Detrital-zircon fission-track ages for the “Hoh Formation”: Implications for late Cenozoic evolution of the Cascadia subduction wedge

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ABSTRACT

We report new fission-track (FT) ages for detrital zircons for 34 sandstone samples and 2 volcanic ash beds from the “Hoh Formation,” exposed along the western side of the Olympic Mountains of western Washington State. The “Hoh Formation” is now formally known as the coastal unit of the Olympic Structural Complex, or Coastal OSC for short. About 35 zircons were dated per sample. The fission-track grain-age (FTGA) distributions are all strongly discordant; grain ages range from 10 to older than 100 Ma. Low vitrinite-refection values, short etch times for the zircons, and a broad range of grain ages indicate that the zircon FT ages are unreset and thus preserve information about cooling events in the source region for these sediments. Five areas were sampled repeatedly and yield similar FTGA distributions, demonstrating that sampling errors are not a problem. We show that almost all of the samples contain a well-defined young component that was probably derived from a contemporaneous active volcanic source, presumably the adjacent Cascadia arc. Binomial peak-fitting was used to estimate the FT minimum age, which is the age of the youngest concordant fraction of zircon FT grain ages in a FTGA distribution. In most cases, minimum ages are similar to fossil ages where available. This result supports our contention that zircon FT minimum ages from volcaniclastic sandstones commonly can be used as a proxy for depositional age.

Our zircon FT minimum ages indicate that the Coastal OSC is made up mainly of lower Miocene (ca. 24 to 16 Ma) sedimentary rocks. We use these age data, together with other geologic constraints, to reconstruct a tectonic history. Sedimentary rocks of the Coastal OSC were derived from a mixed-source region that included an active volcanic arc and also older units, including Cretaceous metamorphic rocks, probably located in the Omineca crystalline belt in the Canadian Rockies. The upper part of the Clallam Formation, located on the northern side of the Olympic Peninsula, appears to be a remnant of the sedimentary system that fed the Coastal OSC. The sediments that formed the Coastal OSC were initially deposited seaward of the Cascadia trench, at water depths of >2000 m. This debris was deposited seaward of the Cascadia trench, at water depths of >2000 m, and subsequently accreted beneath the frontal 50 to 100 km of the wedge. Owing to continued accretion at the front of the wedge, and erosion of the forearc high in back of the wedge, these lower Miocene sediments were moved rearward within the Cascadia subduction wedge. A simple relationship based on the cross-sectional area of the wedge and a steady accretion flux indicates that it would have taken ~22 m.y. for the Coastal OSC to reach its present position 140 km landward of the toe of the wedge. This estimate is in good agreement with the unit’s early Miocene age.

Keywords: Cascadia subduction wedge, Olympic Mountains, Coastal OSC, Hoh Formation, zircon fission-track dating.

INTRODUCTION

The Olympic Mountains of northwest Washington State (Fig. 1) mark the first part of the Cascadia forearc to emerge above sea level, starting at ca. 15 Ma (Brandon and Vance, 1992; Brandon et al., 1998). Uplift and erosion of the forearc high in this area provide a deep window into the subduction wedge. Tabor and Cady (1978a, 1978b) used the informal name Olympic core for these rocks. Brandon and Vance (1992) suggested Olympic subduction complex. To be consistent with Stratigraphic Code (Salvador, 1994), we designate here the formal name Olympic Structural Complex (OSC). This name is a direct replacement for the previous informal names. Following Tabor and Cady (1978a, 1978b), the OSC refers to the imbricated assemblage of turbidite sandstone, siltstone, and lesser igneous rocks that structurally underlie the Crescent terrane and the Calawah and Hurricane Ridge faults, and their lateral equivalents (Figs. 1 and 2). These rocks are mainly Eocene to middle Miocene in age, but note there is a small enigmatic slice of Mesozoic rocks exposed in the northwest corner of the Olympic Peninsula. The OSC could be extended to include all equivalent rocks within the Cascadia subduction wedge, but this is generally not done given that the Olympic Mountains represent the only exposure of the Cascadia wedge north of the California border.

Tabor and Cady (1978a, 1978b) mapped five informal lithic assemblages within the OSC. Brandon and Vance (1992) reorganized...
The Cascadia subduction wedge (Fig. 1) is a doubly vergent wedge (Willett et al., 1993) formed by 35 m.y. of subduction of the Juan de Fuca plate beneath the Cascadia margin (e.g., Davis and Hyndman, 1989; Aalto et al., 1995, 1998). The active wedge is 200 to 250 km wide, bounded to the west by the Cascadia trench and to the east by a forearc low, marked by the Willamette Valley, Puget Sound, and Georgia Straits. The forearc high, which corresponds to the Oregon-Washington Coast
Ranges and the Insular Ranges of Vancouver Island, marks the transition between the pro- and retroside of this doubly vergent system.

Accretion over the past 35 m.y. has allowed the wedge to grow to its present large size (e.g., Rau, 1973, 1975; Tabor and Cady, 1978a, 1978b; Dickinson and Seely, 1979; Brandon and Vance, 1992; Brandon et al., 1998). The seaward deformation front of the wedge is currently ~140 km west of the present coastline of the Olympic Peninsula, and seismic data indicate that the 2–3 km thickness of sediment carried into the subduction zone on the Juan de Fuca plate gets thickened to 20 km beneath the Washington coast and ~35 km beneath the central part of the Olympic Mountains (Clowes et al., 1987; Brandon et al., 1998; Parsons et al., 1998; Pazzaglia and Brandon, 2001). The modern convergence rate along this part of the Cascadia subduction zone is 36 mm/yr (DeMets et al., 1990; DeMets and Dixon, 1999), and virtually all sediment on the Juan de Fuca plate appears to be incorporated by accretion into the Cascadia subduction wedge (Davis and Hyndman, 1989; Davis et al., 1990; Pazzaglia and Brandon, 2001; Batt et al., 2001).

Along most of its length, the Cascadia subduction wedge includes a relatively coherent tectonic sequence, called by some the “Coast Range terrane” (or “Siletz terrane” by others), made up of early Eocene oceanic crust and overlying marine sedimentary rocks. The base of the Coast Range terrane is exposed in the Olympics and is defined by a major thrust fault, locally called the Hurricane Ridge fault or Calawah fault (Fig. 2). The Coast Range terrane can be viewed as a structural lid that overlies the accretionary part of the wedge (Clowes et al., 1987; Brandon et al., 1998). We consider the lid to be part of the active wedge, because it shows clear evidence of uplift and broad folding associated with the maintenance of critical taper on the two sides of the doubly vergent wedge (see discussion in Pazzaglia and Brandon, 2001).

