Petrology, geochemistry, and provenance of the Galice Formation, Klamath Mountains, Oregon and California

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ABSTRACT

The Upper Jurassic Galice Formation of the Klamath Mountains, Oregon-California, overlies the ca. 162-Ma Josephine ophiolite and the slightly younger Rogue–Chetco volcano-plutonic arc complex. The Galice Formation that overlies the Josephine ophiolite consists of a siliceous hemipelagic sequence, which grades upward into a thick turbidite sequence. Bedded hemipelagic rocks and scarce sandstone, however, also occur at several localities within the Josephine ophiolite pillow basalts. Corrected paleoflow current data suggest that the Galice Formation was derived predominantly from the east and north. Detrital modes of sandstones from the Galice Formation indicate an arc source as well as a predominantly chert-argillite source with minor metamorphic rocks. A sandstone located ~20 m below the top of the Josephine ophiolite has detrital modes and heavy mineral suites similar to the turbidite sandstones. Detrital Cr-spinel compositions from the turbidite and intra-pillow lava sandstones are also similar, indicating supra-subduction zone mantle peridotite and volcanic sources. Published detrital zircon data from a turbidite sandstone chiefly give a bimodal age distribution of 153 Ma and ca. 227 Ma but with a minor Proterozoic component. Whole-rock geochemistry from intra-pillow lava sedimentary rocks, the hemipelagic sequence, and the turbidites suggest a mixture between mafic and crustal sources. It is suggested that the source area for the intra-pillow lava sedimentary rocks, hemipelagic sequence, and turbidites resulted from the mixing of arc and accreted terranes. These data indicate that the source areas for the Galice Formation were already established by ca. 162 Ma, probably during a Middle Jurassic orogeny that predated formation of the Josephine basin.

Keywords: Upper Jurassic, Galice Formation, hemipelagic sequence, turbidites, provenance, detrital zircons, Josephine ophiolite, California, Oregon

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INTRODUCTION

The Upper Jurassic Galice Formation occurs within the western Klamath terrane (western Jurassic belt) of the Klamath Mountains, Oregon-California (Fig. 1). It consists predominantly of slate and lesser metasandstone, but volcanic (predominantly volcaniclastic) members are locally present. The type section of the Galice Formation conformably overlies both the ca. 160–157-Ma Rogue–Chetco volcano-plutonic arc complex and ca. 164–162-Ma Josephine ophiolite (Harper et al., 1994; Fig. 1). A lower hemipelagic sequence is present where the Galice Formation overlies the Josephine ophiolite (Fig. 2). The Galice Formation also occurs within the Elk outlier (Fig. 1), where it is in fault contact with sheeted dikes that are correlated with the Josephine ophiolite (Giaramita and Harper, this volume).

The western Klamath terrane is the youngest of a series of eastward-dipping, imbricated thrust sheets within the Klamath Mountains (Irwin, 1964, 1966, 1994; Burchfiel and Davis, 1981). The roof thrust of the western Klamath terrane, the Orleans thrust (Fig. 1), is a major crustal boundary with geological and geophysical data suggesting 40 to >100 km of displacement, respectively (Jachens et al., 1986). The basal thrust, the Madstone Cabin thrust (Fig. 1), juxtaposes the Josephine ophiolite over the Chetco intrusive complex of similar age (Dick, 1976; Harper et al., 1994; Yule, 1996). The Galice Formation is regionally metamorphosed from prehnite-pumpellyite to lower greenschist facies (Harper et al., 1988) and is unconformably overlain by Lower Cretaceous nonmarine and marine conglomerate and sandstone (Harper et al., 1994).

The Galice Formation and its basement formed in a supra-subduction setting, as indicated by the presence of the underlying Rogue–Chetco arc complex and members of similar lithology in the Galice Formation (Garcia, 1982) and a high proportion of arc detritus in metasandstones (Snake, 1977; Harper, 1984; Wyld and Wright, 1988; Harper et al., 1994). The Galice basin apparently formed as a result of extension within a west-facing arc, ultimately resulting in seafloor spreading that produced the Josephine ophiolite (Snake, 1977; Saleeby et al., 1982; Harper and Wright, 1984; Wyld and Wright, 1988; Harper et al., 1994; Yule et al., this volume). Seafloor spreading was postulated to have occurred in a back-arc basin behind the Rogue–Chetco arc, but the presence of boninites in the Josephine ophiolite and the apparently slightly younger age of the Rogue–Chetco arc led Harper (2004) to suggest that arc rifting and initial seafloor spreading took place in the fore-arc. Similarly, the modern Lau back-arc basin apparently first formed in the fore-arc, but became situated in the back-arc after a trenchward jump in the arc axis (Parson and Wright, 1996). Collapse of the basin occurred during the Late Jurassic Nevadan orogeny by underthrusting of the Galice Formation and its basement beneath older terranes of the Klamath Mountains, resulting in low-grade regional metamorphism (Snake, 1977; Saleeby et al., 1982; Harper and Wright, 1984). Geochronologic and biostratigraphic data suggest that the Galice Formation was deposited...
during the Nevadan orogeny, perhaps in a trenchlike basin along the western edge of the Klamath Mountains (Harper et al., 1994). The large displacement along the Orleans thrust suggests that the Josephine ophiolite was partially subducted during the Nevadan orogeny.

This chapter focuses on the provenance of the Galice Formation, based on new and existing data for detrital modes, chemistry of detrital Cr-spinel, whole-rock sedimentary geochemistry, and ages of detrital zircon. In addition, a comparison is made between the upper Galice Formation and clastic sedimentary rocks in the basal hemipelagic sequence overlying and interbedded with the Josephine ophiolite, including a newly discovered sandstone bed within the pillow lava unit.

**GALICE FORMATION**

Diller (1907) originally named the section of slate, sandstone, and minor conglomerate exposed on Galice Creek in southwest Oregon, west of Grants Pass, the “Galice Formation” (Fig. 1). Wells and Walker (1953) mapped the Galice 7.5-min quadrangle and recognized volcanic members lithologically similar to the underlying Rogue Formation. Cater and Wells (1953) correlated sedimentary and volcanic rocks in northern California (Gasquet 7.5-min quadrangle) with the Galice Formation, and Irwin (1960, 1966) showed the Galice Formation extending the entire length of the Klamath Mountains province. Although volcanic members are present in the type Galice Formation, Vail (1977) and Harper (1980, 1984) showed that a volcanic member mapped by Wells et al. (1949) and Cater and Wells (1953) is actually the upper part of a complete ophiolite complex (Josephine ophiolite). Volcanic members are generally absent in the Galice Formation that overlies the Josephine ophiolite (Harper, 1984), but one is present within the Galice Formation overlying the Devils Elbow remnant of the Josephine ophiolite in the southern Klamath Mountains (Wylde and Wright, 1988).

Harper (1980, 1994) suggested that the Galice Formation can be subdivided into three units (Fig. 2). The lower unit represents a hemipelagic sequence, whereas the upper unit is comprised of a turbidite sequence consisting of interbedded sandstone, siltstone, and radiolarian argillite. Lying between these two units is an ~55-m-thick transitional unit that represents the transition from hemipelagic to turbidite deposition (Fig. 2).

The ages of the Josephine ophiolite and Rogue–Chetco arc, which conformably underlie the Galice Formation, were initially thought to be the same (ca. 157 Ma; Saleeby et al., 1982; Saleeby, 1984). Based on this, Harper (1984) and Pinto-Auso and Harper (1985) included the hemipelagic sequence directly overlying the Josephine ophiolite within the Galice Formation. Subsequent biostratigraphic (Pessagno and Blome, 1990) and radiometric data (Harper et al. 1994) indicate, however, that the hemipelagic sequence is older than the type Galice Formation. This finding led Pessagno et al. (1993, 2000) to exclude the hemipelagic sequence (their “volcanopelagic succession”) from the Galice Formation. Pessagno et al. (1993, 2000) suggested a hiatus occurs ~50 m stratigraphically above the Josephine ophiolite that separates the turbidite part of the Galice Formation from the underlying hemipelagic sequence (Fig. 2). They cited the much poorer preservation of radiolarians below this contact in support of their interpretation. They further suggested that the hemipelagic sequence is lithologically and genetically unrelated to the overlying turbidite; this suggestion, however, is inconsistent with petrographic and geochemical data presented below. Our observations show that bedding in rocks beneath the apparent hiatus is disrupted over a thickness of ~10 m below the hemipelagic-turbidite transition (Fig. 2). Beds are discontinuous, but not folded, and are cut by numerous small low-angle faults. This disruption appears to have occurred before complete lithification (Harper, this volume), because deformed beds and some small faults are cut by ca. 150-Ma syn-Nevadan dikes, some of which have amoeboid margins suggestive of intrusion into wet sediment (Harper, this volume).

