Linking deep and shallow crustal processes during regional transtension in an exhumed continental arc, North Cascades, northwestern Cordillera (USA)

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ABSTRACT

The North Cascades orogen (northwestern USA) provides an exceptional natural laboratory with which to evaluate potential temporal and kinematic links between processes operating at a wide range of crustal levels during collapse of a continental arc, and particularly the compatibility of strain between the upper and lower crust. This magmatic arc reached a crustal thickness of ≥55 km in the mid-Cretaceous. Eocene collapse of the arc during regional transtension was marked by magmatism, migmatization, ductile flow, and exhumation of deep crust (8–12 kbar) rocks in the Cascades crystalline core coeval with subsidence and rapid deposition in nonmarine basins adjacent to the core, and intrusion of dike complexes. The Skagit Gneiss Complex is the larger of two regions of exhumed deep crust with Eocene cooling ages in the Cascades core, and it consists primarily of tonalitic orthogneiss emplaced mainly in two episodes of ca. 73–59 Ma and 50–45 Ma. Metamorphism, melt crystallization, and ductile deformation of migmatitic metapelite overlap the orthogneiss emplacement, occurring (possibly intermittently) from ca. 71 to 53 Ma; the youngest orthogneisses overlap ⁴₀Ar/³⁹Ar biotite dates, compatible with rapid cooling. Gently to moderately dipping foliation, horizontal orogen-parallel (northwest-southeast) mineral lineation, sizable constrictional domains, and strong stretching parallel to lineation of hinges of mesoscopic folds in the Skagit Gneiss Complex are compatible with transtension linked to dextral-normal displacement of the Ross Lake fault zone, the northeastern boundary of the Cascades core. The other deeply exhumed domain, the 9–12 kbar Swakane Biotite Gneiss, has a broadly north-trending, gently plunging lineation and gently to moderately dipping foliation, which are associated with top-to-the-north noncoaxial shear. This gneiss is separated from overlying metamorphic rocks by a folded detachment fault. The Eocene Swauk and Chumstick basins flank the southern end of the Cascades core in the Swauk basin, sediments were deposited in part at ca. 51 Ma, folded shortly afterward, and then covered by ca. 49 Ma Teanaway basalts and intruded by associated mafic dikes. Directly after dike intrusion, the fault-bounded Chumstick basin subsided rapidly. Extension directions from these dikes and from Eocene dikes that intruded the Cascades core are dominantly oblique to the overall trend of the orogen (275°−310° vs. ~320°, respectively) and to the northwest-south–east to north-south ductile flow direction in the Skagit and Swakane rocks. This discordance implies that coeval extensional strain was decoupled between the brittle and ductile crust. Strain orientations at all depths in the Cascades core contrast with the approximately east-west extension driven by orogenic collapse in coeval metamorphic core complexes ~200 km to the east. Arc-oblique to arc-parallel flow in the Cascades core probably resulted in part from dextral shear along the plate margin and from along-strike gradients in crustal thickness and temperature.

INTRODUCTION

Collisional orogens and some contractional continental magmatic arcs have gross similarities in their structure, including thickened crust (50–70 km) and broad plateaus that have zones of partially molten lower to middle crust (e.g., Tibet and Altiplano-Puna; Nelson et al., 1996; Schilling et al., 2006; Ward et al., 2014). Studies of exhumed orogenic crust and models for lateral and vertical crustal flow have most commonly been formulated based on collisional orogens, such as the Himalayas (e.g., Law et al., 2006, and references therein). Much less is known about the deep levels of continental arcs, which contain the rocks that record the thermal, rheological, and mechanical transition of an orogen from contractional crustal thickening to extension and/or transtension. For example, in the North American Cordillera, more research has concentrated on deep crustal flow and exhumation of the hinterland metamorphic core complexes (e.g., MacCready et al., 1997; Vanderhaeghe and Teyssier, 1997; Teyssier et al., 2005; Gervais and Brown, 2011) than on the Mesozoic arc system to the west (Fig. 1). Similarly, lower to middle crustal flow
is poorly documented in other exhumed continental magmatic arcs, with a few exceptions (e.g., Fiordland, New Zealand; Klepeis et al., 2004; Klepeis and King, 2009; Betka and Klepeis, 2013). In this study we focus on the 96–45 Ma North Cascades magmatic arc of the State of Washington (USA) and southwest British Columbia (Canada), which is the southern continuation of the Coast Mountains batholith. This arc likely reached thicknesses of ≥55 km (Miller and Paterson, 2001) in the Late Cretaceous, and a wide range of Cretaceous crustal levels (0 to ≥40 km) is exposed. It is interpreted as the western margin of a proposed Late Cretaceous to Eocene orogenic plateau, the eastern margin of which is marked by metamorphic core complexes (Whitney et al., 2004). Eocene migmatization, plutonism, and exhumation of some of the deepest (8–12 kbar) rocks of the Cascades crystalline core were in part coeval with motion on dextral strike-slip and oblique-slip faults, formation of nonmarine, transtensional basins, and intrusion of extensive basaltic to rhyolitic dike swarms (e.g., Haugerud et al., 1991; Gordon et al., 2010a; Eddy et al., 2016). This orogen thus offers an exceptional opportunity to evaluate links between contemporaneous deep and shallow crustal processes during regional transtension. Furthermore, the record of how different crustal levels evolved provides a view of the construction and collapse of a thick continental arc that may be analogous in some respects to the Cenozoic central Andes (Western Cordillera and Altiplano) (e.g., Scheuber and Reutter, 1992; Beck and Zandt, 2002), where deep crust has not been exhumed.

We integrate a large body of different types of data to arrive at a synthesis of processes occurring from the middle crust to the surface during a time interval largely centered on the Eocene (60–45 Ma). We address the potential interplay of deformation, metamorphism, partial melting, and magmatism at different depths, including the temporal and dynamic relations of these processes to exhumation of arc orogenic crust, formation of nonmarine basins, and intrusion of dike swarms. Particular emphasis is placed on how strain was partitioned at different depths in the crust during transtension and whether the upper crust was decoupled from the deeper, ductilely flowing lower crust (e.g., Tibetan Plateau, Royden et al., 1997; Altiplano-Puna, Husson and Semere, 2003; Gerbault et al., 2005; Ouimet and Cook, 2010).

Overall, the North Cascades arc developed during a time interval largely centered on the Eocene (60–45 Ma). We address the potential interplay of deformation, metamorphism, partial melting, and magmatism at different depths, including the temporal and dynamic relations of these processes to exhumation of arc orogenic crust, formation of nonmarine basins, and intrusion of dike swarms. Particular emphasis is placed on how strain was partitioned at different depths in the crust during transtension and whether the upper crust was decoupled from the deeper, ductilely flowing lower crust. This orogen thus offers an exceptional opportunity to evaluate links between contemporaneous deep and shallow crustal processes during regional transtension. Furthermore, the record of how different crustal levels evolved provides a view of the construction and collapse of a thick continental arc that may be analogous in some respects to the Cenozoic central Andes (Western Cordillera and Altiplano).
Figure 2. Geologic map emphasizing Eocene tectonic elements of the Cascades core and adjacent areas. Dextral strike-slip motion on the Straight Creek fault has been restored. Eocene plutons and units with young cooling ages (Skagit Gneiss Complex and Swakane Biotite Gneiss) in the Cascades core are emphasized. CM—Cooper Mountain batholith; DH—Duncan Hill pluton; GH—Golden Horn batholith; LFZ—Leavenworth fault zone; RC—Railroad Creek pluton; RLFZ—Ross Lake fault zone. Eocene fold traces are shown in Eocene basins and Cascades core. Sources are described in the text.
This unit and approximately coeval orthogneisses represent a Late Triassic arc. Much of the southern part of the Cascades core consists of the dominantly metapelitic and metapsammitic Chiwaukum Schist of the Nason terrane (e.g., Plummer, 1980; Tabor et al., 1987b), and includes Early Cretaceous protoliths (Brown and Gehrels, 2007). The structurally deepest terrane is the Late Cretaceous metapsammitic Swakane Biotite Gneiss (Fig. 2) (Matzel et al., 2004; Paterson et al., 2004; Gatewood and Stowell, 2012).