The focus here is on the Coastal OSC, which is the youngest and westernmost unit exposed in the Olympic Mountains (Fig. 2). It consists mainly of a monotonous sequence of sandstones and shales, typically well bedded but locally chaotic, with rare interspersed lenses and blocks of volcanic rock. Clastic rocks are mostly turbidites, and volcanic rocks are geochemically similar to the thick basaltic basement in the overlying Coast Range terrane (Applegate and Brandon, 1989). Microfossils in the Coastal OSC are overwhelmingly early Miocene in age and record deposition in lower-bathyal water depths, between ~2000 and 4000 m (Rau, 1975, 1979; Ingle, 1980). There is a small number of younger localities that indicates middle or late Miocene ages (seven reported in Rau, 1975) and a few others that indicate Eocene ages (four reported in Rau, 1979). Definitive Oligocene localities have not been recognized.

The more eastern parts of the Olympic subduction complex commonly show prehnite-pumpellyite facies metamorphism and a spaced pressure-solution cleavage (Tabor and Cady, 1978a, 1978b; Brandon and Calderwood, 1990). In contrast, clastic rocks of the Coastal OSC are typically unmetamorphosed, although zeolites are locally present (Stewart, 1974). Basaltic blocks do show static greenschist assemblages, probably related to seafloor metamorphism, which would have occurred prior to incorporation of the blocks into the mélangé units. Cleavage fabrics are also notably absent, except for scaly fabrics locally developed in mélangé units.

The Coastal OSC shows clear evidence of northeast-southwest shortening, in the direction of convergence. Bedding generally strikes to the northwest-southeast and is commonly steep and locally overturned. Where recognized, folds are typically overturned to the southwest. The section shows widespread evidence of repetition by thrust faults (Rau, 1975, 1979; Orange, 1990).

Mélangé and broken formation (sensu Hsu, 1968) are expected, given the subduction-zone setting, but these deformational styles are observed in less than ~20% of the Coastal OSC. In fact, mélangé is even less common (<10%) in the other parts of the subduction complex (Tabor et al., 1970; Tabor and Cady, 1978a). These observations contrast with the large fraction of mélangé found in Miocene parts of the Cascadia wedge exposed in the Franciscan Complex in northern California (see Fig. 1; King Range and False Cape terranes of McLaughlin et al., 1982, 1994; Aalto et al. [1995]).

Local bodies of mélangé in the Coastal OSC have a matrix of intensely “scaly” siltstone with blocks of sandstone, conglomerate, and volcanic rock (Weissenborn and Snively, 1968; Rau, 1970, 1975, 1979; Rau and Grocock, 1974; Orange, 1990; Orange et al.,
Published maps generally lump rare exposures of broken formation together with the more widespread coherent stratigraphic units. Rau (1979) and Snively and Kvenvolden (1989) recognized local areas where the more coherent parts of the Coastal OSC appear to be resting depositionally on broken formation. Modern subduction wedges are commonly mantled by small “trench-slope basins” (Moore and Karig, 1976) that accumulate in local depressions in the surface of the wedge. The coherent sequences in the Coastal OSC may represent trench-slope basins, deposited after the underlying section was accreted.

The other alternative is that the chaotic units record mass wasting. Mass wasting is now widely recognized at sediment-rich subduction zones, like Cascadia (e.g., Goldfinger et al., 2000). Mass-wasting deposits commonly reach the trench (e.g., Davis and Hyndman, 1989), where they become depositionally interfingered with trench-basin strata. Thus, stratigraphic relationships are currently not sufficient to judge whether the coherent units in the Coastal OSC were deposited seaward or landward of the toe of the wedge.

The paleobathymetric evidence provided by benthic foraminifera provide a key constraint in that they record deposition in water deeper than ~2000 m. Rau (1975, 1979) recognized that deep-water benthic foraminifera were very common in the Coastal OSC, but he preferred a conservative estimate of >200 m for water depth. One of the reasons is that the foraminiferal assemblages were a mixture of both deep- and shallow-water taxa. Down-slope transport is now recognized as the reason for mixed depth provenance of benthic foraminifera in turbidite sequences. Thus, paleodepth is now defined by using the deepest-water taxa in the assemblage (Ingle, 1980).

We have applied the paleobathymetry biofacies of Ingle (1980) to define paleobathymetry by using the benthic foraminifera reports by Rau (1975, 1979) for the Coastal OSC. The conclusion is that the unit was deposited in the lower-bathyal biofacies, indicating water depths of 2000 to 4000 m. The diagnostic taxa are Cibicides pseudoungerianus evolutus (Cushman and Hobson) and Gyroidina solandii (d’Orbigny), which are very common in all of Rau’s samples. Less common diagnostic taxa are Melonis pompilioides (Fichtel and Moll) (referred to by Rau, 1979, by an older name: Nonion pompilioides), and Plectofrondicularia californica (Cushman and Stewart).

At present, the surface of the Cascadia trench basin seaward of the wedge is at a water depth between 2000 to 2500 m. The relatively shallow depth is due to the young age of the Juan de Fuca plate, which is ca. 8 Ma at the trench. The age of the plate at the trench has remained relatively constant at ~8 m.y. over the past ~35 m.y. (Wilson, 1988), so the paleodepth of the trench basin would not have been much different than the modern. Thus, we conclude that most of the Coastal OSC was probably deposited in the trench basin, seaward of the Cascadia wedge. We recognize, however, that some of the subduction complex may represent slope basins, deposited after accretion. Any slope-basin sequences must have been deposited soon after accretion in order to account for the foraminiferal evidence for initial deposition at depths of >2000 m. Thus, one can consider the age of the Coastal OSC as defining when sedimentary materials first were added to the front of the wedge. This evidence is used in the Discussion to constrain transport times through the wedge.