For the purposes of this chapter, we include the hemipelagic sequence along with the turbidite in the Galice Formation (sensu lato), because the transitional unit (Fig. 2) is clearly lithologically gradational in character between the hemipelagic sequence and the Galice turbidite. Furthermore, as discussed below, sandstones in the transitional unit and in the lower part of the basal turbidite are very similar to the scarce sandstones that occur within both the hemipelagic sequence and pillow lavas of the Josephine ophiolite.
Structures in the Galice Formation that formed during the Late Jurassic Nevadan orogeny include slaty cleavage and associated overturned folds, stretching lineations, fibrous extension veins, and small thrust faults (Kays, 1968; Snoke, 1977; Harper, 1980, 1984; Norman, 1984; Gray, 1985, this volume; Wyld, 1985; Harper, this volume). The slaty cleavage and associated flattening of sand and pebble grains varies from very weak, as in the area of Cave Junction, Oregon (Jones, 1988; Fig. 1), to very strong. Strain data (i.e., Harper, 1980; Cashman, 1988; Jones, 1988) show a southward increase in strain from the Cave Junction area into northern California, coincident with an increase in metamorphic grade from prehnite-pumpellylite to lower greenschist facies (Harper, 1980; Harper et al., 1988). Paleomagnetic data suggest that the Galice Formation in southwestern Oregon may have undergone clockwise rotation of as much as 100° (Schultz and Levi, 1983; Bogen, 1986; Harper and Park, 1986).

Sedimentary Rocks within the Josephine Ophiolite

Pure, light-green chert is common between pillows in pillow lavas throughout the Josephine ophiolite. Although uncommon, bedded sedimentary rocks up to 5 m in thickness locally occur within the pillow lava unit of the Josephine ophiolite (Pinto-Auso and Harper, 1985; Kuhns and Baitis, 1987; Harper et al., 1988; Zierenberg et al., 1988; Figs. 2 and A-1). Radiolarians from these rocks indicate a late Callovian age, consistent with the ca. 162-Ma age of the Josephine ophiolite (Pessagno et al., 1988; Zierenberg et al., 1988). These sedimentary rocks are interbedded with Josephine lavas as well as separate mineralized lavas and massive-sulfide deposits from overlying lava flows (Pinto-Auso and Harper, 1985; Kuhns and Baitis, 1987; Zierenberg et al., 1988). Although most of the intra-pillow lava sedimentary rocks are bedded, some in the Turner-Albright mine area on the Oregon-California border southwest of Cave Junction, Oregon, are diamicrites that probably formed as debris flows (Fig. A-1). These deposits and the presence of abundant talus breccias suggest the presence of fault scarps during formation of the Josephine ophiolite (Kuhns and Baitis, 1987; Zierenberg et al., 1988). The diamicrites are very similar in hand sample, and in terms of geochemistry and color (black and green; Fig. A-1) to samples analyzed from the hemipelagic sequence. They are dominantly argillites, with green and black varieties interbedded on millimeter to centimeter scales. Where the regional metamorphic grade is lower-greenschist facies, they are slaty. Petrographic observations and geochemistry indicate that they are a mixture of mud, radiolarians, and hydrothermal sediment (Pinto-Auso and Harper, 1985; Kuhns and Baitis, 1987; Zierenberg et al., 1988; Figs. 2 and A-1). Less common rock types in the bedded sequences in the ophiolite include red radiolarian argillite (or slate) and chert. Detrital muscovite observed in black argillite lamella indicates a terrigenous component. The green argillite lamellae are probably tuffaceous, as shown for similar, better-studied green argillites in the hemipelagic sequence (Pinto-Auso and Harper, 1985).

An uncommon 2- to 10-cm-thick sandstone bed was found beneath the uppermost lava flow of the Josephine ophiolite, ~20 m below the depositional contact with the hemipelagic sequence in the type section of the pillow lava unit (Harper et al., 2002; MacDonald et al., 2004; Fig. 2). Harper (1994) originally identified this sandstone as a tuff, based on its characteristics in hand sample, including green color, but subsequent petrographic observations indicate the presence of terrigenous detritus, as discussed below. The green color is apparently due to ocean-ridge hydrothermal alteration, which has affected the entire pillow lava unit up to the depositional contact at this locality (Harper, 1995). Several meters along strike of where it was originally discovered, this bed has a medium-gray color similar to sandstones in the overlying Galice Formation (Harper et al., 1988).

Hemipelagic Sequence

The hemipelagic sequence is ~45-m thick and consists predominantly of green to black slaty radiolarian argillite with lesser green to black radiolarian chert (Pinto-Auso and Harper, 1985; Pessagno et al., 2000). Argillite in the hemipelagic sequence is a mixture of radiolarians, terrigenous and tuffaceous detritus, and hydrothermal sediment (Pinto-Auso and Harper, 1985; Kuhns and Baitis, 1987; Harper et al., 1988). The tuffaceous component is commonly evident from the presence of angular plagioclase, quartz (some bipyramidal; Fig. 3A) and, uncommonly, altered glass, silt, and sand (Fig. 3B). A terrigenous component from the presence of detrital muscovite is evident in thin section. The Fe/(Fe + Mn + Al) ratios indicate a relatively small metalliferous component derived from hydrothermal springs (Fig. 4), even for those overlying massive-sulfide deposits of hydrothermal origin (Kuhns and Baitis, 1987; Zierenberg et al., 1988).

Elevated metal contents are common in samples from the hemipelagic sequence, especially 8–23 m above the depositional contact with the underlying Josephine ophiolite, and some samples have sufficiently high Fe/(Fe + Mn + Al) to be classified as metalliferous (Pinto-Auso and Harper, 1985; Fig. 4). In the field, metalliferous sedimentary rocks are indicated by a shiny manganiferous coating or, less commonly, by a red color. The geochemistry and petrography of the metalliferous samples indicate that they all have substantial clastic detritus and variable amounts of radiolarians (Pinto-Auso and Harper, 1985). In addition, most of the rocks in the hemipelagic sequence, as well as those interbedded with pillow lavas of the Josephine ophiolite, have Fe/(Fe + Mn + Al) that is elevated relative to terrigenous sediments (Fig. 4), implying a small metalliferous component. Metalliferous sediments typically directly overlie ophiolites and modern ocean crust (e.g., Barrett et al., 1987). Pinto-Auso and Harper (1985) suggested that the stratigraphic
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Figure 3. (A) Photomicrograph of a tuffaceous chert showing bipyramidal (volcanic) quartz and plagioclase. A 0.25-mm scale is located on figure. Crossed polarizers. (B) Photomicrograph of tuffaceous radiolarian argillite from the hemipelagic sequence showing well-preserved, cone-shaped Radiolaria (arrow), altered vesicular glass (left center), and angular quartz grains (clear). Plane light. Horizontal scale is 2 mm. (C) Photomicrograph of metasandstone from the Galice Formation showing felsic volcanic clast with resorbed quartz phenocryst. Colorless high-relief grain below this clast is augite. Note foliation in the matrix. Horizontal scale is ~1.5 mm. (D) Photomicrograph of metasandstone from the Galice Formation showing a clast of quartz-mica schist. At the lower left of this grain is a mafic volcanic clast. Dark seams are parallel to foliation. Horizontal scale is ~0.8 mm.

position of the metalliferous sediments, 8–23 m above the Josephine ophiolite, was the result of low-temperature, off-axis springs analogous to those along the flanks of the Galapagos Rise. Harper (2003) proposed that the metalliferous horizon was the result of the passage of a propagating spreading center, with the metalliferous component representing distal fallout from high-temperature, on-axis hot springs. The massive-sulfide deposits show that hydrothermal springs that could produce Fe- and Mn-rich metalliferous sediments were present during eruption of the Josephine pillow lavas. Thus, the low metal content of intra-pillow lava sedimentary rocks, including those directly overlying the massive sulfide deposits, and of sedimentary rocks comprising the lower 8 m of the hemipelagic sequence is apparently the result of dilution by abundant terrigenous mud.