Much of the Cascades core was intruded, metamorphosed, and ductilely shortened in the mid-Cretaceous (ca. 100–85 Ma) synchronous with regional thrusting and folding (Fig. 1) (e.g., Misch, 1966; Brown, 1987; Mcgroder, 1991).
Crustal thickness in the core probably reached ≥55 km at ca. 90 Ma (Miller and Paterson, 2001). After the major shortening event, magmatism (79–45 Ma), metamorphism, and deformation were focused in the northeastern part of the Cascades core (Chelan block; Fig. 1). The Swakane Biotite Gneiss was buried to ~35–40 km depth by 88 Ma (Matzel et al., 2004; Paterson et al., 2004; Gatewood and Stowell, 2012). High-angle Paleogene faults (Fig. 1) underwent variable amounts of dextral strike slip and dip slip, and a transition from components of reverse slip to normal slip on some of these faults likely coincided with a regional change from transpression to transtension (Miller and Bowring, 1990; Haugerud et al., 1991).

The youngest 40Ar/39Ar and K-Ar ages (e.g., Wernicke and Getty, 1997; Paterson et al., 2004) and ductile deformation are found in the two domains of deep crust that underwent substantial exhumation accompanied by Eocene near-isothermal decompression and subsequent rapid cooling (Fig. 2). These domains include the orthogneiss-dominated, partially migmatitic 8–10 kbar Skagit Gneiss Complex, and the largely nonmigmatitic 9–12 kbar Skawane Biotite Gneiss (Misch, 1968; Haugerud et al., 1991; Whitney, 1992b; Valley et al., 2003; Gordon et al., 2010a). Ductile deformation, late magmatism, partial melting of Skagit rocks, and exhumation of these domains overlapped with regional transtension, indicated in the shallow crust by coeval formation and rapid subsidence of Eocene nonmarine basins bounded by oblique-slip (dextral and normal) faults (e.g., Johnson, 1985; Evans, 1994; Eddy et al., 2016), and intrusion of abundant Eocene dikes. These basins accumulated clastic rocks that in the central Washington Cascades may have reached ≥12 km in thickness (Tabor et al., 1984; Evans, 1994).

Several studies have argued that ridge subduction occurred at ca. 50 Ma at the latitude of northern Washington (Thorkelson and Taylor, 1989; Cowan, 2003; Haussler et al., 2003; Madsen et al., 2006) and created a slab window beneath the Cascades at that time. In the Cascades core, magmatism changed from dominantly tonalite at 96–60 Ma to mainly granodiorite at 50–45 Ma (Misch, 1966; Miller et al., 2009a). Most of the widespread dikes are inferred to be ca. 49–45 Ma (Tabor et al., 1984; Eddy et al., 2016). Dikes occur in granitic swarms, some of which are probably related to shallow plutons in the Cascades core; the most voluminous are in a mafic swarm that intruded the Eocene Swauk basin, and others are isolated bodies that intruded parts of the core. All magmatism, metamorphism, and ductile deformation ended by ca. 43 Ma, when the crystalline core rocks were exhumed to the near surface, as recorded by 40Ar/39Ar and K-Ar hornblende and biotite dates (Engels et al., 1976; Wernicke and Getty, 1997; Tabor et al., 2004; Matzel, 2004). Subsequent subduction of the Farallon plate resulted in the modern Cascades arc (e.g., Tepper, 1996) starting ca. 40 Ma (du Bray and John, 2011).

In the following we integrate new data from our multifaceted research with those of previous workers to provide a view of the Eocene crustal architecture and processes operating throughout the crustal section. These data include material in theses by some of the authors (Michels, 2008; Shea, 2008; Doran, 2009; McLean, 2012), publications arising from group research (Miller and Bowring, 1990; Gordon et al., 2010a, 2010b; Wintzer, 2012), and other publications, notably Misch (1968), Haugerud et al. (1991), Paterson et al. (2004), and Eddy et al. (2016). The main new data sets presented herein include high-precision U-Pb (chemical abrasion–isotope dilution–therm ionization mass spectrometry, CA- ID-TIMS) zircon dating of 10 orthogneiss bodies in the Skagit Gneiss Complex (Fig. 3; Table 1; see Supplemental Table 1 and http://geochron.org/dataset/html/geochron_dataset_2016_01_22_kCnxS for details of dates); mapping and structural analysis of the central and southern parts of the gneiss complex and the upper contact of the complex; structural fabrics in Eocene plutons; and orientations and extension directions of Eocene dikes in the shallow crust. The data description begins with the deep crust, focusing on the PT-t-d (pressure-temperature-time-deformation) history of the Skagit Gneiss Complex and to a lesser extent the Swakane Biotite Gneiss, followed by analysis of fabrics and regional strain in Eocene plutons, and then on high-angle faults, nonmarine basins, and dike swarms in the shallow crust.

### TABLE 1. SUMMARY OF U-Pb ZIRCON CRYSTALLIZATION AGES

| Abbreviation | Sample | Age (Ma) | Unit |
|--------------|--------|---------|------|
| DL           | NC-197 | 44.857 ± 0.023 | Diablo Lake orthogneiss (see text) |
| FL           | H2OFG  | 47.197 ± 0.020 | Flecked gneiss (Miller et al., 2009b, p. 402) |
| FP           | ES-32  | 72.787 ± 0.046* | Ferry Peak orthogneiss (Shea, 2008) |
| GG           | SGC-54 | 49.377 ± 0.025 | Tonalitic orthogneiss (Shea, 2008) |
| PC           | NC-581 | 48.490 ± 0.027 | Purple Creek orthogneiss (Michels, 2008) |
| PO           | SGC-02 | 48.158 ± 0.032 | Promylyotic orthogneiss (Miller et al., 2009b, p. 398) |
| RF           | SGC-95 | 49.198 ± 0.043 | Rainbow Falls orthogneiss (Michels, 2008) |
| SL           | NC-92  | 49.484 ± 0.026 | Sunrise Lake sheet; cuts foliation in tonalite |
| SLs          | NC-95  | 49.198 ± 0.043 | Stehekin orthogneiss (Michels, 2008) |

Note: Abbreviations as in text Figure 3. Sample identifications and details as in Supplemental Table 1 (see text footnote 1); see http://geochron.org/dataset/html/geochron_dataset_2016_01_22_kCnxS.

*Youngest grain; significant inheritance.
PALEOGENE MIDDLE TO DEEP CRUST

Skagit Gneiss Complex

Map Pattern and Rock Units

The Skagit Gneiss Complex extends for >100 km along strike, and is as much as 30 km wide (Misch, 1966, 1968; Haugerud et al., 1991) (Figs. 2 and 3). It is bounded on the northeast by the Ross Lake fault zone and is intruded on the southwest by Eocene plutons or is in contact with the Napeequa unit, Late Triassic arc rocks, or Cretaceous orthogneisses (Fig. 3). An intrusive contact with the ca. 48 Ma Cooper Mountain batholith marks the southern boundary and the complex is truncated by the intersection of the Ross Lake and Straight Creek–Fraser faults to the north (Figs. 2 and 3). The Skagit Gneiss Complex consists of orthogneiss, “banded gneiss,” which is strongly foliated and commonly migmatitic gneiss of uncertain protoliths, and metasupracrustal rocks (Misch, 1968). Orthogneiss makes up >75% (Haugerud et al., 1991) of the complex, and on the basis of our mapping, probably >90% of its central and southern parts. Much of the northern half is migmatitic (Misch, 1968).