We have already noted the similarities between the Coastal OSC and correlative accretionary complexes of the False Cape and King Range terranes, exposed on the Pacific Coast in northern California (Fig. 1). These units represent the only other part of the Cascadia subduction wedge where upper Cenozoic accreted rocks are subaerially exposed. The False Cape terrane is composed mainly of scaly argillite and sandstone, and was deformed in early Miocene time (Aalto et al., 1995). The King Range terrane, which contains the youngest rocks of the Coastal belt of the Franciscan Complex, was accreted at the front of the wedge in the middle Miocene, at ca. 15 Ma (McLaughlin et al., 1982, 1994). The King Range terrane is similar to the Coastal OSC in that both include broken formations and mélanges, a general absence of metamorphism, and middle Miocene turbidites that may be accreted sediments initially deposited in the trench (McLaughlin et al., 1982). An important difference is that the False Cape terrane, and perhaps the False Cape terrane as well, appear to have been strongly affected by northward migration of the Mendocino triple junction, which may be largely responsible for driving the present-day uplift and rapid erosion of the wedge in the northern California area (Dumi¢ru, 1991).

Zircon FT analyses for 1264 zircons from 36 samples, all medium- to coarse-grained sandstones, are shown on Tables 1, 2, and 3. Maps showing the distribution of sample localities and zircon FT data are available from the Data Repository.1 Probability density plots for selected examples are shown on Figure 5. Peak-fitting results (Tables 1, 2, and 3) are shown compared to the IUGS time scale of Remane (2000) on Figures 6 and 7.

Zircon separates were prepared by standard techniques using the external-detector method (Wagner and van den Haute, 1992). Both ends of the irradiation package contained fluence monitors, using the SRM 612 U-enriched glass standard, and a mount of Fish Canyon Tuff zircons. Samples were irradiated with thermal neutrons in the TRIGA reactor at Oregon State University (a well-thermalized reactor) using a nominal fluence of 1.0 × 10²¹ neutrons/cm². Following irradiation, the mica detectors were etched in 40 percent hydrofluoric acid for 18 minutes. The effective monitor track density was determined for each sample by using the sample position in the irradiation package to interpolate the densities measured in the two fluence monitors. Fission tracks were counted using oil immersion and a Zeiss microscope at 1250×. Based on 8 analyses, the Zeta value (Hurford and Green, 1983) for R. Stewart is 331.03 ± 6.66 (±1 standard error). Four samples were dated by J. Garver (1990, personal commun.) using methods similar to ours (Garver et al., 1999). His Zeta is 320.67 ± 7.79.

The analyses are organized into three groups. Table 1 consists of 25 samples from the coherent well-stratified turbidite sequences in the Coastal OSC. Five areas, shown as Groups A–E on Table 1 and Figures 3 and 4, were densely sampled, with samples <1 km apart, to check reproducibility of our results. Table 2 includes ten samples collected from sandstone blocks in Coastal OSC mélangé. Table 3 is included for comparative purposes, and consists of samples from adjacent parts of both the Upper and Lower OSC. Five of these samples are from Brandon and Vance (1992), recalculated using BINOMFIT.

EVIDENCE AGAINST THERMAL RESETTLEMENT

Our interpretation of these zircon FT ages depends on whether the ages have been thermally reset. We are particularly concerned because the sediments that formed the Coastal OSC were probably first deposited outboard of the subduction zone, on a young, hot Juan de Fuca plate.
Two observations indicate that resetting did not occur. The first comes from vitrinite reflectance data, which range from 0.48% to 2.08% (Snively and Kvenvolden, 1989; Orange and Underwood, 1995); most values are between 0.5% and 1.29%. Values >1.0% are commonly associated with fault zones where rocks of differing thermal maturation may have been juxtaposed. Arne and Zentilli (1994) showed that reflectance values of 0.7% to 0.9% are diagnostic of thermal histories needed to totally reset apatite FT ages. This prediction is consistent with common partial resetting of apatite FT ages in the Coastal OSC (Brandon et al., 1998). In contrast, resetting of zircon FT ages appears to be associated with vitrinite reflectance values of >~6% (Green et al., 1996; http://www.geotrack.com.au/zfta.htm; Kamp, 2001).

Thus, we would not expect any resetting of the zircon FT ages in the Coastal OSC.

TABLE 1. DETRITAL-ZIRCON FT AGES FROM COHERENTLY BEDDED SANDSTONES, COASTAL OSC

<table>
<thead>
<tr>
<th>Lab number</th>
<th>N</th>
<th>Minimum-age peak</th>
<th>Old peaks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Age</td>
<td>95% Conf. Int.</td>
</tr>
<tr>
<td>Single samples</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>168*</td>
<td>36</td>
<td>11.0</td>
<td>+3.9</td>
</tr>
<tr>
<td>108*</td>
<td>50</td>
<td>13.9</td>
<td>+3.0</td>
</tr>
<tr>
<td>41*</td>
<td>43</td>
<td>16.0</td>
<td>+1.5</td>
</tr>
<tr>
<td>183*</td>
<td>42</td>
<td>16.2</td>
<td>+1.5</td>
</tr>
<tr>
<td>162*</td>
<td>13</td>
<td>18.3</td>
<td>+3.3</td>
</tr>
<tr>
<td>154</td>
<td>51</td>
<td>18.7</td>
<td>+2.2</td>
</tr>
<tr>
<td>151*</td>
<td>50</td>
<td>19.9</td>
<td>+3.6</td>
</tr>
<tr>
<td>143*</td>
<td>27</td>
<td>21.2</td>
<td>18.5</td>
</tr>
<tr>
<td>170*</td>
<td>37</td>
<td>24.9</td>
<td>+2.7</td>
</tr>
<tr>
<td>195</td>
<td>18</td>
<td>25.9</td>
<td>+6.2</td>
</tr>
<tr>
<td>Group A</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>42B*</td>
<td>41</td>
<td>14.2</td>
<td>+2.5</td>
</tr>
<tr>
<td>42B*</td>
<td>27</td>
<td>25.4</td>
<td>+6.3</td>
</tr>
<tr>
<td>COMBINED</td>
<td>68</td>
<td>14.3</td>
<td>+2.3</td>
</tr>
<tr>
<td>(continued for fourth peak)</td>
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<td></td>
<td></td>
</tr>
</tbody>
</table>

Notes: Peak ages and 95% confidence interval were estimated by using the binomial-fit method (Galbraith and Green, 1990). N—total number of dated grains; %—percent of total number of dated grains in an individual peak. COMBINED—data for all samples in group. Bold laboratory numbers correspond to samples illustrated in Figure 5.

*Samples used for comparison of minimum FT ages with fossil ages.