Sandstone and scarce volcanic pebbly mudstone are uncommon rock types in the hemipelagic sequence (Pinto-Auso and Harper, 1985; Harper, 1994; Pessagno et al., 2000). The pebbly mudstone occurs ~40 m above the depositional contact with the Josephine ophiolite. It contains volcanic clasts up to 20 cm in diameter and has a radiolarian-tuffaceous matrix containing biotite (Harper, 1994). Sandstone was observed at two localities within the hemipelagic sequence; one within an isolated outcrop but the other ~7–8 m above the contact with the Josephine ophiolite (Pinto-Auso and Harper, 1985; Fig. 5). The sandstone beds are graded (Fig. 5) and compositionally very similar to those in the basal part of the turbidite sequence, with ~95% of the grains consisting of lithic volcanic rock fragments, plagioclase, and clinopyroxene. Other clasts, especially in the very coarse fraction, are predominantly chert and siliceous argillite. Scouring of underlying chert is evident along the base of one of these sandstone beds; this bed also contains a 3-cm angular rip-up clast of underlying chert (Fig. 5). Pessagno and Blome (1990) also reported the presence of “pelagic limestone,” but these occur as nodules and probably formed by replacement of radiolarian argillite. Formation of the nodules occurred before formation of slaty cleavage (i.e., diagenetic) because radio-
faulting or to submarine landsliding prior to deposition of the hemipelagic-turbidite transition unit.

**Transition Zone**

An ~55-m-thick transition zone exists between the lower hemipelagic sequence and the upper turbidite of that part of the Galice Formation that overlies the Josephine ophiolite (Harper, 1994; Fig. 2). This zone consists predominately of radiolarian argillite with minor sandstone. The proportion of radiolarians and the proportion and size of clasts (mostly silt) are generally higher than those of radiolarian argillites in the hemipelagic sequence, and there is a negligible hydrothermal component (Pinto-Auso and Harper, 1985). Radiolarian-bearing limestone nodules occur throughout the transition zone, being abundant at the top of the zone (Harper, 1994). The lower contact for the transition zone is defined only for one section located along the Middle Fork of the Smith River, a sequence that includes the type section of the pillow lava unit of the Josephine ophiolite (Harper, 1994, 2003). In this section, which was studied by Pessagno and Blome (1990), as well as by Pinto-Auso and Harper (1985), the base of the transition zone is marked by the lowest occurrence of sandstone beds above the Josephine ophiolite.

Pessagno and Blome (1990) included rocks of the transition zone in the Galice turbidite, but we reserve use of the term “turbidite” for rocks above what we call the transition zone (Fig. 2). The above authors used the lowest occurrence of sandstones, as

Figure 4. Plot illustrating apparent mixing relationships for sedimentary rocks within Josephine ophiolite, hemipelagic sequence, and Galice Formation. Mixing curves are for East Pacific Rise (EPR), hydrothermal sediment, and Pacific pelagic clay (PC), and EPR hydrothermal sediment and mid-ocean ridge basalt (MORB; values from Barrett, 1981). Also shown is field for Margi umbers, Cyprus, which are pure metalliferous sediment (Ravizza et al., 1999). Figure modified from Pinto-Auso and Harper (1985) and Harper et al. (1988).

Figure 5. Graded bed from hemipelagic sequence of the Galice Formation. Bed beneath sandstone is a radiolarian chert. On the back side of this sample there is a 3-cm subangular rip-up clast of underlying sediment located within sandstone.
well as the presence of the hiatus discussed in the preceding paragraph, to define the contact between the hemipelagic sequence (their “volcanopelagic unit”) and the transition zone (their “Galice Formation sensu lato”). As noted above, a sandstone bed occurs within the pillow lava unit of the ophiolite (at this locality); at other localities, sandstone beds occur within the hemipelagic sequence (Fig. 5). Thus, the use of the first sandstone bed to define the base of the transition zone in the section (Fig. 2) is only valid for this particular location. We interpret the contact to be gradational, with an upward increase in proportion of sandstone beds and decreasing proportion of radiolarians in argillites.

According to the biostratigraphy of Pessagno et al. (1993, 2000), the transition zone ranges in age from middle Oxfordian to middle Kimmeridgian. Recently, Miller et al. (2003) estimated the age of deposition of a massive sandstone from the base of the overlying turbidite sequence to be ca. 153 Ma, based on the age of the youngest detrital zircon age spectrum (Miller et al., 2003). Thus, deposition of the transition zone appears to have ranged from ca. 157 to ca. 153 Ma.

**Turbidite of the Galice Formation**

The base of the turbidite sequence, as defined in this chapter, occurs ~100 m above the top of the Josephine ophiolite at the locality represented in Figure 2. The Galice Formation in its type area, where it overlies the Rogue Formation, consists entirely of turbidite (i.e., no hemipelagic sequence) that interferes with various volcanic members. The Galice turbidite consists predominantly of slate and thin-bedded to massive metasandstone, along with scarce pebble conglomerate. Sandstones are predominantly feldspathic litharenites, are commonly graded, show scouring along their bases, and display partial to complete Bouma sequences. Load and flame structures, sole marks, and mud rip-up clasts are locally common. Trace fossils from slates include *Chondrites*, *Cosmophorae*, and *Spirobranchus*, which indicate abyssal, or perhaps bathyal, water depths (A.A. Ekdale, written commun., 1980). In the section represented in Figure 2, and extending at least 10 km farther north and 1 km farther south, a 200- to 300-m-thick massive sandstone unit overlies the transition zone between the hemipelagic and turbidite sequences (Cater and Wells, 1953; Harper, 1980). The thickness of the turbidite cannot be accurately determined, due to repetition by folding and faulting, but the original stratigraphic thickness was probably at least several kilometers, based on its large outcrop area (Fig. 1).

The Galice turbidite, in both its type area and where it overlies the Josephine ophiolite, contains the bivalve *Buchia concentrica*, which has a known age range of middle Oxfordian to late Kimmeridgian (Imlay et al., 1959; Imlay, 1980). The age of the base of the turbidite based on the biostratigraphy of Pessagno et al. (1993, 2000) is middle Kimmeridgian. The absolute age of the turbidite is constrained by the ca. 153-Ma age of the youngest detrital zircons (Miller et al., 2003) and the age of crosscutting dikes, sills, and small plutons, which are as old as ca. 150 Ma (Harper et al., 1994).

**PALEOFLOW DATA**

Paleoflow data were obtained by Park-Jones (1988) for the type Galice Formation overlying the Rogue Formation west of Grants Pass, Oregon, and by us for the Galice Formation overlying the Josephine ophiolite where it is exposed in Rough and Ready Creek, southwest of Cave Junction, Oregon. Folds in the Galice Formation usually are plunging, so paleoflow data were restored by first rotating the fold axis (usually bedding-cleavage intersection) to horizontal, followed by rotation of bedding to horizontal. The resulting bidirectional and unidirectional flow directions are plotted in Figure 6. Predominantly west paleocurrent directions (in present-day coordinates) were found for the type Galice Formation (Park-Jones, 1988; Fig. 6), compared to predominantly north directions for the Galice Formation overlying the Josephine ophiolite (Fig. 6). The paleocurrent data probably need to be corrected for clockwise rotation of the Klamath Mountains, which has been suggested to explain paleomagnetic declination anomalies (Mankinen and Irwin, 1982; Schultz and Levi, 1983; Bogen, 1986; Renne and Scott, 1988; Smith and Harper, 1993). Some workers (e.g., Mankinen and Irwin, 1982; Bogen, 1986) argued that the entire Klamath Mountains rotated as a single block, whereas others have argued that the arcuate shape of the Klamath Mountains is the result of oroclinal bending, possibly related to formation of the Columbia embayment (Renne and Scott, 1988; Saleeby and Harper, 1993). The strike of slaty cleavage at both localities for which there are paleocurrent data is ~020°–030° (Park-Jones, 1988; this study). Figure 6B and D show the paleocurrent data after correcting for an assumed clockwise rotation of 65°, which assumes the western Klamath terrane had an original trend of ~340°, parallel to the overall structure of correlative rocks in the Sierra Nevada foothills, which have not been rotated (Bogen et al., 1985; Frei, 1986). The inferred ~65° rotation for the Galice Formation in southwest Oregon is less than the ~100° and ~78° clockwise rotations inferred for a volcanic member in the type Galice section and for the Grants Pass pluton that intrudes the Galice Formation (Schultz and Levi, 1983; Bogen, 1986), respectively, and the ~100° for the Josephine ophiolite (Smith and Harper, 1993). The actual amount of clockwise rotation may be less than suggested by the paleomagnetic data because of unknown amounts of post-Nevadan tilting, observed in Cretaceous sedimentary rocks elsewhere in the west-central and northwest Klamath Mountains, and possibly because of insufficient magnetic cleaning (Harper and Park, 1986; Renne and Scott, 1986).