The dominant protoliths of Skagit orthogneisses are biotite tonalite and leuco- tonalite; hornblende-bearing tonalite and biotite granodiorite are moderately abundant, and granite to quartz monzonite and diorite orthogneiss occur locally (Misch, 1968; Haugerud et al., 1991; Miller et al., 1994; our mapping). Crystallization ages of the orthogneiss protoliths range from ca. 89 to 45 Ma, and the largest volume of orthogneiss probably intruded between ca. 73 and 59 Ma (Fig. 3; U-Pb CA-ID-TIMS zircon data in Table 1; other ID-TIMS zircon dates from Mattinson, 1972; Miller et al., 1989; Miller and Bowring, 1990; Haugerud et al., 1991). Most of these orthogneisses are sheeted, or heterogeneous on the outcrop to 100 m scale, and are intimately intruded by 10-cm- thick to 10-m-thick leucocratic, commonly pegmatitic sheets. An apparent intrusive hiatus occurred between the ca. 59 and 50–45 Ma orthogneisses (Gordon et al., 2010a); these crystallization ages overlap with those of the granite to quartz monzonite and diorite orthogneiss ranging from <1 m to ~750 m across, of which most are considered in terms of three domains (shown in Figs. 3 and 5) that are distinguished by different foliation orientations. Foliation is generally defined by aligned biotite, quartz, and quartz-plagioclase aggregates, and in some rocks by hornblende. Most of the complex is foliated and lineated, but fabric intensity varies considerably. The weakest fabrics are in orthogneiss and particularly in the younger (ca. 50–48 Ma) orthogneiss in the central and southern domains. In addition, constrictional fabrics are common in some areas, and are recognized qualitatively by stronger lineation than foliation, and by strain markers, such as in metamorphosed fragmental rocks and quartz-plagioclase aggregates in orthogneisses (Misch, 1968; our observations). They are most notable in the young (ca. 45 Ma) Diablo Lake orthogneiss (Misch, 1968; Haugerud et al., 1991; Winter, 2012) and parts of the stepover in the Ross Lake fault zone.

In much of the gneiss complex, foliation strikes northwest (Fig. 5A), and on a regional scale, the complex forms a northwest-southeast–trending anticlinorium (Tabor et al., 1989). Mineral lineation typically plunges gently to the northwest or southeast (e.g., Brown and Talbot, 1989), but is dominal in orientation (Fig. 5B). Outcrop-scale folds are common in well-layered gneiss and migmatite in the northern domain. Earliest structures are tight to
isoclinal, gently inclined to recumbent folds that typically have wavelengths of <25 cm (Misch, 1977a; Wintzer, 2012). Locally, the dominant foliation is axial-planar to folded leucosomes and compositional layers, but it is generally folded by the earliest structures. Hinge lines are variably oriented reflecting subsequent refolding, and the average orientation of axial surfaces is approximately parallel to foliation (Wintzer, 2012; our observations). The dominant mesoscopic folds are upright to inclined structures that range from gentle to tight, and have wavelengths of ~1 cm to 20 m (Fig. 4A) (Misch, 1968; Haugerud et al., 1991; Wintzer, 2012). These folds are probably parasitic to the upright map-scale folds (Fig. 5C). Hinge lines mostly plunge southeast, are subparallel to mineral lineations, and are commonly stretched parallel to lineation (Fig. 4B).

Map-scale folds of foliation and unit contacts are mainly upright to steeply inclined structures (Figs. 6A, 6B; e.g., Misch, 1966; Tabor and Haugerud, 1999; Tabor et al., 2003; Miller et al., 2009b; Wintzer, 2012). They most commonly plunge gently southeast or northwest, but in parts of the complex the axial traces are curved. In the south, symmetric gentle to open map-scale folds of foliation have axial traces that trend west-northwest near the Cooper Mountain...
batholith, and curve to a northwest trend to the north (Fig. 5C). Gently plunging mineral lineation similarly swings from east-southeast near the batholith to northwest farther to the north, and to northwest-southeast and northeast in the northern part of the complex (Fig. 5B). Most map-scale folds die out northward from the Cooper Mountain batholith, whereas the longest (8 km) wavelength and westernmost antiform continues for ≥30 km. In the central domain, foliation defines a broadly fan-like antiform; in the west, it strikes northwest and dips southwest, and in the east, it strikes north-northeast and dips east. Lineation shows a similar swing (Adams, 1961; our data). In the northern domain, gently southeast-plunging map-scale folds have wavelengths of a few kilometers (Fig. 6A, A’). The northwest-striking foliation of this corridor changes in the east near the stepover in the Ross Lake fault zone where strikes...
rotate to northeast and east-west (Fig. 5A), and dips are lower compared to the south. Mineral lineation here mostly plunges gently to the south, or to the south-southwest and north-northeast, in contrast to the gentle southeast plunge of lineation away from the stepover.

The latest folds are defined by the large swings in the regional trends of foliation, lineation, and hinge lines (Fig. 5), and probably plunge moderately to steeply. These open folds record approximately orogen-parallel shortening that deformed foliation in rocks as young as the ca. 47.2 Ma sheet (FL, Fig. 3; Table 1) in the northern domain.

Non coaxial flow is recognized in some areas at the outcrop scale to microscale by asymmetric structures, but well-defined ductile shear zones are uncommon at the 10 m scale or larger. These asymmetric structures, which are...
used to determine shear sense, include S-C relations, shear bands (C’ surfaces), oblique quartz foliation, asymmetric plagioclase porphyroclast systems (Fig. 4C), mica fish, asymmetric leucosomes and boudins, and foliation curvature. These features are concentrated in weaker micaceous layers, particularly in well-layered gneisses next to orthogneiss bodies and pegmatitic sheets. Shear zones are generally subparallel to the dominant foliation and compositional layering. Widespread, but less common, generally ≤15-cm-thick shear zones cut foliation at moderate to high angles, are generally steeper than the foliation-parallel ones, and are highly variable in strike. Lineations in the late zones also show considerable scatter, but the majority plunges gently to moderately to the east-southeast and southeast. Similar kinematic indicators occur in both types of shear zones.
In a few outcrops, asymmetric fabrics in foliation-parallel shear zones show a reversal in shear sense from sinistral to dextral across fold hinges. The best-documented cases have top-to-the-northwest shear in both limbs and in the hinge. This reversal and the shear direction (mineral lineation) do not result from flexural shear, as the lineations and hinge lines are sub-parallel (cf. Goscombe and Trouw, 1999). On the basis of these observations, and following Wintzer (2012), we reevaluated kinematic patterns for the foliation-parallel shear zones after unfolding the upright, gently plunging folds (such as in Fig. 4A) and restoring foliations to an approximately subhorizontal orientation.

