STRUCTURAL GEOLOGY OF THE NORTHERN GALICE FORMATION,
WESTERN KLAMATH MOUNTAINS,
OREGON AND CALIFORNIA

A thesis presented to the Faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of Master of Science
College of Science and Mathematics
Department of Geological Sciences

F. Ross Jones
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ABSTRACT

The Galice Formation is a thick, turbiditic flysch sequence that depositionally overlies the Josephine ophiolite in the western Klamath Mountains of northwestern California and southwestern Oregon. The Josephine ophiolite and Galice have been interpreted as the basement and sedimentary cover of a Late Jurassic back-arc basin that opened proximal to the continental margin. During the Nevadan orogeny (ca. 151-147 Ma), the ophiolite and overlying Galice sediments were thrust eastward beneath the continental margin along the regional Orleans fault. Two distinct Nevadan deformational phases are recognized in the Galice—an initial $D_1$ or "main-phase" deformation and a later brittle deformation dominated by thrusting.

$D_1$ involved development of tight to isoclinal northwest-vergent $F_1$ folds which possess an axial-planar $S_1$ cleavage. $S_1$ ranges from a weak pressure-solution cleavage in the northernmost Galice, to a strong slaty cleavage or schistosity in the southern Galice. $S_1$ is ubiquitously parallel to bedding in the absence of recognizable $F_1$ fold closures. Bedding is commonly transposed into parallelism with cleavage in southern exposures of the Galice.

Syntectonic fibers in pressure shadows of pyrite grains reflect a north-south gradient in Nevadan penetrative strain. The fibers are straight and parallel to the trace of slaty cleavage, suggesting that cleavage formed during a
stage of pure flattening. Percent shortening normal to $S_1$ determined from fibers in the northern Galice (SW Oregon) is approximately 33%. However, fibers from slates in northern California record an average of 70% shortening normal to $S_1$. A large increase in the intensity of $S_1$ towards the south also suggests a gradient in finite strain.

Main-phase Nevadan deformation was followed by an episode of brittle shearing. Nevadan thrusts reveal top-to-northwest sense to displacement, similar to vergence observed from $F_1$ folds. A second generation of folds, $F_2$, appear to be genetically related to the thrusts. Nevadan thrusts clearly overprint main-phase structures ($F_1$, $S_1$). In the northernmost Galice, angular $F_1$ folds are thrusted-out parallel to bedding in one of the fold limbs. Further to the south, $F_1$ folds are more typically truncated by thrusts that are parallel to $S_1$ cleavage planes. Small-scale thrust systems in thinly-bedded rocks have produced peculiar bedding truncations or "pseudocross-bedding". These small thrusts locally truncate the $S_1$ cleavage.

Fundamental conclusions obtained in this study are:
1) The Galice Formation is tightly folded on both outcrop and regional scales.
2) There is a north-to-south increase in Nevadan regional metamorphic grade and penetrative strain. This gradient may reflect a small southward component of underthrusting along the Orleans fault.
3) Nevadan vergence from folds and thrusts in the Galice is dominantly northwestward.

4) Most Nevadan thrusting in the Galice took place after the main-phase penetrative deformation and peak metamorphism. This ductile-to-brittle transition may have resulted from continued underthrusting during regional uplift.
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Chapter 1: INTRODUCTION