**DETRITAL MODES**

Sandstones within the Galice Formation are generally lithic wackes and feldspatholithic wackes (Snoke, 1977; Harper, 1980; Wyld, 1985). Sandstones are generally foliated (e.g., Fig.
Detective modes for Galice sandstones were determined by Harper (1980), Norman (1984), and Wyld (1985). Medium- to coarse-grained sandstones were point-counted following the procedures of Dickinson (1970) and the terminology of Ingersoll and Suczek (1979) as well as Ingersoll et al. (1984). The detrital modes should be considered semiquantitative, due to partial recrystallization and, in many samples, deformation during low-grade metamorphism. Matrix generally constitutes >20%, and in some samples >30%, of the counted points. Such matrix contents are much higher than those for modern turbidites and are probably the result of alteration of lithic fragments (e.g., Dickinson, 1970, 1985). Quantitative modal analysis is also compromised by the textural similarity of many felsic volcanic rock fragments to chert; only Harper (1980) used feldspar staining to remove this uncertainty.

Structural complexity of the Galice Formation does not allow for accurate determination of stratigraphic height above basement. Harper (1980) and Wyld (1985), however, had sufficient stratigraphic control to broadly group their samples into basal, lower, and middle-upper.

Harper (1980) found that 62–92% of the volcanic rock fragments in samples of Galice sandstone have microlitic textures that are typical of andesites. Altered glass (replaced by chlorite) ranges from ~1–23% of volcanic clasts, although most samples have <10%. The proportion of felsic clasts is highly variable, ranging from 3 to 33% of volcanic clasts. Clasts with lathtwork texture (mafic) are much less common, never comprising >4% of the volcanic clasts.

The lithic- and feldspathic-rich nature of the sandstones is evident on various triangular diagrams (Fig. 7). Some sandstones plot as moderately quartz-rich on the quartz-feldspar-lithic (Q–F–L) diagram (Fig. 7A), but chert dominates the Q fraction, and monocrystalline quartz (Qm) contents are low, as evident on the monocrystalline quartz-feldspar-lithic diagram (Qm–F–L; Fig. 7B). Polycrystalline quartz (Qp), consisting of quartz sandstone, and quartz siltstone, is generally a minor component (Fig. 7C), but abundant chert and argillite results in high values of lithic sedimentary clasts (Lsm in Fig. 7D; Lsm = sedimentary lithic + metasedimentary lithics, Lv = volcanic lithic). Detrital modes of sandstones for the Galice Formation from its type area (Harper, 1980), near the Oregon-California border (Harper, 1980; Norman, 1984), and from where it overlies the Devils Elbow remnant of the Josephine ophiolite in the southern Klamath Mountains, are similar (Fig. 7). Those samples overlying the Devils Elbow remnant, however, tend to be less rich in lithic fragments (Fig. 7A and B) and have more monocrystalline quartz (Fig. 7B).

A trend of increasing proportion of total quartz (Q; mostly chert), Qm, Qp, and Lsm (mostly siliceous argillite and chert) with increasing stratigraphic height is evident in Figure 7 for samples overlying the Devils Elbow remnant of the Josephine ophiolite in the southern Klamath Mountains (Wyld, 1985). A similar trend is evident for the Galice Formation overlying the Josephine ophiolite in its type area in northwestern California.
Figure 7. Triangular plots of point-count data for sandstones from Galice Formation in its type area (Harper, 1980), where it overlies the Josephine ophiolite near the Oregon-California border (Harper, 1980; Norman, 1984), and overlying Devils Elbow remnant of the Josephine ophiolite in the southern Klamath Mountains (Wyld, 1985). Q—monocrystalline quartz + polycrystalline quartz + chert; F—feldspar (all plagioclase in these samples); L—unstable lithics; Q_m—monocrystalline quartz; L_t—total polycrystalline lithic fragments, including stable quartzose; Q_p—chert + polycrystalline quartz; L_v—lithic volcanics; L_sm—sedimentary + metasedimentary lithics. (A) Q–F–L diagram after Dickinson et al. (1983). Shaded field is for correlative rocks in the Sierra Nevada foothills, including the Mariposa Formation (Behrmann and Parkinson; 1978). (B) Q_m–F–L diagram after Dickinson et al. (1983). (C) Q_p–L_v–L_sm diagram after Dickinson (1985). Inferred end-member sources for Galice sandstones are indicated. (D) L_sm+Q_p–F–L_v diagram.
bias is evident in individual sandstone beds that have pebbly volcanic clasts than in associated sandstones. Such a grain-size near 0 to ~20%. This result implies a much lower proportion of which consists almost essentially of volcanic clasts, varies from zose sandstone clasts (Q; Fig. 8). The igneous component, erates are rich in chert (C) and generally have very few quart clasts in the 0.5–6 cm diameter range. Galice conglomerate (1985) counted clasts larger than 1 mm, whereas Seiders (1991) obtained detrital modes for conglomerates from the Galice For tion by point counting (Fig. 8). In their modal analysis, Wyld obtained detrital modes for conglomerates from the Galice sandstones, but this may be due to the replacement of plagioclase by prehnite.

Wyld (1985) and Seiders (1991, written commun., 1992) obtained detrital modes for conglomerates from the Galice Formation by point counting (Fig. 8). In their modal analysis, Wyld (1985) counted clasts larger than 1 mm, whereas Seiders (1991) counted clasts in the 0.5–6 cm diameter range. Galice conglomerates are rich in chert (C) and generally have very few quartz sandstone clasts (Q; Fig. 8). The igneous component, which consists almost essentially of volcanic clasts, varies from near 0 to ~20%. This result implies a much lower proportion of volcanic clasts than in associated sandstones. Such a grain-size bias is evident in individual sandstone beds that have pebbly bases. Two of Wyld’s (1985) conglomerate samples from the lower Galice Formation in the southern Klamath Mountains have much higher Q and are transitional between typical Galice Formation chert-rich conglomerates and correlative conglomerates of the Sierra Nevadan foothills (Fig. 8).

HEAVY MINERALS

Occurrence

Snoke (1972, 1977), Harper (1980, 1984), and Wyld (1985) reported heavy-mineral assemblages in the Galice Formation. The most common heavy minerals are zircon (euhedral to very well rounded), tourmaline, apatite, biotite, muscovite, Cr-spinel, and, in volcanic-rich basal turbidite sandstones, clinopyroxene and hornblende. Less common to rare are garnet, epidote, staurolite, and glaucophane. The 2- to 5-cm-thick sandstone bed within the Josephine pillow lava unit has abundant clinopyroxene (~5 modal %), with lesser hornblende, Cr-spinel, and rare glaucophane (violet to clear pleochroism); zircon has not yet been found in this sample; these heavy minerals, as well as the detrital modes, are similar to those of volcanic-rich sandstones within the hemipelagic-turbidite transition zone and basal turbidite.

Cr-Spinel

McLennan et al. (1993) suggested that Cr-spinel most likely indicates an ophiolitic source area. Spinels can occur in any primitive mafic volcanic rocks, however, as well as in any type of ultramafic rock. Cookenboo et al. (1997) and Lee (1999) stressed the importance of using the chemical composition of detrital Cr-spinels to infer rock types in the source area. Lee (1999) suggested that Cr-spinels are significant in paleogeographic reconstructions because their deposition occurs within proximity of their source area.