TABLE 2. DETRITAL-ZIRCON FT AGES FROM SANDSTONE BLOCKS IN MÉLANGE, COASTAL OSC

<table>
<thead>
<tr>
<th>Lab number</th>
<th>N</th>
<th>Minimum-age peak</th>
<th>Old peaks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Age</td>
<td>95% Conf. Int.</td>
</tr>
<tr>
<td>Single samples</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>117*</td>
<td>45</td>
<td>14.6</td>
<td>+4.1</td>
</tr>
<tr>
<td>37</td>
<td>46</td>
<td>17.7</td>
<td>+5.0</td>
</tr>
<tr>
<td>176</td>
<td>15</td>
<td>19.9</td>
<td>+6.9</td>
</tr>
<tr>
<td>118*</td>
<td>50</td>
<td>20.4</td>
<td>+4.9</td>
</tr>
<tr>
<td>140</td>
<td>50</td>
<td>22.0</td>
<td>+2.7</td>
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<tr>
<td>175</td>
<td>41</td>
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<td>178</td>
<td>17</td>
<td>32.3</td>
<td>+9.0</td>
</tr>
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<td>173</td>
<td>18</td>
<td>35.4</td>
<td>+7.5</td>
</tr>
<tr>
<td>191</td>
<td>37</td>
<td>36.2</td>
<td>+17.1</td>
</tr>
<tr>
<td>152</td>
<td>23</td>
<td>37.5</td>
<td>+18.3</td>
</tr>
</tbody>
</table>

Notes: See Table 1.
**TABLE 3. DETRITAL ZIRCON FT AGES FROM ADJACENT PARTS OF THE OLYMPIC SUBDUCTION COMPLEX**

<table>
<thead>
<tr>
<th>Lab number</th>
<th>N</th>
<th>Minimum-age peak</th>
<th>Old peaks</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Age</td>
<td>95% Conf. Int.</td>
<td>%</td>
</tr>
<tr>
<td>Lower OSC</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ZD6</td>
<td>50</td>
<td>18.5</td>
<td>2.1</td>
</tr>
<tr>
<td>ZD50</td>
<td>25</td>
<td>26.5</td>
<td>3.8</td>
</tr>
<tr>
<td>S3</td>
<td>51</td>
<td>19.2</td>
<td>2.1</td>
</tr>
<tr>
<td>Upper OSC</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ZD22</td>
<td>50</td>
<td>30.9</td>
<td>2.0</td>
</tr>
<tr>
<td>ZD38</td>
<td>61</td>
<td>43.1</td>
<td>3.9</td>
</tr>
<tr>
<td>ZD44</td>
<td>25</td>
<td>47.8</td>
<td>3.8</td>
</tr>
</tbody>
</table>

Notes: Samples with “ZD” prefix are from Brandon and Vance (1992) and are recalculated here by using binomial peak-fitting method. See Table 1 for other details.

**Figure 3. Simplified geologic map of the Coastal OSC after Rau (1975, 1979), showing sample locations by laboratory number (Tables 1 to 3).** The Coastal OSC is divided into coherent turbidite sequences (light gray) and mélangé sequences (dark gray). White circles—sandstone samples from coherent sequences of the Coastal OSC; gray circles—sandstone samples from blocks in mélangé. Italic labels mark groups A to E, which represent areas where a coherent sequence was sampled more than once. The Upper OSC and Lower OSC (Fig. 2) are undifferentiated on this map but lie to the east of the thrust fault bounding the east side of the Coastal OSC. Comparative samples from those units are reported in Table 3 and marked on the map with white squares. ZD6 and ZD50 are from the Lower OSC, and ZD38, ZD44, and ZD22 are from the Upper OSC.

Mountains, in the vicinity of Mount Olympus (Fig. 2). In that area, resetting is marked by a reduction in the range of grain ages and by an increase in the amount of time needed to etch fission tracks. However, temperatures in that area failed to bring the zircons to a common reset age (i.e., the “reset” samples all failed the χ² test), indicating that zircons had heterogeneous annealing properties during resetting.

Grain-to-grain variations in radiation damage are the most likely cause for this heterogeneous annealing. Most of the radiation damage in zircon is produced not by fission decay of U, but by alpha decay of U and Th. Alpha decay produces much smaller tracks, but occurs at a much higher rate relative to fission decay. Experimental work shows that the time needed to etch and reveal fission tracks increases as the amount of alpha damage decreases and that alpha damage and fission tracks in zircon start to anneal at similar temperatures (Tagami et al., 1990). Furthermore, the annealing temperature for fission tracks is known to increase with decreasing alpha damage (Kasuya and Naeser, 1988).

From these observations, Brandon and Vance (1992) suggested the following guidelines for detecting regions in a study area where partial resetting may have occurred: (1) The sample shows a marked increase in etch time compared to other samples in the study area. The actual change in etch time depends on the procedures used and also on the time since thermal resetting. (2) The sample shows a systematic younging of old grain ages. Zircons with old detrital FT ages will tend to have the greatest amount of alpha damage and thus should show the largest reduction in FT ages. In turn, the youngest fraction of detrital-zircon FT grain ages should be the most resistant to partial resetting, especially if the detrital FT ages are close in age to the heating event.

All of our samples show a large range in grain ages and normal etch times (4–6 h). As a result, we are confident that the measured FT ages have not changed since deposition and thus provide an accurate record of pre-depositional cooling events in the source region.

**PEAK FITTING AND MINIMUM AGES**

The FTGA distributions for our samples typically show a large span in grain ages, from ca. 10 Ma to older than 100 Ma. The probability for the χ² test (Galbraith, 1981) is effectively zero for 34 of the samples and 9% and 15% for samples 152 and 191, respectively. This result indicates that the variance
must be coupled with a test to find the maximum number of significant components in the distribution, as discussed subsequently.

It is useful to ask why the binomial distribution is important for this problem. The measured data are the counts of spontaneous and induced tracks for each grain, designated as \( r_i \) and \( s_i \), with the index indicating the \( i \)th grain in a sample of \( i = 1 \) to \( N \) dated grains (where \( N \) is total number of dated grains). The variables \( r_i \) and \( s_i \) are Poisson distributed, but they can be converted into a single binomial-distributed variable by the transformation \( r_i/(r_i + s_i) \) (Galbraith and Green, 1990). This variable can be approximated by a Gaussian distribution when values of \( r_i \) and \( s_i \) are large, which is commonly the case for zircons, given their relatively high U content, which implies a high track density. However, as the cooling age and U content decrease, \( r_i \) and \( s_i \) become small. Thus, the Gaussian approximation tends to break down for young cooling ages, especially for apatites, which commonly have low U contents. In these cases, decomposition methods based on the Gaussian distribution (e.g., GAUSSFIT by Brandon, 1992; MIX by Sambridge and Compston, 1994) will perform poorly. In contrast, the binomial peak-fitting method will provide unbiased estimates of peak ages for track counts of any size.

