity. Buried marshes similar in age to regional subsidence events have been attributed to upper-plate structures in South Slough, southern Oregon coast, and Humboldt Bay, northern California coast, with little visible evidence of significant rapid subsidence in other bays in this region, such as the Siuslaw River (Clarke & Carver 1992; Nelson 1992; Nelson & Personius 1996; Nelson et al. 1996a). If rapid subsidence is not regionally extensive, these are examples of crustal structures that were triggered by subduction zone events. Regional subsidence in this area may be small and only detectable by biostratigraphic investigations (compare with Mathewes & Clague (1994)), and pronounced sudden subsidence may only be recorded where local structures were triggered. Upper-plate structures are likely to have longer recurrence intervals than subduction zone earthquakes and may not be triggered by every subduction event.

Local structures may also produce tectonic subsidence independently of subduction events. Independent coseismic subsidence has been suggested as a likely cause of marsh burials on the southern Oregon and northern California coasts (Nelson 1992; Nelson & Personius 1996).

Preservation of buried marshes. The sequence of Cascadia buried marshes indicates net submergence of the land or net relative sea level rise of 2–5 m in the last 2000–4000 years. If the earthquake strain cycle were completely elastic and no other factors were involved, coseismic subsidence and interseismic uplift would cancel out and no buried marshes would be preserved. This argument is used for the Nankai subduction zone, where coseismically uplifted terraces are only preserved permanently when synchronous upper-plate uplift occurs (Sugiyama 1994). If similar patterns of deformation to those at Nankai occur along the Cascadia subduction zone, permanent deformation by local structures may help to preserve marsh burial. Not all subduction zone earthquakes would be recorded (e.g. SZE2 in Fig. 14) and subduction zone earthquakes would appear to be less frequent with longer recurrence intervals.

One major difference between the Nankai and Cascadia coastlines is the sense of coseismic motion: Nankai experiences uplift whereas Cascadia experiences subsidence. Therefore, preservation of buried marshes in Cascadia may also be influenced by the following non-tectonic factors, producing relative sea-level rise: (1) late Holocene eustatic sea-level rise; (2) isostatic forebulge collapse following the last glacial maximum; (3) compaction; or (4) changes in the geometry of coastal estuaries (this could cause relative sea-level rise or fall). These factors would not contribute to the preservation of uplifted terraces in Nankai, unless sea level was falling in late Holocene time, or this region experienced some form of isostatic uplift. Buried marsh preservation in Cascadia could also be attributed to regional tectonic subsidence resulting from, for example, subduction erosion. The rates and effects of these factors in late Holocene time are poorly known and therefore their contributions to the preservation of buried marshes can only be approximated. The rate of late Holocene global eustatic sea-level rise is hotly debated, with some estimates pointing to negligible rise during the last 5000 years (P. Clark, W. R. Peltier, pers. comm., 1997), very low rates (Clark & Lingle 1979; Bard et al. 1996), or a value which is currently very difficult to separate from local and regional factors.
including isostatic and tectonic factors, which dominate relative sea-level rise (Bloom & Yonekura 1990; Nelson et al. 1996b). Estimates of forebulge collapse on much of the Cascadia margin associated with isostatic re-equilibration following the last glacial maximum (LGM) are given by the models of Peltier (1996). Subsidence rates (or relative sea-level rise) for much of the Washington and Oregon coastline are estimated as 0–1 mm/year, with a maximum on the northern Oregon coast (M2 model of Peltier (1996)). However, the northern Olympic Peninsula and Vancouver Island, which underlay the Cordilleran ice sheet during the LGM, should be experiencing isostatic rebound. Tectonic subsidence may indeed dominate the preservation of buried marshes, but until other variables are better resolved, this hypothesis remains untested.