Spinels from four sandstones were analyzed for this study. One sample is from the bed within the Josephine ophiolite pillow lava unit (sample PC-2b; Table 1), another sample from the basal Galice turbidite overlying the ophiolite (sample LJC-23; Table 1 and Fig. 2) and the other two samples from the turbidite sequence (samples DB-1 and CJ-7; Table 1 and Fig. 2). The Cr-spinels were analyzed with a JEOL 733 Superprobe at the Department of Earth and Environmental Sciences, Rensselaer Polytechnic Institute, Troy, New York, and with a JEOL 8900 electron microprobe at Binghamton University, Binghamton, New York. The elements Ti, Al, Cr, Mn, Mg, Ni, and Fe were analyzed using a 15-kV accelerating voltage, 15-nÅ beam current, and a 1-μm beam diameter. Elements Al, Cr, Mg, and Fe were counted for 40 seconds whereas Ti, Mn, and Ni were counted for 100 seconds each. Sample USNM 117075 from Tiebaghi Mine, New Caledonia, was used as an Al, Cr, Mg, and Fe standard at Rensselaer Polytechnic Institute. Other elemental standards used at Rensselaer Polytechnic Institute were rutile for Ti, tephroite for Mn, and diopside glass for Ni. Analy-

Figure 8. Conglomerate point count data for the Galice Formation and the Mariposa Formation in the central Sierran foothills illustrating provenance differences. I—igneous rocks; Q—quartz sandstone and quartzite; C—chert; Fm—formation. Data from Wyld (1985) and Seiders (1991, written commun., 1992).
sis of the standard USNM 117075 was done throughout the session at Rensselaer Polytechnic Institute to assess proper calibration. Standards used at Binghamton University were TiO₂ for Ti, Al₂O₃ for Al, chromite for Cr, spessartine garnet for Mn, MgO for Mg, Ni-metal for Ni, and hematite for Fe.

The compositions of detrital spinels from the sandstone within the Josephine ophiolite pillow lava unit and the sandstone from the basal turbidite are given in Table 1 and are plotted in Figures 9 and 10. The detrital spinels show a wide range of compositions. Some plot entirely within the field for mantle peridotites, whereas others plot entirely in the field for volcanic spinels, including the mid-ocean ridge basalt (MORB) field and between the MORB and ocean-island—or within-plate—basalt (OIB) fields (Fig. 9). Other detrital spinels plot in the area of overlap for volcanic and mantle peridotite spinels (Fig. 9).

Those detrital spinels identified as originating from mantle peridotites in Figure 9 have Cr/(Cr + Al) ratios (Cr§) that overlap those of abyssal (mid-ocean ridge origin) and suprasubduction zone (SSZ) peridotites (Fig. 10A). All, however, have Mg/(Mg + Fe²⁺) ratios that are lower than those for abyssal peridotite spinels; the Mg/(Mg + Fe²⁺) ratios for all the detrital mantle spinels fall within the field for SSZ peridotites,

### Table 1. Representative Analyses of Cr-Rich Spinel

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<th>TiO₂</th>
<th>Al₂O₃</th>
<th>Cr₂O₃</th>
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<td>0.57</td>
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</table>

Note: The ferric iron content of each analysis was determined by assuming stoichiometry, following the methods of Barnes and Roeder (2001).

*From Figure 2.

†Sample analyzed at Rensselaer Polytechnic Institute.

§Sample analyzed at Binghamton University.

b.d.—below detection limits.
a result that implies that they are all derived from SSZ ophiolites (Fig. 10A).

Those detrital spinels identified as volcanic (MORB to MORB-OIB) in Figure 9 have Cr/(Cr + Al) ratios similar to the upper range of most modern MORB, but are shifted to lower Mg/(Mg + Fe²⁺) ratios (Fig. 10B). This shift to lower Mg/(Mg + Fe²⁺) ratios is probably due, in part, to the OIB component that is indicated in Figure 9, as OIB basalts have Mg/(Mg + Fe²⁺) ratios that extend to much lower values. However, it may also be due to crystal fractionation of olivine, which causes the Mg/(Mg + Fe²⁺) of spinel to increase (Dick and Bullen, 1984; Kamenetsky et al., 2001; Fig. 10B).

Figure 10C shows those detrital Cr-spinels that could have been derived from either mantle peridotite or volcanic rocks based on the classification shown in Figure 9. Most of these spinels, as well as some of the peridotite spinels (Fig. 10A), have high Cr/(Cr + Al) ratios, indicative of high degrees of partial melting, which is typical of arcs (Dick and Bullen, 1984). Four of the detrital spinels from the turbidite sample and one from the intra-pillow sample have Cr/(Cr + Al) ratios of ~0.8, which is characteristic of boninites (Fig. 10C) and mantle that has undergone very high degrees of melting or mantle that has interacted with a boninitic magma (e.g., Dick and Bullen, 1984; Harper, 2004).

**AGE OF SOURCE AREAS**

The age of detrital zircons within Galice sandstones was studied by Miller and Saleeby (1995) and Miller et al. (2003)
through U/Pb age dating. Miller and Saleeby (1995), using multiple grain fractions from various localities throughout the Galice Formation, obtained U-Pb data that plot as a chord on a concordia diagram, with an average upper intercept of ca. 1600 Ma and a lower intercept of ca. 215 Ma, although the youngest zircons approach ca. 170 Ma. Surprisingly, Miller and Saleeby (1995) found no difference in the age distributions of euhedral and well-rounded (recycled) zircon populations. All four of their Galice samples show the same spread on a concordia diagram, which implies no major variations in the age of detrital zircons in these samples. A fifth sample gives an older upper-intercept age; however, this sample is not from the Galice Formation but from the Lems Ridge olistostrome that underlies the Galice, which probably represents a fragment of older Klamath basement in the western Klamath terrane that is correlative with the lower Mesozoic Rattlesnake Creek terrane (Ohr, 1987).

Recently, Miller et al. (2003) reported ion-microprobe single-crystal U/Pb ages for detrital zircons from a Galice Formation sandstone near the base of the turbidite sequence at a Buchia concentrica locality. Most zircons from this sample are Late Jurassic, averaging ca. 153 Ma. Zircons of ca. 227 Ma are common, and a few Paleozoic and Proterozoic zircons were found. The high proportion of volcanic clasts in the sandstone, including altered glass, indicates erosion of an active arc; thus, the ca. 153-Ma age most likely approximates the time of turbidite deposition, consistent with the Late Jurassic age of the Galice turbidite.

The Nd and Sr isotopic values for slates from the Galice Formation turbidite sequence are rather uniform and indicate derivation from both Precambrian continental rocks and young mantle-derived rocks (Frost et al., this volume). These workers also found that the continental isotopic signal is greater in slates than in sandstones of the turbidite sequence, a relationship that may be consistent with differences in the whole-rock geochemistry discussed below.

Two chert pebbles from the Galice Formation were dated as Late Triassic using radiolaria (D.L. Jones, personal commun., 1979; Harper, 1980). One of the pebbles is from a volcanic-rich sandstone near the base of the turbidite sequence overlying the Josephine ophiolite, and the other pebble is from a conglomerate near the base of the type Galice Formation overlying the Rogue Formation.

**GEOCHEMISTRY**

Major- and trace-element geochemistry of sedimentary rocks can provide invaluable information about the original tectonic setting and the provenance of a sedimentary unit (Roser and Korsch, 1986; McLennan et al., 1993). Unlike petrographic methods for tectonic discrimination that are limited to using coarser grain sizes, chemical analyses of sedimentary rocks have no grain-size restrictions (Roser and Korsch, 1986). McLennan (1989) notes, however, that concentrations of heavy minerals in sandstones can affect rare-earth-element (REE) abundances and

![Figure 11. Provenance discriminant-function diagram of Roser and Korsch (1988). Only samples from the Galice turbidite are plotted. Siliceous argillite and chert from the hemipelagic sequence and Josephine ophiolite are not plotted because of their Fe-rich hemipelagic component. Also plotted are the average values for North American shale composite (NASC), post-Archean Australian shale composite (PAAS), upper continental crust (UC; Gromet et al., 1984; McLennan, 1989; McLennan, 2001), and fields for modern back-arc basin turbidites (McLennan et al., 1990) and volcaniclastic sediments from the Lau back arc (Bednarz and Schmincke, 1994). Symbols are the same as in Figure 5.](image-url)
ratios, and that abundances of elements such as rare earths are diluted by the greater concentration of quartz in sandstones. Biogenic CaCO₃ or SiO₂ can also dilute concentrations of other elements within a sample (Roser and Korsch, 1986, 1988). Diagenetic reactions related to burial can affect silicate and calcite abundances of sandstones (Galloway, 1974). Thus, caution should be used when making interpretations based on geochemical analyses of sedimentary rocks, and ratios should be used when possible. Because we do not have CO₂ values for our analyses, we cannot correct for dilution resulting from CaCO₃.