In the southern domain, after unfolding we found examples of the top-to-the-southeast and top-to-the-northwest shear. In the central part of the gneiss complex, we rarely observed shear-sense indicators outside of the dextral Ross Lake fault zone (see following). In the northern domain, shear-sense indicators are widespread (Wintzer, 2012; our data), including in orthogneiss as young as ca. 48–47 Ma, but are not recognized in the ca. 45 Ma Diablo Lake orthogneiss.
bodies. Both top-to-the-northwest (mostly dextral) and top-to-the-southeast (mostly sinistral) shear are recorded throughout the corridor. Noncoaxial shear initiated at amphibolite facies conditions, as cummingtonite and sillimanite are in C-surfaces and plagioclase is recrystallized in foliation-subparallel zones (Gordon et al., 2010a), and some of the shear zones at moderate angles to foliation in the southern domain appear to have been filled with melt (Fig. 4D). Shear continued during cooling to greenschist facies conditions in both types of shear zones, as in some zones quartz shows basal \( \langle a \rangle \) slip, bulging recrystallization, and considerable unstrained strain, and plagioclase is microfractured and not recrystallized (cf. Gordon et al., 2010a, 2010b; Wintzer, 2012). Overall patterns of noncoaxial flow are clearly complicated in the Skagit Gneiss Complex. We thus further refined our analyses by looking at shear sense in some of the youngest dated orthogneisses, including the 47.192 Ma ± 0.002 Ma, plagioclase-phryic, sphene-coated, orthogneiss sheet and a 48.490 Ma ± 0.027 Ma (PO, Fig. 3; Table 1) protomylonitic orthogneiss body. There appears to be an association of the youngest greenschist facies microstructures and top-to-the-northwest shear, but we also found examples with top-to-the-southeast shear.

We estimated the minimum shortening of the gneiss complex recorded by map-scale upright folds in two cross sections (Figs. 6A, 6B) by assuming flexural slip and using simple line-length balancing. Results are ≥33% in the northern domain and ≥25% in the southern domain. The magnitude of shortening by foliation development and the homogeneous shortening indicated by the thickening of layers in fold hinges is unknown.

**Timing of Deformation and Cooling**

The Skagit Gneiss Complex underwent a protracted history of metamorphism and deformation (Table 2). The overprinting by the extensive ca. 73–45 Ma magmatism, metamorphism, and deformation makes it difficult to correlate any structures within the complex to the mid-Cretaceous (ca. 100–80 Ma) regional shortening that is well documented in the Wenatchee block of the Cascades core and in low-grade rocks adjacent to the core. Similarly, the high-P metamorphism that initiated by 75 Ma in the core (Miller et al., 1993a, 1993b; Brown et al., 1994) has been overprinted.

Orthogneisses that yield inferred crystallization ages as young as 47.2 Ma (FL, Fig. 3; Table 1), ca. 69–51 Ma leucosomes, and metamafic crustal rocks with monazite dates as young as 47 Ma (Gordon et al., 2010a) display the dominant foliation and lineation in the Skagit Gneiss Complex. In two of the dated migmatite localities, a 65–64 Ma and a 53 Ma discordant leucosome cut the foliation of the host metapelite and biotite gneiss, indicating that foliation initiated by 65 Ma in at least part of the gneiss complex. Both discordant leucosomes were later deformed under greenschist-facies conditions (Gordon et al., 2010a).

Fabric development was in part time transgressive. Solid-state foliation and lineation in orthogneisses in the southern part of the complex, and folding of the 49.484 ± 0.026 Ma Sunrise Lake tonalite (SL, Fig. 3; Table 1) predated intrusion of the ca. 48 Ma Cooper Mountain batholith. A 49.198 ± 0.043 Ma felsic sheet (SL1, Fig. 3; Table 1) that cuts foliation in the tonalite may better define the timing of deformation. In comparison, the 48.5–472 Ma orthogneisses (FG, PO, Fig. 3; Table 1) in the northern part of the complex record significant solid-state deformation. Moreover, ca. 45 Ma Diablo Lake orthogneiss bodies crosscut foliation, although these orthogneisses have strong solid-state lineation (Misch, 1988; Haugerud et al., 1991; Wintzer, 2012). All ductile deformation ended before intrusion of the 34 Ma phase of the Chiliwack batholith (Fig. 3).

Most K-Ar and \(^{40}\text{Ar}/^{39}\text{Ar}\) biotite and hornblende dates from the Skagit Gneiss Complex range from ca. 50 to 44 Ma (Engels et al., 1976; Wernicke and Getty, 1997; Tabor et al., 2003; Gordon et al., 2010b). A sample from near (~2.5–5 km) dated migmatites gives \(^{40}\text{Ar}/^{39}\text{Ar}\) hornblende and biotite dates of 47.1 ± 1.3 Ma and 45.2 ± 0.2 Ma, respectively (Wernicke and Getty, 1997), and muscovite from two structurally higher pegmatite samples on Ruby Mountain (Fig. 3) yield muscovite \(^{40}\text{Ar}/^{39}\text{Ar}\) dates of 47.1 ± 0.3 Ma and 46.8 ± 0.3 Ma (Gordon et al., 2010b). Overall, the temporal overlap of the youngest zircon (ca. 47 Ma) and monazite (ca. 49–46 Ma) dates from migmatites and orthogneisses with the \(^{40}\text{Ar}/^{39}\text{Ar}\) and K-Ar dates indicates rapid cooling at 47–45 Ma.

**Boundaries of the Skagit Gneiss Complex**

The relationships of the Skagit Gneiss Complex to adjacent, generally lower grade rocks are important for models of crustal flow and partitioning of deformation at different crustal levels. Lengthy segments of the western and southern boundaries are intrusive contacts with younger weakly deformed plutons. Where the gneiss complex is in contact with older rocks on the west (Fig. 3), foliation and lineation are concordant across the contact (Cater and Wright, 1967; Miller, 1987; Tabor et al., 2003; our mapping). The northeast boundary in the Ross Lake fault zone and the upper contact with the Napeequa unit are more complicated.

The ≥10-km-wide Ross Lake fault system (Misch, 1966; Miller, 1994) is part of a 500-km-long zone of Paleogene, northwest-striking high-angle faults (Fig. 1) (e.g., Monger, 1986). The northernmost strand of the Ross Lake fault is a vertical mylonite zone in British Columbia that separates upper-amphibolite-facies rocks of the Skagit Gneiss Complex from sub-greenschist-facies rocks to the east (McTaggart and Thompson, 1967). Dextral shear and 6–12 km of north-east-side-down normal slip occurred from ca. 50(?) to after 45 Ma (Haugerud, 1985) and predated Oligocene magmatic rocks to the south. In Washington, the fault is intruded by the ca. 48 Ma Ruby Creek plutonic belt, and steps westward 7 km across a gently to moderately dipping stepover zone to the Gabriel Peak tectonic belt (Figs. 3 and 4E) (Miller et al., 1994; Gordon et al., 2010b). This mylonitic, moderately to steeply north-east-dipping belt records dextral shear in the north and reverse shear farther south (Fig. 3). Transpressional deformation in the belt occurred from 65 Ma (and earlier?) to ca. 58 Ma (Miller and Bowring, 1990; Miller et al., 1994). The tectonic belt merges to the south with the North Creek fault to form the Foggy Dew fault zone (Fig. 3), which separates sillimanite-bearing mylonites from greenschist-facies rocks to the east, and underwent oblique dextral-normal slip (down to the east) between ca. 50 and 48 Ma (Miller and Bowring, 1990).
The gently to moderately northeast-dipping upper contact of the Skagit Gneiss Complex is best exposed on Ruby Mountain (Fig. 3), where we analyzed the structure. There, orthogneiss and local pelitic schist are structurally below amphibolite, quartzite, and metaperidotite of the Napeequa unit, and Skagit and Napeequa rocks are both intruded by trondhjemitic pegmatites. This contact is part of a shear zone marked by a strain gradient with pronounced intensification of lineation and the formation of constrictional (L >> S) fabrics. Protomylonitic gneiss near the contact has moderate- to high-temperature microstructures, marked by recrystallization of hornblende, plagioclase, and quartz, and is cut by widespread contact-parallel ductile shear zones. In addition, kinematic indicators give conflicting senses of noncoaxial shear from outcrop to outcrop. In one scenario, the intense fabrics and juxtaposition of different rock types represent a detachment zone; however, the inconsistent sense of shear, similar P-T conditions of 8–10 kbar at 650 °C, and muscovite 40Ar/39Ar cooling ages (ca. 47 Ma) on both sides of the contact argue against this interpretation (Gordon et al., 2010b). Thus, the Skagit-Napeeqa contact was overprinted by a high-temperature shear zone that was localized at this contact. Localization probably resulted from the rheological contrast between the Skagit orthogneiss and Napeequa rocks, possibly reactivating an older shear zone, and/or deformation may have transposed an intrusive contact.