Quantitative subsidence calculations. In general, subsidence patterns measured along the Cascadia subduction zone appear to be fairly consistent, with subsidence of 0.5–2 m for each burial event. Very large differences in the amount of subsidence per event, which might indicate localized subsidence contributions, have not been observed, as pointed out by Clague (1997). The large deviation in uplift magnitude recorded across the Patton Bay fault in Alaska, is not observed in subsidence on the Cascadia margin. However, the Patton Bay fault, within the coseismically uplifted zone, is within the accretionary prism and close to the deformation front, and therefore might be expected to experience more pronounced deformation. Measurements of Cascadia subsidence are invariably imprecise because biostratigraphic markers such as diatoms and plant assemblages have large vertical water depth ranges, but fairly large differences in subsidence can be detected (Atwater & Hemphill-Haley 1996). Small variations in measured subsidence through careful lithostratigraphic and biostratigraphic studies, such as those of Long & Shennan (1994), Nelson et al. (1996b), and Shennan et al. (1996), may eventually allow the contribution of local structures to the subsidence record to be determined.

The estuarine stratigraphy at many Cascadia sites is strikingly similar to that observed at passive margin sites (Long & Shennan 1994), where a coseismic origin is unlikely. Some Cascadia subsidence events may therefore have non-seismic origins such as natural succession of intertidal environments from local changes in sea level, sedimentation rates, and ocean currents (Long & Shennan 1994; Nelson et al. 1996a, b).

N–S compression

Most active structures on the inner shelf and coast in the Cascadia subduction zone are characterized by roughly E–W trends, in contrast to the predominantly N–S to NW–SE trends which result from plate convergence on the continental slope. We agree with Snavely (1987) and Goldfinger et al. (1992b) that this landward region of the forearc is under N–S compression, which is in agreement with regional N–S compression throughout the continental northwestern USA derived from late Tertiary upper-plate fault orientations, earthquake focal mechanisms, and borehole breakouts (Werner et al. 1990; Zoback & Zoback 1989). The regional N–S compressional stress field extends onto the middle to outer shelf in Washington and much of Oregon. In contrast, the southern Oregon and northern California shelf and coastal region are within the active accretionary prism, and deformation is in response to plate convergence leading to structures with N–S to NW–SE trends. Wang et al. (1995) suggested that the NE-directed strain accumulation caused by plate convergence can be considered a time-dependent local perturbation superimposed on the regional N–S compressive stress field, and thus the regional stress field and cyclic loading may coexist. The transition from regional N–S compression to predominantly plate convergence driven compression represents a significant structural domain boundary. This transition may act as a backstop and may be related to the long-term average position of the downdip end of the seismogenic locked zone. Despite the apparent independence of upper-plate structures from subduction zone deformation, it seems likely that these structures could be triggered by rupture of the subduction zone. Independent fault movement in response to regional compression is also possible and therefore these structures pose independent seismic hazards.

Conclusions

Evidence of Quaternary deformation on the Cascadia coast and inner shelf is widespread, with crustal downwarping or fault offset coincident with many coastal lowlands. Rapidly buried marshes at these locations may be due to elastic strain release on the subduction megathrust, downwarping or fault displacement on upper-plate crustal structures, or both. Calculations of Cascadia subduction zone earthquake recurrence intervals, rupture zones, and magnitudes based on correlations of marsh burial
events between sites may be complicated by the possibility of localized crustal fault movements and fold growth, in addition to non-seismic origins of the observed stratigraphy. The earthquake record is likely to be more difficult to resolve with the interaction of these multiple factors (Fig. 14). Prehistoric subduction zone earthquakes may have been of lower magnitude than previously estimated. Recurrence intervals for such earthquakes may be overestimated, if some events are not preserved as a result of little permanent deformation, or underestimated, if anomalous local subsidence events are wrongly linked to similar-age regional events along the margin. Higher-resolution records of marsh chronology and estimates of subsidence may eventually lead to the separation of regional and local factors. Meanwhile, the use of these records for modelling of the subduction earthquake cycle and prediction of prehistoric earthquake rupture zones should be undertaken with caution. The inner shelf and coastal structures are consistent with the regional N–S compressional stress field and inconsistent with subduction-driven compression. Despite low seismicity, these crustal faults may be seismic and pose significant shaking and ground deformation hazards to the coastal communities.

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