We have, thus, followed the advice of Roser and Korsch (1986, 1988) and recalculated major-element values on a CaO-free basis before plotting on some diagrams (e.g., Fig. 11).

Published and unpublished geochemical data for the Galice Formation are available from other studies (Coleman, 1972; Pinto-Ausio and Harper, 1985; Kuhns and Baitis, 1987; Park-Jones, 1988; Zierenberg et al., 1988; Barnes et al., 1995; Frost et al., this volume; Table 2). In addition, we report a chemical analysis for a sandstone sample (D25C) that was analyzed along with samples of Park-Jones (1988) by X-ray fluorescence at McGill University. Only major-element data of Coleman (1972) are used, as his trace-element data are semiquantitative; sample

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<th>TABLE 2. SELECTED ANALYSES OF GALICE FORMATION SAMPLES</th>
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<tr>
<td><strong>LOI</strong></td>
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Notes: JO—overlies Josephine ophiolite; LOI—loss on ignition; n.d.—no data.
*Data from Park-Jones (1988) except sample D25c, which is previously unpublished.
†Fe₂O₃TT—total iron as Fe₂O₃.
If all trace elements reported, analysis of Ba, Rb, Sr, Pb, Y, Zr, Ni, Cr, and V by X-ray fluorescence; others by instrumental neutron activation analysis.
B6C from Pinto-Auso and Harper (1985) was not used because of its extremely high CaO concentration.

A metalliferous component in sediments is best detected using the ratio Al/[Al + Fe + Mn], which is <0.35 for modern metalliferous sediments and >0.5 for nonmetalliferous pelagic clay (Bostrom and Peterson, 1969; Bostrom, 1973). This component can be derived either from hydrothermal springs or precipitation from seawater (as with manganese nodules). Negative, rather than positive, Ce anomalies on REE patterns for samples overlying and within the Josephine ophiolite indicate a hydrothermal origin. Figure 4, from Pinto-Auso and Harper (1985), shows that siliceous argillite and chert samples from the hemipelagic sequence and pillow lava unit of the Josephine ophiolite fall along a mixing line between modern terrigenous and low-temperature hydrothermal sediments. As discussed above, samples from within the pillow lava unit of the Josephine ophiolite and from the lower 8 m of the hemipelagic sequence are not metalliferous, although most have a metalliferous component (i.e., Al/[Al + Fe + Mn] <0.5); in fact, the least metalliferous hemipelagic sediments actually occur within the pillow lavas. This is somewhat surprising, because stratiform massive-sulfide deposits of hydrothermal origin are present in the

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Josephine ophiolite (Kuhns and Baitis, 1987; Zierenberg et al., 1988), but are apparently the result of dilution by abundant terrigenous detritus, as indicated by Figure 4. “Off-axis” metaliferous sediments are present 8–23 m above the base of the hemipelagic sequence (Fig. 2), either due to precipitation from low-temperature, off-axis springs (Pinto-Auso and Harper, 1985) or from distal fallout of high-temperature springs at a younger, propagating spreading center (Harper, 2003). All slate and sandstone samples from the Galice turbidite, including those from the Elk outlier, have \( \text{Al/}(\text{Al}+\text{Fe}+\text{Mn}) > 0.5 \) (Fig. 4), indicating a negligible metaliferous component.

Interpretations about source area and tectonic setting based on sediment geochemistry can be made from several plots (Figs. 11, 12, and 13). The majority of the turbidite samples, including those from the Elk outlier, plot in the intermediate-igneous and mafic-igneous provenance fields on Figure 11 (Roser and Korsch, 1988), which uses linear equations of major-element ratios except for \( \text{CaO} \) and \( \text{SiO}_2 \) (because they can be elevated by biogenic contributions). The remaining turbidite samples plot in and near the quartzose sedimentary provenance field (Fig. 11). Samples from the hemipelagic sequence and pillow lava unit are not plotted on Figure 11 because their high-Fe hydrothermal components cause these samples to plot far to the left. Figure 11 also shows that Galice turbidite samples fall within the field for back-arc basin turbidites of McLennan et al. (1990) and are displaced away from average continental crust (UC) and sediments largely derived from continental crust (NASC and PAAS in Fig. 11).

Figure 12 uses the trace elements Th, Zr, and Sc, which are generally immobile during metamorphism. This diagram is useful in that sediments typical of passive continental margins have elevated Zr/Sc ratios due to zircon enrichment through sediment recycling (McLennan et al., 1993; Fig. 12). Active-margin samples, however, plot along a trend between mafic and continental sources. Figure 12 shows that the Galice Formation samples plot along this trend, indicating insignificant sediment recycling, although the presence of well-rounded detrital zircons in sandstones indicates some recycling has occurred (Harper, 1980; Miller and Saleeby, 1995). Figure 11 suggests that the Galice samples were derived, on average, from sources that are more mafic than typical continental crust; perhaps from the mixing of continental and intermediate arc material. The chert sample from the hemipelagic sequence above the Josephine pillow lavas plots at significantly higher Zr/Sc values compared to other samples (Fig. 12); although the nonbiogenic component in the chert should consist of terrigenous or tuffaceous material similar to that in the hemipelagic rocks and thus plot along the same trend as other Galice samples. Hence, it may not be valid to plot cherts on this diagram (McLennan et al., 1993). Moreover, the analytical uncertainties for Zr and Sc are relatively high for this particular sample since they are so diluted by \( \text{SiO}_2 \).

A more complete comparison of Galice samples with post-Archean Australian shale (PAAS), derived largely from continental sources, can be made by normalizing immobile trace and rare-earth elements to PAAS (Fig. 13). McLennan (1989) suggested that samples should be normalized to PAAS rather than to North American shale composite (NASC), due to the possible inclusion of abnormal samples in NASC. Reference samples are shown in Figure 13A and B for comparison with those from the Galice Formation. The NASC and average upper continental crust generally have flat PAAS-normalized patterns (Fig. 13A), reflecting the dominantly continental sources for the shale.

Figure 12. \( \text{Th/Sc} \) vs. \( \text{Zr/Sc} \) plot modified from McLennan et al. (1993) by the addition of average upper continental crust (representing cratonic sources; McLennan, 2001), a modern island-arc tholeiite (representing mafic sources; Pearce et al., 1995), and the field for modern back-arc basin turbidites (McLennan et al., 1990). Symbols are the same as in Figure 4.
composites. A Cyprus umber, which is essentially pure metalliferous sediment, shows enrichment with respect to PAAS, especially in Ni and V, except for negative Ce and Cr anomalies (Fig. 13A). The back-arc basin sediment from the Celebes Sea and an average island-arc tholeiite (IAT) have remarkably similar PAAS-normalized patterns (Fig. 13B); both are depleted in the light REEs compared to PAAS, have negative Hf, Th, and Cr anomalies, and have positive Sc anomalies (Fig. 13B). Back-arc basin sediment from the Japan Sea has a flatter PAAS-normalized pattern and a positive Ba anomaly (Fig. 13B), which is probably due to a large continental, or calc-alkaline volcanic, source component; this sample displays a slight negative slope from La through Eu, a result that is similar to the IAT sample and the Celebes Sea back-arc basin sediment (Fig. 13B).