### EOCENE DUCTILE DEFORMATION IN THE SWAKANE BIOTITE GNEISS

The second major domain of rocks with Eocene 40Ar/39Ar cooling ages and ductile deformation is south of the Skagit Gneiss Complex within the Cretaceous Swakane Biotite Gneiss. The Dinkelman décollement separates the main body of Swakane Biotite Gneiss from the structurally overlying Napeequa unit and...
91–72 Ma arc plutons (Fig. 2) (Matzel et al., 2004; Paterson et al., 2004). A narrow belt of Swakane Biotite Gneiss is also exposed southwest of the Entiat fault (Fig. 2). In the following we briefly summarize the major features of the gneisses that are most relevant to the Eocene evolution of the rocks, using previously published data.

The dominantly arkosic Cretaceous clastic protoliths of the Swakane Biotite Gneiss were buried and metamorphosed at peak conditions of ~640–750 °C at 9–12 kbar (Valley et al., 2003; Gatewood and Stowell, 2012). Sm-Nd garnet dates of 73.5 ± 1.2 to 66.8 ± 0.7 Ma from the gneiss (Gatewood and Stowell, 2012) and a U-Pb zircon age of 68.36 ± 0.07 Ma from a syntectonic leucogranite sheet (Matzel et al., 2004) are interpreted to date the amphibolite facies metamorphism. Hornblende 40Ar/39Ar and K-Ar dates of 579 ± 0.5 Ma and 50.8 ± 1.4 Ma, respectively, and biotite dates ranging from 49.5 ± 0.3 to 46.2 ± 1.5 Ma record the cooling of the Swakane Biotite Gneiss in the Chelan block (Tabor et al., 1987a: Matzel, 2004; Paterson et al., 2004). Some of the gneiss was at the surface by ca. 48 Ma, as it is basement to part of the Eocene Chumstick basin (e.g., Gresens et al., 1981; Evans, 1994).

Structures associated with burial were probably largely obscured by pervasive ductile, non coaxial shear that occurred at decreasing temperatures, presumably during exhumation (Paterson et al., 2004). Gently dipping foliations and the Dinkelman décollement are folded into a ≥10 km wavelength, gently northwest-plunging antiform. Lineation plunges gently to the north to north-northeast (Paterson et al., 2004; Miller et al., 2006). Mesoscopic fold hinge lines are more scattered, ranging from northwest to north-south, to southwest. Abundant kinematic indicators show that the gneiss records dextral shear on the southwest limb and sinistral shear on the northeast hinge of the antiform (Paterson et al., 2004). Restoration of foliation to shallower dip by unfolding of the antiform results in uniform top-to-the-north to top-to-the-north-northeast shear (Paterson et al., 2004). Latest major motion on the Dinkelman detachment is also top-to-the-north, oblique to the strike of the orogen (~315°–320°). Differences in 40Ar/39Ar and K-Ar cooling ages between the upper and lower plates of ~10–13 and 10–16 m.y. for hornblende and biotite, respectively, suggest that late motion was extensional and resulted in excision of ≥8 km of crust (Matzel, 2004; Paterson et al., 2004). undeformed dikes intrude both the hanging wall and footwall of the décollement, although we have not been able to trace any single dike across the structure. The two dated dikes in the footwall close to the décollement are 48.4 ± 2.2 Ma (K-Ar hornblende) and 47.8 ± 1.9 Ma (K-Ar biotite) (Tabor et al., 1987a). Displacement on the décollement probably occurred until ca. 48–47 Ma, as indicated by 40Ar/39Ar and K-Ar cooling ages.

**STRAIN PATTERNS FROM EPIZONAL EOCENE PLUTONS**

Shallow to middle crustal Eocene (ca. 48–45 Ma) plutons intrude, or are near, the Skagit Gneiss Complex, and include the Cooper Mountain, Duncan Hill, and Railroad Creek plutons (Figs. 2 and 3). These intrusions generally display weak solid-state deformation (our mapping), but overlap temporally with late ductile deformation of the complex. Orientations (Figs. 5A, 5B) of magmatic foliation and lineation, defined best by biotite and to a lesser extent plagioclase and hornblende in these plutons, may provide “snapshots” of the regional strain field during emplacement (e.g., Paterson et al., 1998). Foliations in the intrusions are in places discordant to pluton contacts and subparallel to host-rock structures and the regional structural grain (Figs. 5A, 5B), and/or intensify in strength independent of contacts. Thus, these fabrics are interpreted to record tectonic strain and compliment results from the Skagit Gneiss Complex and Swakane Biotite Gneiss.

On the southwest side of the Skagit Gneiss Complex are the markedly elongate, ca. 46–45 Ma granodioritic Duncan Hill and Railroad Creek plutons (Fig. 3) (Cater, 1982; Dellinger, 1996). The tilted Duncan Hill pluton (~2–13 km paleodepth) has a northwest-striking, vertical to steeply northeast-dipping, moderate to strong magmatic to solid-state foliation and northwest-trending lineation in its deeper northwest end (Figs. 5A, 5B) (Cater, 1982; Dellinger, 1996; our data) that has been interpreted to record regional strain by Haugerud et al. (1991). In comparison, our structural mapping indicates that the Railroad Creek pluton also has variable fabric intensity. In the south, the eastern margin of the pluton has a very weak magmatic foliation. A moderately strong magmatic fabric that is variably overprinted by solid-state deformation is found to the west, from the pluton interior to the margin. Foliation strikes north to northwest and dips steeply, and lineation trends northwest or southeast, and plunges gently (Figs. 5A, 5B). Lineation is stronger than foliation near the western contact. The strain gradient and solid-state deformation in the interior of the pluton are compatible with an origin by regional tectonic strain.

The western part of the ca. 48 Ma Cooper Mountain batholith (Shea, 2008) has a weak northwest-striking foliation where we have mapped it (Fig. 5A). Anisotropy of magnetic susceptibility data from the batholith generally show northwest-striking magnetic foliation and gently to moderately northwest- or southeast-plunging magnetic lineation, which has been interpreted to record regional tectonic strain (Fawcett et al., 2003).

In summary, magmatic and mostly weak solid-state foliation in Eocene plutons on average strike northwest, are steep, and probably record northwest-southwest shortening. Lineations are dominantly subhorizontal and northwest or southeast trending, and likely indicate weak northwest-southeast tectonic stretch.

**EOCENE FAULTS**

Most major map-scale Eocene faults cutting and bounding the Cascades core and Eocene basins strike north to northwest and are steep. The most prominent is the right-lateral Straight Creek-Fraser fault, which forms the western boundary of the Cascades core (Figs. 1 and 2) and has displaced core rocks ~110–160 km southward from equivalent rocks in southern British Columbia (e.g., Misch, 1977b; Vance and Miller, 1981; Tabor et al., 1984; Monger,
boundary of the Chumstick basin and separates it from the Eocene Swauk For-
ing point, the hinge line of the folded contact between the Swakane Biotite
gneiss and Napeequa unit (Fig. 2).