Binomial peak fitting is based on the maximum-likelihood method, which means that the best-fit solution is determined directly by comparing the distribution of the grain data to a predicted mixed binomial distribution. This approach is a significant improvement over the more common least-squares method, which requires more restrictive assumptions about the distribution of the residuals.

A problem with all of the peak-fitting methods is that they require an initial guess of the number and ages of the peaks in the distribution. The number of peaks could be as large as the number of grains in the FTGA distribution. So we are left to ask, how many peaks are significant? It is also important to start with an initial guess that is not too far away from the best-fit solution; otherwise the calculation may fail to find the best-fit solution.

We have used a Windows program BINOMFIT, written by M. Brandon, to find best-fit components for the FT data presented here. The current version of BINOMFIT does an iterative search of peak ages and number of peaks to find the best-fit set of significant components in the mixed distribution. The program has an automated version of the \( F^2 \) test approach outlined in Brandon (1992). The program considers a large set of trial solutions. The trick is to organize the search and to apply appropriate tests to find the best solution.

The quality of fit for each of the trial solutions is scored by using the \( \chi^2 \) statistic, in a manner similar to the conventional \( \chi^2 \) test. The strategy for the search is to iterate the search to include a successively larger number of peaks and to try a large number of initial guesses during each iteration to ensure that an optimal solution is found at each step. The initial guesses for peak ages are generated by using the probability density plot for the FTGA distribution. The density plot is estimated by using the method in Brandon (1996), and the first and second derivatives of the plot are used to find bumps and humps in the plot. The probability density at each candidate age is used to guess the potential size of the peak. We have tried an alternative approach, using evenly spaced ages, but the computation takes longer and does not provide any obvious advantages.

The next step is to iterate through an increasing number of peaks. The first iteration produces a single component age, with an age identical to the pooled age and a \( \chi^2 \) value identical to that produced by the conventional \( \chi^2 \) test. The next iteration finds a best-fit two-
component solution by using all combinations of initial peaks as initial guesses. The best solution is the one with the lowest \( \chi^2 \) value. The program then considers a three-component solution, then a four-component solution, and so on.

With each iteration, the \( F \)-test is used to determine whether the introduction of an additional component has produced a significant improvement in the \( \chi^2 \) statistic (Brandon, 1992). In general, one will find that the \( \chi^2 \) statistic gets smaller with the introduction of a new component. In part, this effect is due to the fact that the additional component provides the model with greater flexibility to fit the data. We need to know whether the improvement in fit is large compared with the expected random variability associated with measurements. In other words, we want to be assured that the additional peak is fitting signal and not noise.

This question is nicely addressed by the \( F \) test. To explain, consider two solutions for \( m \) and \( m + 1 \) peaks and the quality of those fits as indicated by \( \chi^2_m \) and \( \chi^2_{m+1} \). The \( F \) statistic is given by \( F = (m + 1)(\chi^2_m - \chi^2_{m+1})/\chi^2_{m+1} \). When \( F \) is large, then the improvement in fit associated with the additional peak is considered significant. The \( F \) distribution is used to assign a probability \( P(F) \), which is the probability that random variation alone could produce the observed \( F \) statistic. We consider \( P(F) < \sim 5\% \) to indicate that the improvement in fit is significant. Thus, we can find the optimal number of significant peaks by adding peaks until we get a value of \( P(F) > \sim 5\% \).

All of the minimum ages and older peak ages reported here have been estimated by using this method (Tables 1–3). Uncertainties are cited at the 95% confidence level. Note that the uncertainties are asymmetric: the older interval is larger than the younger interval. Those interested in testing our analysis can download the BINOMFIT program and all of our FT grain data at http://www.geology.yale.edu/brandon.

RESULTS

Well-stratified Turbidite Sequences

For most samples in this group (Table 1), BINOMFIT found just two significant peaks. Probability density plots and histograms for selected samples from the coherent sequences are shown on Figure 5. Particularly interesting are analyses 143 and 150 (Table 1, Fig. 5), tuff samples that preserve abundant delicate glass shards and display clear evidence of contamination by older zircons. FTGA distributions for these samples are so similar to those from sandstones in the Coastal OSC that we conclude the provenance for these sediments must have included active volcanic sources.

The distribution of FT minimum ages for samples from the coherent well-stratified turbidite sequences (Table 1) are tightly clustered between 26 and 11 Ma (Fig. 6), essentially identical to the early and middle (?) Miocene age assignments of fossils from the Coastal OSC (Rau, 1975, 1979). In fact, at \( \pm 2\ SE \), only one sample has a minimum age significantly different from early Miocene, an excursion that could have happened by chance alone in this group of 25 dates. Combined distributions for grouped samples from the co-
Samples, Coherent Sandstones in Coastal OSC

![Graph showing minimum ages and older peak ages for coherent sandstone sequences in the Coastal OSC](image)

Figure 6. Plot of minimum ages and older peak ages for coherent sandstone sequences in the Coastal OSC (Table 1). Samples 168 to 195 are from individual localities. Black triangles and gray circles—minimum ages and older peak ages, respectively. Samples in groups A to E are from localities where continuous sequences were repeatedly dated. Open triangles and circles—minimum ages and older peak ages, respectively, for single samples in each group. Black triangles and gray circles—minimum ages and older peak ages, respectively, for the combined grain-age distributions, each of which contains all of the grain ages dated for samples in that group. Error bars show the 95% confidence intervals. Abbreviations for stratigraphic units are after Remane (2000): LM—late Miocene, MM—middle Miocene, EM—early Miocene, LO—late Oligocene, EO—early Oligocene, LE—late Eocene, ME—middle Eocene, EE—early Eocene, LP—late Paleocene, EP—early Paleocene, LK—late Cretaceous.

 coherent sequences (Table 1, Fig. 6) demonstrate that peak ages are reproducible from sample to sample. The peak ages for the combined distributions are generally similar to those for the individual sample distributions.