Galice Formation slate from the type area and from the turbidite sequence overlying the Josephine ophiolite has similar patterns when normalized to PAAS (Fig. 13C). They show depletion in light REEs, a negative Th anomaly, and a positive $P_2O_5$ anomaly. A sandstone from the basal part of the turbidite overlying the Josephine ophiolite has a similar pattern to the slates, except for a negative Ba anomaly and positive Hf, Sc, and Cr anomalies (Fig. 13D). The Cr anomaly is probably from detrital Cr-spinel, as it is common in thin section, and the Sc anomaly is probably from the presence of detrital augite, which is abundant in this and most sandstone from the basal turbidite overlying the Josephine ophiolite (Harper, 1980).

The intra-pillow siliceous argillite, which is interbedded with pillow lavas of the Josephine ophiolite, has a similar pattern to slate from the Galice Formation turbidite (Fig. 13C), suggesting a similar terrigenous source. This sample, however, has
a negative rather than a positive \( P_2O_5 \) anomaly; however, because only one of the slates has a significant \( P_2O_5 \) anomaly and the sandstone has none, \( P \) may have been during diagenesis and metamorphism. The pattern for the sandstone sample (Fig. 13D) is shifted downward relative to the slates, a relationship that is readily explained by dilution from biogenic silica, which is evident in thin section from the presence of radiolarians. As expected, this dilution effect is even greater for the chert from the hemipelagic sequence overlying the Josephine ophiolite (Fig. 13C). The pattern for the chert is similar to that for the intra-pillow siliceous argillite (Fig. 13C), suggesting a similar terrigenous component, although the chert has a prominent positive Ba anomaly of unknown origin (seawater?), similar to the metalliferous sediment. The metalliferous sediment sample has a pattern that has similarities to both the slates of the Galice turbidite and to the Cyprus umber (e.g., negative Ce and Cr anomalies); this observation is consistent with the conclusion of Pinto-Auso and Harper (1985) that the metalliferous sediments in the hemipelagic sequence all have substantial terrigenous components, as evident from elevated Al/(Al + Fe + Mn) ratios (Fig. 4).

**DISCUSSION**

**Provenance**

Plots of detrital modes (Fig. 7), sediment geochemistry (Figs. 11, 12, and 13), and Nd and Sr isotopes (Frost et al., this volume) show trends for Galice turbidite samples that can be explained by mixing between two sources. The abundance of volcanic rock fragments in sandstones and conglomerates (Figs. 7 and 8) indicates that one of these sources is volcanic, and this source is almost certainly a magmatic arc. The non-arc source could represent basement of the arc, or a geographically distinct continental-like source.

The microlitic texture of the most volcanic clasts suggest that the arc source was composed primarily of andesite, although the abundance of detrital clinopyroxene (up to 5 modal %) is perhaps more consistent with a basaltic-andesite or basalt source. A negative slope for light REE, negative Th anomaly, and possibly more consistent with a basaltic-andesite or basalt source.

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from the Galice Formation where it overlies the Josephine ophiolite (Harper, 1980; Fig. 6D). The Klamath Mountains contain abundant chert and siliceous argillite, much of it Triassic (like two pebbles from the Galice Formation), as well as metamorphic and ophiolitic rocks (e.g., Irwin, 1966, 1972; Davis et al., 1978; Saleeby and Busby-Spera, 1992). The basement to all but the eastern Klamath arc rocks, including the Rogue–Chetco arc (Yule, 1996; Yule et al., this volume), is the early Mesozoic Rattlesnake Creek terrane (Wright and Wyld, 1994; Hacker et al., 1995). The Rattlesnake Creek terrane contains Triassic chert as well as mafic volcanic rocks that could have supplied MORB to MORB-OIB affinity detrital Cr-spinel (Fig. 9), although OIB affinity volcanic rocks are also present in the Sawyers Bar terrace of the central Klamath Mountains (Hacker et al., 1993). Units in the Klamath Mountains that could have supplied metamorphic detritus include the greenschist- to amphibolite-facies central metamorphic belt (e.g., Irwin, 1966; Davis et al., 1978). As for detrital glaucophane, the blueschist-facies Stuart Fork Terrane (Hotz et al., 1977) and Skookum Gulch Schist (Cotkin et al., 1992), as well as blueschist blocks in mélange of the Eastern Hayfork unit in the Sawyers Bar terrane (Wright, 1982; Hacker and Ernst, 1993), could have been the source(s). The large supra-subduction zone Trinity ophiolite could have supplied detrital peridotitic and boninitic Cr-spinel (Quick, 1981; Metcalf et al., 2000; Fig. 10).

Possible contributors of arc detritus from east of the Galice Formation include Middle Jurassic rocks in the eastern Klamath Mountains (e.g., Renne and Scott, 1988), the ca. 177- to 168-Ma Hayfork arc (western Hayfork terrane and related plutonic rocks; Wright and Fahan, 1988), and the ca. 159- to 165-Ma Wooley Creek plutonic belt (e.g., Hacker et al., 1995). These are not the appropriate age to have provided the ca. 153-Ma detrital zircons in the Galice Formation, nor are they likely to have yielded volcanic glass that is abundant in the volcanic-rich sandstones of the basal turbidite. Three plutons along the northeastern margin of the Klamath Mountains, however, may be the subvolcanic remnants of a ca. 153-Ma zircon source. Irwin and Wooden (1999) reported 154- and 156-Ma U/Pb zircon ages for the Jacksonville and White Rock plutons in the northern Klamath Mountains, respectively, and Yule et al. (this volume) interpret an age of ca. 152–156 Ma from complex U/Pb zircon age data for the younger of two phases of the Ashland pluton. Frost et al. (this volume) suggest that the Rogue–Chetco arc (Fig. 1) was the source of the ca. 153-Ma volcanic detritus in Galice sandstones and slates, but this seems unlikely because: (1) the arc axis, probably represented by the plutonic Chetco complex lies west of the Galice Formation, inconsistent with paleocurrent data; (2) U/Pb zircon ages for the Chetco Complex are too old, ranging from 160 to 157 Ma (Yule, 1996; Yule et al., this volume); (3) the Rogue Formation underlies the Galice Formation and is thus too old, although one 40Ar/39Ar hornblende age is 153.4 Ma (Hacker et al., 1995); and (4) some of the components in the sandstones (e.g., glaucophane) are not known from the basement rocks of the Rogue–Chetco arc complex, which are dominantly ophiolitic (Yule, 1996; Yule et al., this volume). Volcanic members in the Galice Formation, which include coarse breccias, were obviously derived from nearby volcanic sources, but volcanic detritus in the Galice sandstones and slates had to have been mixed with non-arc sources before deposition.

Derivation of the ca. 227-Ma detrital zircons within the Galice Formation from the Klamath Mountain terranes to the east does appear unlikely, because there is no known source in the Klamath Mountains. The only known igneous rocks of the appropriate age in the Klamath Mountains are dacites in the Pit Formation of the eastern Klamath Mountains (Albers and Robertson, 1961), but these apparently are barren of zircon (J.E. Wright, written commun., 2004). Hacker et al. (1993) infer that arc rocks of this age are present in the Sawyers Bar terrane in the central Klamath Mountains, but even if they are, these rocks are predominantly mafic and unlikely to have yielded much zircon. The ca. 227-Ma detrital zircon, however, might be present in Upper Triassic to Middle Jurassic clastic rocks of the Klamath Mountains. A pre-Cretaceous reconstruction by Wyld and Wright (2001) places the Pine Nut terrane and the basinal Luning assemblage just east of the Klamath Mountains. Both the Pine Nut terrane and the Luning assemblage contain strata with peak detrital zircon ages of ca. 225 Ma (Manuszak et al., 2000), in addition to rocks of ca. 230 Ma in the former (Dilles and Wright, 1988); thus, these terranes are potential sources of the ca. 227-Ma detrital zircon in Galice sandstones.