It has been interpreted as a dextral strike-slip fault largely on the basis of its
length, stratigraphic relations, and rapid depositional rates of the Chumstick
basin (e.g., Johnson, 1985; Evans, 1994). Cheney and Hayman (2009) sug-

The steep north-northwest–striking Leavenworth fault forms the western
boundary of the Chumstick basin and separates it from the Eocene Swauk For-

The northwest-striking Entiat fault is the eastern boundary of the Eocene
basin (e.g., Johnson, 1984). The fault probably has

The Swauk and Chuckanut strata were openly to tightly folded. Folds in
the Swauk Formation mostly have wavelengths of 100 m to 2 km and steep
axial planes (Fig. 6C) (Tabor et al., 1982, 2000; Doran, 2009). Axial traces of
these folds change from northwest in the west, to mainly east-southeast or
west-northwest in the central region, to largely northwest in the easternmost
part of the Swauk Formation (Tabor et al., 1982; our mapping; Fig. 7).

The Eocene nonmarine basins (Figs. 2 and 7) that flank the Cascades core
consist of very thick sequences of sandstone and mudstone, with variable
amounts of conglomerate and local tuff (e.g., Johnson, 1984). Most important
for comparison to the processes occurring in the deeper crust are the Chucka-
ut, Swauk, and Chumstick sequences (Fig. 2), which overlap in age with the
Eocene crystalline rocks.

The Chuckanut Formation is west of the Skagit Gneiss Complex, but before
dextral slip on the Straight Creek fault the formation was farther south, west
of the Swauk Formation (e.g., Vance and Miller, 1981). A tuff deep in the Chucka-
ut section is 56.835 ± 0.087 Ma, and an intercalated felsic tuff (Silver Pass
Volcanic Member) in the Swauk Formation is 51.364 ± 0.067 Ma (U-Pb zircon,
CA-ID-TIMS; Eddy et al., 2016).

Figure 7. Simplified map emphasizing
Eocene sedimentary and volcanic rocks,
adjacent units, and Eocene faults, dikes,
and folds in the Swauk basin of the cen-
tral Washington Cascades. Fm.—forma-
tion. C-C is the line of the cross section in
Figure 6C. Lines 1–4 are transects where
numerous dike orientations were meas-
ured and extensions were calculated (see
Table 3). Dikes and some folds are from our
mapping, and other folds and contacts are
modified mainly from Tabor et al. (1982,
2000).
the Teanaway Formation gives a U-Pb zircon age of 49.341 ± 0.070 Ma (Eddy et al., 2016). Conformably overlying the Teanaway lavas are clastic rocks of the mid(?) to late Eocene Roslyn Formation.

The Swauk basin is juxtaposed on the east against the Chumstick basin by the Leavenworth fault, and the Entiat fault forms the northeast boundary of the Chumstick basin (Fig. 7). These two faults have components of dextral slip (Tabor et al., 1984). The Chumstick Formation contains numerous tuffs (McClinicy, 1986); one from near the base of the section is 49.147 ± 0.073 Ma and another ~6000 m higher in the lower-middle part of the section is 47.981 ± 0.059 Ma (U-Pb zircon; CA-ID-TIMS; Eddy et al., 2016), indicating very high sedimentation rates of ~6–7 m/k.y. during the interval between eruption of these tuffs. This depositional rate, the large thickness of the lower half of the formation, migrating depocenters, and monolithologic breccias containing large (averaging >1 m) clasts, some of which are >25 km from their source, provide strong evidence that the Chumstick Formation accumulated in a transtensional basin (Evans, 1994). Eddy et al. (2016) showed that the onset of rapid sedimentation in the basin corresponds to a region-wide reorganization of sedimentation patterns and the onset or acceleration of movement on the regional strike-slip faults.

**EOCENE DIKES**

Dike swarms give insights into the regional strain field in the shallow crust. Felsic to mafic Eocene dikes widely intrude Cascades core rocks and the Eocene Swauk Formation.

The best studied, most abundant, and southernmost dikes are in the ca. 49.3 Ma (U-Pb date for Teanaway lavas) mafic Teanaway dike swarm, which primarily intrudes the Swauk Formation (Fig. 8; Foster, 1958; Tabor et al., 1984; Doran, 2009). This swarm extends ~75 km from east to west and a maximum of 18 km from north to south (Fig. 7). We measured orientations and thicknesses of 392 Teanaway dikes throughout the basin. In 4 detailed transects in the western and central parts of the swarm (Fig. 7), average strikes range from 036° to 017° (Table 3), dips range from 72° to 62°, and thicknesses are from 20 to 12 m. By comparison, 145 dikes intruding the eastern part of the basin have a mean strike of 040°, dip of 78°, and thickness of 14 m (Mendoza, 2008). Average subhorizontal extension directions inferred for the dikes range from 306°–126° to 287°–107° in the western and central regions, to 310°–130° in the east. Minimum extensions of ~16% and 43% were determined from 2 of the strike-normal transects across the western and central Swauk basin (Table 3). These extension values were calculated from the number and thickness of dikes over the length of a transect.

Orientations of Eocene dikes (Fig. 9) that intrude the Cascades core are less well documented. The few dated dikes that intrude the Chelan block are ca. 48–46 Ma (K-Ar; Tabor et al., 1987a). Granite and rhyolite dikes next to the shallow part of the ca. 46 Ma Duncan Hill pluton commonly strike northeast (Tabor et al., 1987a; Delliger, 1996; Bryant and Miller, 2014). Northwest-striking dikes have also been reported from southeast of the pluton (Waters, 1926; Tabor et al., 1987a; Sylva and Miller, 2014). Granitic dikes next to the southern margin of the ca. 48 Ma Cooper Mountain batholith on average strike 005° (Raviola, 1988). Our study of dikes scattered throughout the Skagit Gneiss Complex indicates that they have a broad range of modal compositions and textures, are typically 10 cm to 2 m thick, and some are younger than ca. 48.2 Ma ortho-
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| Transect | Number of dikes | Mean trend (°) | Transect length (km) | Total dike thickness (km) | Extension (%) |
|----------|----------------|----------------|----------------------|--------------------------|--------------|
| 1        | 33             | 036            |                      |                          |              |
| 2        | 36             | 028            | 5.7                  | 0.77                     | 15.5         |
| 3        | 86             | 021            | 7.8                  | 2.08                     | 36.3         |
| 4        | 38             | 017            |                      |                          |              |
| Combined | 193            | 024            |                      |                          |              |

Note: Excludes dikes dipping <50°; does not include data in eastern part of the dike swarm. Transect locations shown in text Figure 7.

Interaction of Shallow and Deeper Crustal Levels

In the North Cascades, a relatively short time interval (ca. 53–45 Ma) was marked by (1) plutonism, metamorphism, partial melting, ductile flow, and rapid cooling of metamorphic rocks; (2) rapid subsidence and deposition in nonmarine basins; (3) brittle faulting; and (4) extensive dike intrusion in the shallow crust (Fig. 10; Table 2). In contrast to some orogens (e.g., Tauern window; Selverstone, 1988; Axen et al., 1998), where the upper rigid crust was not strained while the middle to lower crust underwent ductile flow, different crustal levels of the North Cascades underwent coeval deformation, suggesting a degree of coupling between the crustal layers.

Compatibility of Strain in the Ductile and Brittle Crust

The presence of both Eocene brittle and ductile structures allows further evaluation of structural patterns and the degree of strain coupling between different crustal levels during regional transtension (Fig. 10). In particular, our discussion focuses on deformation styles, extension directions, and noncoaxial shear at different depths, and the Eocene folding that involved the Swauk basin and the Cascades crystalline core.