1.1 Regional Perspective

The Klamath Mountains province is an excellent example of multiple accretion of diverse oceanic terranes at a long-lived convergent plate margin. The lithotectonic framework of the Klamath Mountains was first described by Irwin (1960) who recognized four, broadly arcuate belts in the Klamaths (fig. 1.1). The belts are bounded by eastward-dipping thrust faults and become successively younger to the west. Irwin (1966) named the belts (from structurally highest to lowest): the eastern Klamath belt, the central metamorphic belt, the western Paleozoic and Triassic belt and the western Jurassic belt. More recently, the Klamaths have been interpreted as an "amalgamation" of several allochthonous oceanic terranes which have accreted to the western margin of North America from the early Paleozoic to the middle Mesozoic (Hamilton, 1969; Irwin, 1981,1985). Terranes of the Klamath Mountains largely consist of ultramafic and ophiolitic thrust sheets, oceanic sedimentary and island arc sequences, diverse melanges and calc-alkaline plutons, primarily of Mesozoic age (Hotz, 1971a). The Klamaths are unconformably overlain by lower Cretaceous sedimentary rocks correlated with the Great Valley sequence in northern and central California (Irwin, 1981).
Figure 1.1 Map of the Klamath Mountains province showing the configuration of four major thrust-bounded terranes: EK= eastern Klamath belt, CM= central metamorphic belt, TrPz= western Triassic and Paleozoic belt and WJ= western Jurassic belt. Jcm is the Jurassic Condrey Mountain schist. Plutons were omitted for clarity. From Jachens and others, 1986, fig.1 (modified from Irwin, 1981; Hotz, 1971).
1.2 The Galice Formation

The Galice Formation is exposed in the western Jurassic belt of the Klamath Mountains (fig. 1.2). It was first named by Diller (1907) for a locally-exposed sequence of slate, graywacke and rare pebble conglomerate at the "town" of Galice and along Galice Creek, approximately 25 kilometers northwest of Grants Pass, Oregon. Volcaniclastic units which are interstratified with shales and graywackes in exposures along the Rogue River were subsequently included in the "type" Galice Formation by Wells and Walker (1953). A geological map of the Galice Quadrangle (Wells and Walker, 1953) shows a conformable contact between the Galice Formation and underlying volcanics; however, some authors have interpreted this lithologic boundary as a fault (Park-Jones, 1988). This contact may have been depositional, but is evidently obscured by faulting. Wells and others (1949) estimated the stratigraphic thickness of the Galice to be 8 to 9 kilometers, but this figure is almost certainly too high, considering that it is isoclinaly folded.

Specimens of the bivalve *Buchia concentrica* (Sowerby) were collected from several localities in the type Galice (Diller, 1907; Diller and Kay, 1924). This Late Jurassic index fossil has a known age-range of late Middle Oxfordian to Middle Kimmeridgian, or approximately 159-155 Ma (Imlay,
Figure 1.2 Geological map of the western Jurassic belt (modified from Harding, 1986). The area in the smaller, five-sided box is the northern field area of this study, shown in more detail in figure 1.6.
1980; Harland et al., 1982). A second younger age-constraint is provided by outliers of Cretaceous shallow marine and subaerial sedimentary rocks which overlie the Galice in angular unconformity. Cretaceous outliers overlie the Galice near O''Brien, Oregon, in Douglas County, Oregon, and along Grave and Days Creeks north of Grants Pass (see map by Hotz, 1971b). The Riddle Formation (Diller and Kay, 1924) unconformably overlies the Rogue Formation north of Grants Pass, and has been correlated with the Middle to Late Tithonian (ca. 148-144 Ma) Knoxville Formation in California (Imlay et al., 1959). Massive conglomerates and coarse-grained sandstones unconformably overlie the Galice at Indian Hill near O'’Brien, Oregon (fig. 1.3). These rocks are believed to be the northern equivalents of the lower Cretaceous Horsetown Formation, a member of the Great Valley sequence in California (Wells et al., 1949). Fossils from this outlier are Late Hauterivian to Barremian, or about 126-119 Ma (Imlay et al., 1959). Interestingly, the stratigraphy of this sequence strongly resembles that of the older Knoxville Formation described by Diller and Kay (1924, p.3) as "gray sandstone and some shale with massive conglomerate at the base".

Graywacke and slate are exposed in a north-south trending belt which extends along the entire length of the western Klamath Mountains (Hershey, 1911; Irwin, 1960). Cater and Wells (1953) first correlated these rocks with the type Galice exposed along the Rogue River and Grave Creek.
Figure 1.3 Geological map of the area near O'Brien, Oregon (after Ramp, 1979). Tra= Applegate Group, Jjp= Josephine peridotite, Jg= Galice Formation, Kh= Horsetown Formation and Qs= Quaternary sediments.
This correlation was based on the occurrence of *Buchia concentrica* at localities in northernmost California, and on lithostratigraphic similarities. Similarities in sedimentary petrography further support the conclusion that the two formations are correlative (Harper, 1980a, 1983).