South-directed paleocurrent directions in the type Galice Formation (Fig. 6B) suggest that the main source area for the Galice Formation may have been north of the present Klamath Mountains. Klamath rocks are covered by Cenozoic volcanic rocks to the north, but chert, including Triassic chert, (e.g., Pessagno and Whalen, 1982), and supra-subduction zone ophiolite peridotite (Bishop, 1995) are present in northeastern Oregon. Significant volumes of ca. 153-Ma volcanic rocks do not appear to be present (e.g., Saleeby and Busby-Spera, 1992), although some andesitic dikes and the Sunrise Butte pluton have K-Ar ages in the range of 150–158 Ma (Johnson et al., 1995). Only a few probable Late Jurassic arc rocks are exposed farther north in Washington (Miller et al., 1993). Possible source rocks for the ca. 227-Ma detrital zircons in the Galice may be present in the Blue Mountains, Oregon, Washington, and Idaho (e.g., 262- to 219-Ma Sparta complex and Carnian to Norian volcanic rocks of the Olds Ferry terrane; Avé Lallemand, 1995). The monocrystalline quartz-poor Galice sandstones are unlikely to have had a source to the south, because Paleozoic quartz sandstones in the central and southern Sierra Nevada contain zircons much older (>2.2 Ga) than detrital zircons in the Galice Formation (Harding et al., 2000).

**Regional Variations**

Miller et al. (2003) found that detrital-zircon age populations in a sandstone overlying the Ingalls ophiolitic complex in...
the Cascade Mountains of Washington State were similar to ages from a sandstone of the Galice Formation, with peaks at ca. 153 and ca. 227 Ma. The Ingalls sandstone, however, contains no Precambrian zircon. No modal petrographic data are available for the Ingalls sandstones, although the types of clasts are similar to those in the Galice Formation, including Cr-spinel (Southwick, 1974; Miller et al., 1993).

Terranes similar to those of the western Klamath terrane are present in the Sierra Nevada region (Davis, 1969; Irwin, 2003). Conglomerates from the Mariposa Formation of the central Sierra Nevada foothills, which has long been correlated to the Galice Formation (e.g., Diller, 1907; Davis, 1969), are distinct from Galice conglomerates in that they have a much higher proportion of quartzose clasts (Seiders and Blome, 1988; Seiders, 1991, written commun., 1992; Fig. 8). This observation suggests a Sierran rather than Klamath Mountains source for the Mariposa Formation (Seiders and Blome, 1988), even though paleocurrent data reported by Bogen (1984) indicate south-southeast transport. Most of the conglomerates in the southernmost Galice Formation (Wyld, 1985; Fig. 1) are similar to those from the Galice Formation farther north, in which pebbles consist almost entirely of chert, argillite, or volcanic rocks, but some have elevated contents of quartzose clasts and are thus compositionally transitional to conglomerates of the Mariposa Formation (Fig. 8).

Temporal Variations and Tectonic Implications

The ca. 162- to 153-Ma sedimentary rocks of the hemipelagic sequence, the hemipelagic-turbidite transition, and those within the Josephine ophiolite have similarities to the turbidite (Figs. 11, 12, and 13), except for the presence of a metalliferous (hydrothermal) component in some samples (Fig. 4). There is a striking similarity in petrography, and especially detrital Cr-spinel compositions (Figs. 9 and 10), between the thin sandstone bed within pillow lavas of the Josephine ophiolite (ca. 162 Ma) and sandstones of the lower Galice turbidite (ca. 153 Ma). This observation and similarities in the geochemistry of siliceous argillites and Galice slates suggest that the source area for the Galice turbidite was already established by ca. 162 Ma. This implies that the Galice source area was probably produced during ca. 165-Ma thrusting (“Siskiyou orogeny”) that preceded formation of the Josephine ophiolite (Wright and Fahan, 1988; Hacker et al., 1995; Fig. 14). The inferred basin and source area relationships just after this episode of shortening are shown in Figure 14A.

Tectonic Setting of Deposition

The tectonic setting at ca. 160 Ma for the western Klamath Mountains is well constrained (e.g., Saleeby et al., 1982; Harper and Wright, 1984; Wyld and Wright, 1988; Hacker et al., 1995; Yule, 1996; Yule et al., this volume), and a model in which the Josephine ophiolite is located within a back-arc basin behind the west-facing Rogue–Chetco arc at ca. 160 Ma (Fig. 14A) is generally accepted (e.g., Dickinson et al., 1996). The ca. 177- to 168-Ma western Hayfork arc would represent a remnant arc at this time, and a tectonic highland is inferred because of a ca. 165-Ma shortening event (Wright and Fahan, 1988), sometimes called the “Siskiyou orogeny.” Seafloor spreading in the back-arc basin is inferred to have been transform-dominated (Fig. 14A), based on the east-west orientation (in modern coordinates) of spreading centers inferred from sheeted dike orientations and other structural data (Harper et al., 1985; Alexander and Harper, 1992). Parts of both sides of this basin were underlain by older rifted Klamath crust, in places cut by Josephine-age mafic complexes; the Rogue–Chetco arc was built on older Klamath crust rifted away during seafloor spreading in the back-arc basin (Saleeby et al., 1982; Harper and Wright, 1984; Wyld and Wright, 1988; Yule, 1996; Yule et al., 1996).
west (Fig. 14A). This is probably the explanation for the high terrigenous content of sediments within the Josephine pillow lava unit and hemipelagic sequence, with the terrigenous sediment derived from the nearby “Siskiyou” highlands.

Based on the closeness in age of the Galice turbidite with the oldest plutons that cut the roof (Orleans) thrust, which may have >100-km displacement, Harper et al. (1994) reinterpreted the Galice turbidite as a synorogenic deposit (Fig. 14B). In this scenario, the turbidite was derived from Late Jurassic uplift related to underthrusting of the Josephine ophiolite. The area of uplift is inferred to have been similar to that affected by the pre-Josephine (“Siskiyou”) deformation, which would account for the very similar clasts compositions and detrital modes of the sandstone within the Josephine pillow lavas (ca. 162 Ma) and the basal Galice turbidite (ca. 153 Ma). The great increase in amount of sandstone at the top of the hemipelagic-turbidite transition is based on the peak ca. 153-Ma age of detrital zircon in a sandstone from the basal turbidite (Miller et al., 2003) as well as other geochronologic constraints (Harper et al., 1994). Widespread ca. 150-Ma $^{40}$Ar/$^{39}$Ar hornblende and mica cooling ages in the central Klamath Mountains (Hacker et al., 1995) may record Nevadan underthrusting and uplift.

The presence of the apparent hiatus at the top of the hemipelagic sequence (Pessagno and Blome, 1990; Pessagno et al., 2000), which is underlain by a zone of pre-cleavage deformation (Harper, this volume; Fig. 2), is enigmatic. As discussed above, this contact could be a minor unconformity, a normal fault, or a submarine landslide surface. A possible explanation for this surface is that it is a disconformity resulting from nonsedimentation and/or submarine erosion on a flexural bulge in front of the trench as the ophiolite and overlying Galice were thrust beneath older terranes of the Klamath Mountains during the Nevadan orogeny (Fig. 14). Passage of crust over a flexural bulge in front of a trench can result in an unconformity and normal faulting (e.g., Rowley and Kidd, 1981). Underthrusting of the Josephine ophiolite, which involved at least 40 km of displacement, is broadly analogous to subduction. In addition, relief created by this faulting could have resulted in a submarine landslide that removed strata prior to deposition of the hemipelagic-turbidite unit.

The quartzite-rich conglomerates in the Mariposa Formation in the Sierra Nevada foothills (Fig. 8) may indicate that the Mariposa and associated Late Jurassic arc rocks have not been translated appreciably relative to older rocks in the Sierra Nevada. If this suggestion is correct, then the Mariposa should contain >2.2-Ga detrital zircon known to be present in older rocks of the Sierra Nevada (Harding et al., 2000). Similarly, the Galice sandstones and conglomerates appear to be derived largely from the Klamath Mountains, except that there is no known Klamath source for ca. 227-Ma detrital zircons, and the location that provided ca. 153-Ma zircons is uncertain. Additional dating of zircon in the Klamath Mountains and elsewhere in the Cordillera might better constrain source areas for Galice sedimentary rocks and, in turn, better constrain models for Late Jurassic tectonics and paleogeography of western North America.

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APPENDIX

Figure A-1. Drill core sample from the Turner-Albright mine showing siliceous argillite that is interbedded with lavas of the Josephine ophiolite. The black and green laminations are interpreted as terrigenous and tuffaceous in origin, respectively. The alternating black and green colors are also typical of the hemipelagic sequence overlying the Josephine ophiolite. Note coin for scale.
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