Mesoscopic structures in some of the Skagit rocks are likely composite, as foliation in some places predates 65 Ma leucosomes and elsewhere deforms widespread 49–45 Ma orthogneisses (see preceding discussion). The concordance of fabrics in Eocene orthogneisses with those in older orthogneisses implies that Eocene fabrics are widespread. Fabrics reflecting ductile transtension include the combination of gently to moderately dipping foliations, gently plunging lineations, L > S fabrics in sizable (>10 km²) domains, and strong stretching of fold hinges parallel to lineation (Fig. 4B) (e.g., Robin and Cruden, 1994; Tikoff and Teyssier, 1994; Dewey, 2002; Fossen et al., 2013), most prominently in the ca. 45 Ma Diablo Lake orthogneiss bodies. The extension direction in the Skagit Gneiss Complex is largely northwest-southeast (~330°–150°), except near segments of the Ross Lake fault zone.

Noncoaxial ductile shear in the Skagit Gneiss Complex is localized in weaker metasedimentary rocks, as described above. Both top-to-the-northwest and top-to-the-southeast shear occur in the northern domain with top-to-the-northwest shear more common, particularly in dated Eocene orthogneisses. This shear likely predated upright folding, and thus originally probably occurred on gently dipping surfaces. Overall, we infer that on the map scale, deformation was general shear marked by a component of top-to-the-northwest noncoaxial shear.

The ductile ca. 50–48 Ma strain of the Skagit Gneiss Complex was broadly coaxial with ductile dextral-normal shear in the Ross Lake fault zone, which was top-to-the-northeast in the north and top-to-the-east-southeast in the south (Table 2) (Miller and Bowring, 1990; Haugerud et al., 1991), and the approximately north-south stretching in the dilational stepover of the Ross Lake fault. Eocene deformation may have overlapped earliest movement on the Straight Creek fault (Table 2).

The Swakane Biotite Gneiss is characterized by pervasive, top-to-the-north to north-northeast noncoaxial shear and north-south to north-northeast–south-southwest subhorizontal to moderately plunging stretching (Figs. 9 and 10). The fabrics thus indicate largely subvertical shortening and top-to-the-north subhorizontal flow prior to the late upright folding; some of the flow overlapped with the Eocene deformation in the Skagit rocks (Table 2).
Figure 9. Summary map of Eocene structures and extension directions in the North and Central Cascades, including dikes, stretching lineations, magmatic lineations in plutons, and faults. Eocene plutons (orange with names), dikes, and basins (yellow) are highlighted, as are the Skagit Gneiss Complex and Swakane Biotite Gneiss (purple). NQ—Napeequa unit; RLF—Ross Lake fault; RRC—Railroad Creek pluton; HW—hanging wall; Mtn—mountain.
The two deep-crustal domains in the Chelan block were exhumed in similar manners. The rapid cooling and, presumably, exhumation of the Skagit Gneiss Complex are compatible with vertical thinning associated with subhorizontal stretching and constrictional strain, and a component of normal slip on the dextral Ross Lake fault system. Similarly, exhumation of the Swakane Biotite Gneiss was broadly synchronous with vertical ductile shortening, and probably with dip slip on the Entiat fault.

The overall pattern of ductile Eocene deformation in the Cascades core thus includes subvertical shortening and subhorizontal northwest-southeast to north-south stretching in the Skagit and Swakane rocks; top-to-the-north to north-northeast noncoaxial shear in the Swakane Biotite Gneiss, and less prominent top-to-the-northwest shear in the Skagit Gneiss Complex; and dextral-normal shear in the Ross Lake fault zone (Figs. 9 and 10). The displacement direction in the fault zone is generally oblique to stretching in the gneisses.

The orientation of the stretching lineation relative to the Ross Lake fault zone and the overall trend of the orogen suggest that a simple transtensional model cannot be applied. In a dextral wrench mode, stretching should be oriented ~45° from the strike of the fault zone, at ~095°–275°, whereas in partitioned transtension it should be broadly perpendicular, ~050°–230° (e.g., Tikoff and Teyssier, 1994). Near a major strike-slip zone, high strains may rotate fabrics into subparallelism with the zone, but this seems unlikely to affect patterns at distances of kilometers from the zone. The stretching direction must be influenced by other poorly understood factors, such as along-strike changes in crustal thickness (see following).

Transtension in the brittle crust was manifested by dextral strike-slip faults, widespread dike swarms, and subsidence and thick accumulation of clastic sediments in nonmarine basins (Figs. 9 and 10). Orientations of structures indicate complex partitioning of deformation. Dextral strike slip on the northwest-striking Ross Lake fault zone (Foggy Dew, Ross Lake faults) probably overlapped with motion of the similarly oriented Entiat fault and the north-south–striking Straight Creek–Fraser fault (Fig. 1; Table 1). The north-west-striking faults have components of down-to-the-southwest (Entiat fault) and down-to-the-northeast to east (Ross Lake and Foggy Dew faults) displacement. Orientations of the Eocene dikes indicate that extension was dominantly west-northwest-east-southeast to northwest-southeast, but in detail markedly
variable (Fig. 9). The average strike of the Teanaway dike swarm is ~30° to that of the Straight Creek fault, which is an appropriate orientation for the strain field associated with a transtensional dextral wrench fault (e.g., Sanderson and Marchini, 1984) (Fig. 7).

The magmatic fabrics in Eocene plutons that are interpreted to record regional strain formed at depths intermediate between those of the metamorphic rocks and the dikes. Magmatic lineations trend northwest or southeast and are gently to moderately plunging. These orientations are broadly similar to those in the ductile crust.

Extension directions from ductile and brittle structures and the sense of noncoaxial shear in different areas define a complex Eocene strain field (Figs. 9 and 10). Overall, the ductilely deformed rocks underwent largely orogen-parallel to orogen-oblique, subhorizontal flow. The shallow crustal extension direction inferred from dikes is also oblique to the strike of the orogen and to the flow direction in deeper crustal rocks. Differences in extension between the dikes and Skagit and Swakane rocks average ~20°–25° and 65°–70°, respectively (Fig. 9).

**Compatibility of Folding of the Swauk Formation and Skagit Gneiss Complex**

A distinctive component of the deformation history is the analogous geometry and timing of Eocene folds of the Swauk Formation and Skagit Gneiss Complex that may also be in part synchronous with upright folding of foliation in the Swakane Biotite Gneiss and the Dinkelman décollement. Folding is bracketed between ca. 51 and 49 Ma in the Swauk Formation, and the latest folding occurred ca. 49–48 Ma in the southern part of the gneiss complex and continued until at least 48 Ma (ending by 45 Ma) in the north. Fold geometries are broadly analogous in the Swauk and Skagit rocks (Fig. 6). In both settings, axial traces of the larger folds curve considerably (Figs. 5C and 7). In the Swauk rocks, the trend may result in part from the orientation of boundaries of the basin, and in the west, from rotation due to strike slip on the Straight Creek fault. In the Skagit rocks, the late moderately to steeply plunging folds of foliation described here are difficult to explain. Such large folds with steep axes are commonly thought to not extend into the deep crust (cf. Johnston and Acton, 2003) and may be truncated by a detachment.