Mélange Blocks in the Coastal OSC

Minimum ages from mélangé blocks in the Coastal OSC range from 39 to 15 Ma (Fig. 7, Table 2). Although most of these minimum ages overlap with early Miocene dates from the coherent sequences, four dates are clearly Eocene or Oligocene. Because the total number of dated grains in these samples is small, with \( N \) ranging from 17 to 37 (Table 2), additional dating might resolve younger components. However, mélangé rocks in the Coastal OSC indeed contain Eocene fossil localities (Rau, 1975, 1979), and basalt blocks in mélangé were probably derived from the Eocene Crescent Formation of the adjacent Coast Range terrane (Applegate and Brandon, 1989). These data indicate we should not be surprised to find pre-Miocene sandstone blocks in mélangé in the Coastal OSC.

Aalto et al. (1998) reported Cretaceous to Paleocene \(^{40}\)Ar/\(^{39}\)Ar ages for detrital muscovite from sandstone blocks in mélangé in the Coastal OSC. Zircon FT ages from the same location (samples 140 and 175, Table 2) are as young as early Miocene. The discrepancy between the \(^{40}\)Ar/\(^{39}\)Ar ages and zircon FT dates probably results from the different thermal histories for the detrital muscovite derived from deeply exhumed granitic and metamorphic rocks and for the young zircons derived from contemporaneous volcanic sources. In addition, the effective closure temperature for \(^{40}\)Ar/\(^{39}\)Ar muscovite is about 150° C higher than that for zircon FT.

Detection of Small Peaks

The combined distributions from the coherent sequences illustrate possible problems as-
ing a small number of grains. For example, we would find at least one “14 Ma” zircon if the 14 Ma component was present in this sample.

This approach allows us to evaluate with confidence the problems associated with dating a small number of grains. For example, the minimum age of sample 42a in Group A (Table 1) is 14 Ma, but sample 42b does not have a comparable peak. The combined distribution indicates the 14 Ma component makes up only ~12 percent of the total distribution, suggesting the absence of a 14 Ma peak in sample 42b might be due to random variation associated with sampling. More specifically, did we date enough grains in sample 42b to conclude with confidence that the 14 Ma component is absent?

Reference to the binomial distribution (Fig. 8) indicates if the 14 Ma component makes up only ~12 percent of the parent population, the total number of dated grains (Nₛ) must be ≥25 to ensure, at the 95% probability level, that we would find at least one grain age from this component. To ensure finding 5 or more grains for this component requires that Nₛ ≥80. For sample 42b, Nₛ = 27, so we conclude that we should have found at least one “14 Ma” zircon if the 14 Ma component was present in the sample. However, we must use caution in this interpretation; because of the probabilistic nature of the sample, our prediction will fail occasionally (i.e., 1 time out of 20, given the 95 percent probability level used for our testing). Group E contains a similar example, where the BINOMFIT solution for sample 184 does not resolve a 21 Ma peak, which is present in the other 4 samples in this group. Because Nₛ is small for sample 184 (14 grain ages total), the 21 Ma peak may have been missed.

Peak-fitting results for individual samples in Group D indicate a single well-defined young peak at ~19 Ma, but the combined distribution shows two poorly defined peaks (15 and 21 Ma). This result suggests that the greater number of grain ages in the combined distribution provided the ability to resolve two peaks, whereas the smaller individual samples can only resolve one peak in the same age interval. However, the uncertainties for the ages of the two young peaks are larger because there are fewer grains defining each of the peaks. In fact, based on the uncertainties, there is no significant difference between the minimum ages for the individual samples and that for the combined result. This result shows how the estimated uncertainties can be used to guide judgments about peak resolution.

MINIMUM AGES: VALID PROXIES FOR DEPOSITIONAL AGE?

Key to our study is the question: Is the FT minimum age for a detrital-zircon FTGA distribution from an Olympic Structural Complex sandstone a useful proxy for the depositional age of the sandstone? We can test this interpretation by comparing FT minimum ages with fossil ages from Rau (1975, 1979). Of the samples from the coherent sequences (Table 1), 16 are in demonstrable stratigraphic continuity with localities that have age-diagnostic fossils; these samples were selected with caution, taking into account the complex stratigraphy and structure of the Coastal OSC.

For this test, we focus on the lag time (Fig. 9), defined as the FT minimum age minus the fossil age. The null hypothesis is that the lag time is equal to zero, and we seek examples where this hypothesis fails. A Monte Carlo numerical routine was used to determine the uncertainty for the lag-time estimate, given the specified uncertainties for the FT minimum age and the fossil age. The estimated distribution for the FT minimum age is calculated by converting the age and its uncertainty (Table 1) to a new variable z, which is known to be Gaussian distributed (Galbraith, 1990; Brandon, 1996). Gaussian deviates of z were generated using the GASDEV program (Press et al., 1986, p. 203) and then converted back to age for each sample.
to an age distribution. The result is a simulated distribution of replicate measurements of the FT minimum age for the population of zircons represented by our sample.

The estimated distribution of depositional age was generated by random selection from a uniform distribution defined by the age range indicated by the fossil assemblage (Rau, 1975, 1979; Remane, 2000). The replicated minimum ages and fossil ages were used to create a distribution of lag time made up of 10,000 replicate values. The distribution was sorted and used to find the median and 95% confidence interval for the estimated lag time (50%, 2.5%, and 97.5%, respectively), which are shown in Figure 9.

The overall conclusion is that zircon FT minimum ages seem to do a good job as a proxy for the depositional ages of sandstones in the Coastal OSC. Three samples (108, 42a, 168) are unusual in that they have large negative lag times, which means that the FT minimum age is significantly younger than the depositional age. We have no way to judge the cause of these anomalies. They might be due to incorrect assignment of fossil ages to these samples, to problems with the FT dating, or to random variations in the results.

SOURCE FOR SANDSTONES OF COASTAL OSC

A number of authors have suggested the possibility of significant coast-parallel transport within the Cascadia margin (e.g., Engebretson et al., 1985; Wells and Heller, 1988; England and Wells, 1991). A corollary to this idea is that the sediments that formed the Olympic subduction complex may have been deposited much farther south, perhaps offshore of western Oregon or northern California, and then transported northward to the Olympic Mountains (e.g., Palmer and Lingley, 1989; Davis and Hyndman, 1989; Aalto et al., 1995). Much of the motivation for this idea comes from plate reconstructions that predicted oblique convergence across the southern end of the Cascadia subduction zone (Engebretson et al., 1985; DeMets et al., 1990). These ideas have recently been thrown into question by new geodetic data from McCaffrey et al. (2000) that show that the Cascadia arc and forearc are moving as a separate plate, independent from the North American plate. Their results indicate that convergence at the modern subduction zone is nearly orthogonal along the entire length of the Cascadia margin. An important issue is whether there is any independent evidence for coast-parallel transport of the Coastal OSC. We use provenance information here to evaluate this possibility.