The type Galice Formation has been interpreted as conformably overlying volcanics of the Rogue Formation (Wells and Walker, 1953; Garcia, 1979; Harper and Wright, 1984). The contact is exposed along the Rogue River at Almeda Mine, and along Galice and Grave Creeks, but, as noted above, it may be faulted.

In contrast, a clear depositional contact between the Galice and the Josephine ophiolite is observable at several locations in northwestern California (Vail, 1977; Harper, 1983). Figure 1.4 shows two schematic stratigraphic columns for comparison. The Galice is exposed in two different thrust sheets (Harper and Wright, 1984); the "Galice" overlying the Josephine ophiolite is structurally higher than the type Galice Formation.

1.3 Lithology and Stratigraphy

The northernmost Galice Formation mainly consists of interbedded black shale and lithic wacke (fig. 1.5) which become strongly foliated slates and metagraywackes to the south. Thick sections of shale (0.5 km or more) with very
Figure 1.4 Simplified comparative stratigraphic columns showing the lower portion of the Galice Formation exposed at the type-locality along the Rogue River and Galice Creek, Oregon, and overlying the Josephine ophiolite along the Middle Fork Smith River in northern California.
Figure 1.5 Slightly dismembered beds of alternating shale and graywacke in the Galice Formation along the Illinois River near Cave Junction, Oregon. Beds are nearly vertical and strike approximately north-south.
few graywacke beds are found in some places. Similarly, there are intervals of massive graywacke (1-3 meter thick beds) containing no shale. Thin siltstone beds are commonly found interlaminated with shale. Siltstone is also a component of some graded beds, and is gradational between coarse-grained graywacke and shale. Volcanic rocks were not found in the Galice during field work for this thesis; however, volcaniclastic intervals were mapped by Snoke (1977) near the Preston Peak thrust in northern California.

Diller (1907) noted the occurrence of pebble conglomerate along Galice Creek at the type-locality. This rare conglomerate unit (units?) contains .5 to 4 cm subrounded clasts of chert, shale, volcanics, sandstone and quartzite in a fine sandy matrix (Diller, 1907; Wells et al., 1949). Massive conglomerate beds are more frequently exposed in the lower portion of the Galice, near the depositional contact with the Josephine ophiolite in northern California. Brown and black chert grains, up to 6 mm in diameter, are the dominant constituents of graywacke at the base of most graded beds. In addition to chert, other important components of wackes are detrital quartz, micas, feldspar, chlorite, pyrite, and lithic clasts such as tuffs, quartzite, argillite and serpentinite. Graywackes normally weather to a light tan or brown color.

Shales are gray to black on fresh surfaces and appear brown or reddish-brown in weathered exposures. Galice shales contain abundant opaque carbonaceous material and
pyrite. Most pyrite occurs in aggregates of microcrystalline spherules called "framboids" which will be discussed further in chapter 5. Sinuous dark trails on bedding surfaces are probably the trace fossil Helminthoida (R. Park-Jones, pers. comm.). Pyritized fossil trails were useful in finite strain analysis described in chapter 5.

Shales in the northern Galice near Cave Junction, Oregon are best termed "slaty shales"; however, pelitic rocks become true slates, and locally phyllites, to the south in northern California. Cleavage is usually subparallel to bedding, giving shales a strong bedding-parallel fissility. Foliation in graywackes is weak or absent in the northern Galice, whereas, to the south, wackes have a strong foliation and have been termed "semischists" (Harper, 1980a).

The absence of distinct marker beds or lithologies, combined with structural complexity and limited exposure, makes present stratigraphic correlation within the Galice impossible. Detailed geochronologic work using pollen from Galice shales may prove useful in future studies (G.D. Harper, pers. comm.).