The origin of this short-lived folding is uncertain. Folding may occur during transtension, but the associated shortening is limited (Fletcher and Bartley, 1994; Fossen et al., 2013). If the folds are related to transtension, the low angle (<20°) of many of the hinge lines to the northwest-trending, dextral strike-slip faults (Fig. 5C) that involve the Cascades core suggests that the divergence angle is low (cf. Fossen et al., 2013). The folding more likely reflects a short-lived episode of regional shortening that interrupted the overall transtensional regime. In one interpretation, folding results from collision and accretion of the Siletz-Crescent terrane (Siletzia) (Miller and Umhoefer, 2013; Eddy et al., 2016), which is a large igneous province west of the North Cascades (Fig. 2). This collision has been inferred to account for Eocene deformation in southwest Oregon and Vancouver Island (Fig. 2), where the terrane boundary is marked by thrusting and folding (Johnston and Acton, 2003; Wells et al., 2014). Collision in southwest Oregon occurred between ca. 51 and 47 Ma.

**What Drove Orogen-Parallel Extension?**

In the Paleogene, ca. 55 Ma, there was a fundamental change from regional transpression to transtension in the western Cordillera at the latitude of the Coast Mountains and North Cascades (e.g., Parrish et al., 1988; Miller and Bowring, 1990). This swing in deformation regimes is interpreted as a response to a change in the motion between the North American and Farallon/Kula plates (e.g., Engebretson et al., 1985). In addition, a slab window may have formed during postulated subduction of the Farallon-Kula ridge or Farallon-Resurrection ridge beneath the North American plate ca. 50 Ma (e.g., Breitsprecher et al., 2003; Madsen et al., 2006).

The orogen-parallel to orogen-oblique extension in the brittle and ductile levels of the Washington Cascades indicates that plate motion was not simply partitioned into dextral strike slip parallel to the northwest-trending plate boundary and extension normal to the boundary. Furthermore, orogenic collapse did not result in extension normal (northeast-southwest) to the strike of the arc, as in the metamorphic core complexes to the east.

We postulate that orogen-parallel to orogen-oblique ductile stretching was driven by a combination of dextral shear along the plate margin and along-strike differences in crustal thickness and temperature (cf. Paterson et al., 2004). Crustal thickness of the Cascades core may have decreased from north to south as a result of differences in the magnitudes of mid-Cretaceous shortening. Shortening was driven by eastward underthrusting of the rigid Insular superterrane (McGroder, 1991). In addition, the contractional belt probably terminates at the southern end of the Coast Mountains–North Cascades arc (cf. Monger et al., 1982, 1994), leading to less thickening at this end of the orogen. Much of the southern part of the Cascades core, represented by the Nason terrane, had cooled below biotite K-Ar and 40Ar/39Ar closure temperatures by 80 Ma (e.g., Tabor et al., 1982, 1987a; Matzel, 2004), in marked contrast to the Skagit Gneiss Complex, where upwelling of hot asthenosphere in the inferred slab window may have helped localize late metamorphism, magmatism, and partial melting. The top-to-the-north shear in the Swakane Biotite Gneiss and top-to-the-northwest flow in some of the Skagit Gneiss Complex suggests that during at least part of this deformation, the hot and weak middle to deep crust flowed southward from the region of 79–45 Ma magmatism, metamorphism, and partial melting in the Skagit Gneiss Complex toward the region of thinner and cooler crust in the south. We speculate that this crustal flow is broadly similar to the central Andes where orogen-parallel lower crustal flow is interpreted to have occurred in response to differences in upper crustal shortening in the Altiplano and Puna Plateau (e.g., Hindle et al., 2005; Ouimet and Cook, 2010). Similarly, deep crustal flow discordant to shallower level extension is also thought to have been driven by differences in crustal thickness in the Ruby Mountains metamorphic core complex of Nevada (MacCready et al., 1997).
**Relationships to Metamorphic Core Complexes to the East**

The North Cascades and Omineca metamorphic core complexes to the east were sites of thick (>55 km) crust in the Late Cretaceous–Paleogene, and underwent extension, cooling, and exhumation in the Eocene (e.g., Parrish et al., 1988; Miller and Paterson, 2001; Paterson et al., 2004). Both regions also contain migmatites with similar Eocene melt-crystallization ages (e.g., Vanderhaeghe et al., 1999; Crowley et al., 2001; Hinchey et al., 2006; Kruckenberg et al., 2008; Gordon et al., 2008, 2010a). The two areas are currently connected by a regionally flat Moho, with crustal thicknesses of ~30 km (Cook et al., 1992). Based on these characteristics, Whitney et al. (2004) postulated that the Omineca and North Cascades metamorphic and plutonic belts were dynamically coupled and represent the exhumed eastern and western margins, respectively, of an orogenic plateau.

Pervasive Eocene flow and kinematics recorded in gneisses and mylonites associated with the major detachment of the Shuswap metamorphic core complex of the Omineca belt (Fig. 11) indicate largely east-west extension (e.g., Read and Brown, 1981; Brown and Journeay, 1987; Carr et al., 1987; Parrish et al., 1988; Teyssier et al., 2005). In the southwestern part of this complex, however, the Okanogan dome records west-northwest–east-southeast to northwest-southeast extension (Kruckenberg et al., 2008). In comparison, ductile flow was more northwest-southeast to north-south in the Cascades core. We infer that these differences in part reflect the position of the thick crust relative to the plate margin. In much of the Shuswap complex, extension was probably driven by orogenic collapse in a largely extensional regime (e.g., Teyssier et al., 2005), whereas in the western part of the Okanogan dome and North Cascades, dextral shear associated with the obliquity of the plate boundary modified the strain field and resulted in transtensional flow. Any linked deep crustal flow beneath the hypothesized orogenic plateau must thus be marked by a major swing in the direction of flow. In the Omineca metamorphic core complexes, a broad zone of ductilely deformed rocks has been kinematically and thermally linked for at least part, if not all, of the deformation history to the structurally higher detachment faults during exhumation of the footwall. This relationship may also apply to the Swakane Biotite Gneiss, but it is not apparent for the Skagit Gneiss Complex. Instead, exhumation and dextral shear in the gneiss complex were in part coeval with the dextral-normal shear in the Ross Lake fault system.

**CONCLUSIONS**

The North Cascades orogen exposes a wide range of crustal levels that recorded Eocene transtension differently. The largest domain of deep crust exhumed in the Eocene, the Skagit Gneiss Complex, reveals a long record of magmatism, metamorphism, and partial melting. The complex has gently to moderately dipping foliation, gently northwest- and southeast-plunging lineation, and sizable constrictional domains, all of which formed in part during dextral-normal ductile and brittle shear in the bounding Ross Lake fault system. The other deep-seated domain, the Swakane Biotite Gneiss, records pervasive Eocene top-to-the-north to north-northeast noncoaxial shear. Strain patterns in the Swakane and Skagit rocks are compatible with transtension and orogen-parallel to orogen-oblique subhorizontal stretching.

The shallow crust in the Eocene was marked by dextral and normal slip on regional northwest- and north-south-striking high-angle faults, basin subsidence, rapid deposition, and widespread intrusion of basaltic to rhyolitic dikes. On a regional scale, extension directions from dikes are highly variable, but are mostly oblique to the trend of the orogen.

The brittle and ductile levels of the North Cascades crust were deformed coevally, but record different strain geometries and extension directions. These differences illustrate the problems with extrapolating strain patterns from one
crustal level to another level and using these patterns to interpret ancient plate motions. These differences may reflect the complex regional and local boundary conditions. Ductile flow of the hot and weak Skagit Gneiss Complex and Swakane Biotite Gneiss was probably in response to regional strike slip and extension, vertical shortening and collapse, and along-strike differences in crustal thicknesses in the Cascades core. In particular, the variation in crustal thickness and the presence of thick, hot and weak crustal layer in the northern part of the core may have led to the differences in extension directions.

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