Aalto et al. (1995) proposed that the upper Oligocene–lower Miocene Weaverville Formation in the Klamath Mountains might be a nonmarine remnant of the sedimentary system that fed the Coastal OSC. Aalto et al. (1998) evaluated this interpretation by using 40Ar/39Ar ages for detrital muscovites. The Weaverville Formation was shown to have been derived from local bedrock sources in the Klamath Mountains, whereas sandstone in some of the mélangé blocks of the Coastal OSC was attributed to sources in the Idaho batholith. Heller et al. (1992) argued for a similar Idaho batholith source for detrital muscovite that occurs in older units of the Olympic subduction complex. A link to the Idaho batholith might be taken as evidence for a depositional site for the Coastal OSC farther to the south. However, Brandon and Vance (1992) showed that the distinctive muscovite-bearing source identified by Heller et al. (1992) and Aalto et al. (1998) was part of a long belt of schists and two-mica granites, extending from the Idaho batholith northward into southern Canada as part of the Omineca Crystalline belt. As a result, this provenance link is not very useful for judging coast-parallel transport.

So we turn our attention here to sedimentary units within the Cascadia forearc region. In particular, we consider whether any of the more inboard sedimentary units that underlie the Cascadia forearc might be upstream equivalents of the sedimentary rocks currently found in the Coastal OSC. Such equivalents have never been clearly identified. Candidate units of the right age (early Miocene) include the Astoria Formation of western Oregon and southwestern Washington and the Blakley Harbor and Clallam Formations, exposed along the eastern and northern sides of the Olympic Peninsula (Armentrout et al., 1988; Brandon et al., 1998). The Clallam Formation is equivalent to the Sooke Formation, exposed on the southwest side of Vancouver Island. Collectively, these two units outline the Toﬁno-Makah basin (Fig. 2), which underlies the western side of the Straits of Juan de Fuca and the continental shelf west of Vancouver Island (Snively et al., 1980; Garver and Brandon, 1994).

So, what do we look for to assess a connection with these more inboard units? Sandstone compositions in the Coastal OSC are quite variable, but most rocks are lithic arkoses or lithic wackes, with the lithic fragments

Figure 8. Graph showing probabilities that a sample grain-age distribution will contain at least one grain (gray contours) or at least five grains (black contours) from a component of that distribution. The probabilities are a function of the true size (i.e. expected size) of the component and the total number of grains dated. The calculated probabilities are based on the binomial distribution. See text for further discussion.
composed mainly of volcanic material (Fig. 10). Our zircon FT results also indicate that the Coastal OSC received much sediment from an active volcanic source, probably the Cascade volcanic arc. In contrast, the Astoria and Blakley Harbor Formation are dominated by arkosic sediment, with little volcanic debris (Rau, 1967; Neim et al., 1992). Most of the Clallam Formation is also dominated by arkosic sediment (Anderson, 1985) and lacks young zircon FT ages (Garver and Brandon, 1994). However, the stratigraphically youngest part of the Clallam is volcanoclastic rather than arkosic (facies 5 of Anderson, 1985).

We propose that this upper part of the Clallam Formation (Fig. 2) is the upstream equivalent to the Coastal OSC. This unit is the type locality of the Pillarian molluscan stage of the Pacific Northwest biostratigraphic standard, which has an age of 24–20 Ma (Armentrout, 1987), virtually identical to the depositional age of the Coastal OSC. Petrographic data from Stewart (1970) and Anderson (1985) indicate similar sedimentary compositions for these two units (Fig. 10). Sandstones in the upper Clallam contain abundant andesitic and dacitic rock fragments—undoubtedly derived from the Cascade arc—as well as minor chert detritus and rare metasedimentary and meta-basalt detritus. These constituents are typical of the Coastal OSC sandstones. Neither the upper Clallam nor the Coastal OSC contain high-grade metamorphic detritus.

The Clallam Formation and associated Tofino-Makah basin are tied stratigraphically to Vancouver Island and can be considered a stable reference within the interior of the Cascadia forearc. Thus, our proposed correlation suggests that the Coastal OSC formed in place relative to this part of the Cascadia forearc. In fact, it is useful to note that during the early Miocene, basinal sequences within the Cascadia forearc were overwhelmed by arkosic sediment, mainly derived from east of the arc (Neim et al., 1992). We were unable to find any basinal sequence, other than the upper Clallam, that has the volcanoclastic sandstone composition characteristic of the Coastal OSC.

**DEFORMATION AND ACCRETION OF THE COASTAL OLYMPIC STRUCTURAL COMPLEX**

Most of the Coastal OSC consists of steeply east-dipping strata, with sedimentary structures indicating younging to the east (Rau, 1975, 1979; Tabor and Cady, 1978a). The strata are undoubtedly repeated by thrust faults; otherwise the Coastal OSC would have a...
Figure 11. Schematic cross sections of the Cascadia wedge. (A) Steep imbricate structure as proposed by Rau (1975, 1979) and Tabor and Cady (1978a, 1978b), and (B) domal imbricate structure as proposed by Brandon and Calderwood (1990) and Brandon and Vance (1992). Abbreviations refer to tectonic units exposed in the Olympic Mountains: U—Upper OSC, L—Lower OSC, and C—Coastal OSC.

stratigraphic thickness of >20 km. On the basis of these kinds of arguments, Rau (1973, 1980) and Tabor and Cady (1978a, 1978b) proposed a steeply imbricated structure for the entire Olympics, as schematically shown in Figure 11A. This interpretation predicts that depositional ages should get systematically younger from east to west. This relationship does hold at the scale of the entire Olympic subduction complex. Accreted rocks are latest Eocene in age in the eastern Olympics (Upper OSC), early Miocene in the western Olympics (Lower and Coastal OSC), and Quaternary at the modern toe of the wedge.