A 21 to 31 meter-thick hemipelagic sequence consisting of metalliferous red and black chert interlayered with argillite is frequently found near the contact of the Galice with pillow lavas of the underlying Josephine ophiolite (Pinto-Auso and Harper, 1985). About 100 meters of massive graywacke and pebble conglomerate commonly overlie the
hemipelagic sequence where continuous, unfaulted sections are observed. Repeated massive graywacke intervals in the Galice might be fault-transported equivalents of this coarse basal unit. Thick sequences of black shale and some volcaniclastic intervals have been interpreted as occurring in the structurally highest strata of the type Galice Formation (Wells et al., 1949).

1.4 Tectonic Setting

Reconnaissance mapping by Irwin (1960) has shown that the Galice is the structurally highest unit of the Klamath province and occurs in a distinct belt which he termed the "western Jurassic belt" (WJB). The WJB is bounded to the east by the east-dipping Orleans fault (Hershey, 1911; Jachens et al., 1986), also named the "Preston Peak thrust" by Snoke (1977). At its western margin, the WJB is thrust over Cretaceous and Tertiary rocks of the Franciscan assemblage along the northern extension of the Coast Range thrust. The WJB contains two, east-dipping thrust sheets or "subterranes" (fig. 1.2): (1) A higher thrust sheet consisting of the Josephine Peridotite (Dick, 1976), the Josephine ophiolite (Harper, 1980b) and "Galice" Formation, and (2) A lower thrust sheet, exposed exclusively in southwestern Oregon, containing the Chetco plutonic complex (Hotz, 1971a), mafic gneisses of the Briggs Creek
amphibolite, the Rogue Formation and the type Galice Formation.

The Josephine ophiolite and overlying Galice flysch have been interpreted as the basement and sedimentary cover of a Late Jurassic back-arc basin (Harper, 1980b, Harper and Wright, 1984). The following evidence supports this conclusion:

1) **Sedimentary Petrography.** Graywackes in the Galice show a mixed continental margin/volcanic arc source (Harper, 1980a). Volcanic rock fragments and isolated phenocrysts in coarse wackes and conglomerates suggest that the depositional basin was near an island arc, active in the Late Jurassic. Moreover, the metasedimentary rocks of the type Galice contain intercalated volcanic layers which strongly resemble volcanics of the Rogue Formation (Wells and Walker, 1953; Garcia, 1979). Diverse rock and mineral fragments in the Galice such as detrital glaucophane, chert, chromian spinel, quartzite and mica schist suggest that the Galice was also derived from erosion of sedimentary, metamorphic and ultramafic rocks of older Klamath terranes to the east.

Recent mapping and structural data from the type-Galice strongly suggest that interlayered volcanic units, averaging 15 meters in thickness, are thrust slivers of Rogue volcanics (R. Park-Jones, pers. comm.). Repetition of the volcanic units by tight folding is also possible, based on
large flattening strains evident from Galice slates and
tuffaceous rocks at the type-locality.

2) **Regional Relationships.** Since the Galice overlies both
the Josephine ophiolite and a coeval Late Jurassic island
arc (Garcia, 1979), there is presumed to be a close spatial
and temporal arc-ophiolite association (Harper and Wright,
1984). The predominantly mafic Chetco intrusive complex has
been interpreted as the plutonic core of the Late Jurassic
island arc (Dick, 1976; Garcia, 1982).

3) **Geochronology.** The age of the Josephine ophiolite is 162
+/-1 Ma, determined from U/Pb zircon dating of ophiolitic
plagiogranites (Saleeby et al., 1986). Previous ages of 157
+/-2 Ma for the Josephine ophiolite (Saleeby et al., 1982)
are slightly discordant. Zircons from a silicic tuff from
the Rogue Formation have yielded a U/Pb age of 157+/-2 Ma
(Saleeby, 1984).