Our dating here was motivated, in part, to see whether this pattern of younging was present at the local scale in the western side of the Olympic Peninsula. Brandon and Vance (1992) proposed that the Lower OSC was a more deeply exhumed equivalent of the Coastal OSC. This interpretation was based on three unreset zircon FTGA samples from the Lower OSC with minimum ages ranging from 26 to 18 Ma. Two of these FT minimum ages are shown here in Table 3 and Figures 3, 4, and 7. Our new results for the Coastal OSC demonstrate that a 50-km-wide area, extending from the Pacific Coast to Mount Olympus, is underlain by accreted lower Miocene sedimentary rocks (Fig. 4). There is subtle evidence of younging across this region from ca. 19 Ma at Mount Olympus to ca. 14 Ma at the coast (Fig. 4).

We can make a rough prediction for the expected amount of younging in accreted sedimentary rocks if the structure were steeply dipping as suggested in Figure 11A. It should be approximately equal to the distance across the region divided by the horizontal material velocity relative to the wedge front. Pazzaglia and Brandon (2001) estimated that at the coast, this velocity has been steady at 3 km/ m.y. Thus, a steep imbricate structure should show ~17 m.y. of younging across the 50 km distance from the Pacific Coast to Mount Olympus.

An alternative interpretation (Fig. 11B) is that the Olympics have a more domal structure (Brandon and Vance, 1992; Brandon et al., 1998). The structural lid is thought to have originally extended to the west coast and perhaps farther offshore, as is the case for southern Vancouver Island and also for western Oregon and southwestern Washington (Brandon and Calderwood, 1990). Early emergence of the Olympics at ca. 15 Ma, relative to the rest of the Cascadia forearc high, allowed for uplift and deep erosion of the Olympics, removing the lid and exposing the underlying subduction complex (Brandon and Calderwood, 1990; Brandon et al., 1998).

The more domal structure accounts for the broad expanse of accreted lower Miocene sedimentary rocks. Figure 12 shows schematically how material might have moved through the wedge. This interpretation builds on the
conclusion that in the Olympics sector of the Cascadia margin, the wedge probably reached its present size early in its evolution, at ca. 15 Ma, and that accretion occurred primarily at the front of the wedge (Brandon et al., 1998; Pazzaglia and Brandon, 2001; Batt et al., 2001). Thus, we envision that the Coastal OSC was first accreted beneath the front 50 to 100 km of the wedge, but then moved rearward within the wedge because of further accretion at the front of the wedge and erosion at the back of the wedge. As previously noted, the most western exposures of the Coastal OSC have never been deeply buried and exhumed. Thus, those rocks are shown in Figure 12 as moving along a path near the surface of the wedge. Conversely, the more eastern lower Miocene rocks of the Lower OSC followed a deeper path through the wedge (Fig. 11), reaching maximum depths of as much as 13 km, but ultimately rising to the surface near Mount Olympus (Batt et al., 2001).

We can test this idea by comparing the age of the Coastal OSC to the amount of time needed to reach the coast within a wedge that maintained a steady taper. This calculation is relatively easy because the frontal part of the wedge is not eroding and has probably maintained the same critical taper throughout much of its evolution. In this case, the average horizontal velocity \( u(x) \) at a distance \( x \) from the front of the wedge is related to the accretionary flux \( F_a \) by

\[
\dot{u}(x) = \frac{F_a}{h(x)}
\]  

(1)

where \( h(x) \) is the thickness of the wedge at \( x \) (see Dahlen and Barr [1989] and Pazzaglia and Brandon [2001] for details about this calculation). The only assumptions are that the wedge seaward of the coast has maintained a constant taper and that the accretionary flux has been steady over the time frame of interest. In other words, the full wedge does not need to be at steady state for equation 1 to hold. The accretionary flux is specified in terms of solid rock mass, which is consistent with the fact that most of the porosity is lost within the frontal 10 km of the wedge (Davis and Hyndman, 1989).

We have to integrate the velocity along the particle path to get \( \tau(x) \), the transit time from the trench to a position \( x \) in the wedge. The integration gives

\[
\tau(x) = \frac{A(x)}{F_a}
\]  

(2)

where \( A(x) \) is the cross-sectional area of the wedge between \( x \) and the front of the wedge.

Thermal-kinematic modeling by Batt et al. (2001) indicates a long-term average accretionary flux of 58 km²/m.y. into the Olympics sector of the Cascadia wedge. The profile in Figure 11 was used to calculate \( A(x) \). Equation 2 was then used to estimate the transport time through the wedge (Fig. 13). The prediction is that material accreted at the deformation front of the trench would reach the West Coast in 22 m.y., which is in excellent agreement with our estimate of 24 to 16 Ma for deposition of the Coastal OSC in the trench.

CONCLUSIONS

FT grain ages from detrital zircons in sandstones of the Coastal OSC indicate a mixed source of detrital zircons, derived from the Cascade volcanic arc and from older basement rocks lying to the east of the arc. The youngest components make up, on average, \(~40\%\) to \(50\%\) of a typical zircon FTGA distribution. Comparison with fossil ages indicates that the FT minimum age, which is the pooled age of the youngest component, is a good proxy for the depositional age of sandstones from the Coastal OSC. This result is consistent with other evidence indicating that the volcanic sediment was derived from contemporaneous active volcanoes in the arc. The zircon FT minimum ages indicate that the Coastal OSC is mainly early Miocene in age, which is in close agreement with fossil ages.

Basinal units in the Cascadia forearc are mainly arkosic, which is in contrast to the more volcaniclastic composition of the Coastal OSC. We have found only one coeval unit, within the upper part of the Clallam Formation on the north side of the Olympic Peninsula, that has a similar sedimentary composition. This provenance correlation argues against large-scale coast-parallel transport of the Coastal OSC after accretion. Instead, transport appears to have mainly occurred to the northeast, in association with marginal-normal shortening of the Cascadia wedge, in the direction of plate convergence.

We present a simple model that predicts the average transport time for frontally accreted materials moving through a wedge with a steady taper. The model predicts that frontally accreted sedimentary rocks should take \(~22\) m.y. to reach the present position of the Coastal OSC, located \(~140\) km landward of the front of the wedge. This estimate assumes a steady taper like the modern taper of the wedge and a steady accretionary flux like the modern. The predicted transport time is in close agreement with the early Miocene age of the Coastal OSC and thus is consistent with other evidence for a flux steady state in the Olympics sector of the Cascadia wedge (Brandon et al., 1998; Pazzaglia and Brandon, 2001; Batt et al., 2001).

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Figure 13. Transport time needed to move from the site of initial accretion at the front of the wedge to a location rearward in the wedge, according to equation 2 in the text.
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