This evidence suggests that the Galice was deposited on
the Josephine ophiolite in a Late Jurassic back-arc basin
between a coeval island arc and the continental margin
(Harper and Wright, 1984) represented by the western
further suggested that opening of the Josephine basin
resulted from the splitting of a Middle Jurassic arc.
Recent geochronologic data from the Devils Elbow ophiolite,
the southern equivalent of the Josephine ophiolite, suggest
that rifting began at about 164 Ma (Wyld, 1985; Wyld and
Wright, 1988), 5 m.y. before cessation of the Middle
Jurassic arc (ca. 159 Ma). Magmatism in the Middle Jurassic arc is constrained to the interval 177-159 Ma (Harper and Wright, 1984). The extrusive rocks of the Middle Jurassic arc are probably the volcanics of the western Hayfork terrane in the WPTB. This arc-splitting event produced an active arc/ back-arc basin/ extinct(?) remnant arc geometry proximal to the continental margin in the Late Jurassic (Harper and Wright, 1984).

The Galice contains partial and complete Bouma sedimentation cycles which are indicative of deposition by turbidity currents at medium to abyssal depths. It lithologically resembles Taconian flysch deposits such as the Martinsburg and Austin Glen formations in the central and northern Appalachian orogen. Taconian flysch was deposited in submarine fan systems in deep water during the Middle to Late Ordovician (Rowley and Kidd, 1981; Lash and Drake, 1984).

The Galice turbidites were apparently deposited in greater than 2,000 meters of water. Basal pelagic sediments are siliceous, rather than calcareous, suggesting deposition below the carbonate compensation depth (CCD). The CCD was approximately 2,500 meters below sea level in the Late Jurassic (Bosellini and Winterer, 1975). However, there may not have been a source of calcareous pelagic sediments at this time (W.S.F. Kidd, pers. comm.). Furthermore, calcareous lenses are present locally within the basal pelagic sequence of the Galice.
Nonetheless, depths of modern back-arc basins are generally greater than 2,200 meters (Harper, 1982, and references therein). Pelagic sediments in the basal section of the Galice Formation probably accumulated on a bathymetric "high", coincident with the spreading axis (Pinto-auso and Harper, 1985). The depth of the axial high in modern back-arc basins is normally in excess of 2,000 meters (Harper, 1982).

In addition to rhythmic Bouma sequences, sedimentary structures such as shale rip-up clasts, flute casts and tool marks are evidence for deposition by gravity-driven turbidity currents. Sediment transport was to the west (Harper, 1980a), oceanward from the Late Jurassic continental margin, implying erosion of older rocks of the Klamath Mountains.

Cessation and collapse of the Late Jurassic arc was followed by imbrication of the present western Jurassic belt into two east-dipping thrust sheets. The Josephine ophiolite and overlying Galice were simultaneously underthrust beneath the continental margin and thrust over the Late Jurassic island arc (Rogue-Chetco terrane) during the Nevadan orogeny (Harper and Wright, 1984).

1.5 Study Areas

Field work in the Galice overlying the Josephine ophiolite focused on two areas along approximately 40
kilometers of its exposure. The northern study area is located in Josephine County, Oregon, near the towns of Cave Junction and Kerby (fig. 1.6). It covers much of the NE and NW quadrants of the USGS 15' Cave Junction quadrangle (1953). Figure 1.6 shows the geography of this area for reference to localities cited in this thesis. It also shows the major fault boundaries between the Galice and the Applegate Group (Wells et al., 1949) and Josephine Peridotote (Dick, 1976). The Illinois Valley has good road access, the major arterial being the Redwood Highway (US 199) which extends from Grants Pass, Oregon to the Pacific coast at Crescent City, California. The combined population of the northern Illinois Valley (Cave Junction and vicinity) is about 2,000.

A majority of the exposures of the Galice Formation are found along recent, yet presently inactive, logging road-cuts in the lower hills of the Siskiyou Range east of Kerby. The Siskiyou Range is underlain by the more resistent lithologies of the Applegate Group, and attain a maximum elevation in Oregon of 1,800 meters at Kerby Butte. The Galice is best exposed in steep logging road-cuts where stone has been quarried for road paving material. Semi-continuous fresh exposures of the Galice are interspersed with river gravels along the east and west forks of the Illinois River system near Cave Junction (fig. 1.7). Outcrops of the Galice in this area are, nevertheless, entirely restricted to road-cuts and some stream beds.
Figure 1.6 Reference map for the northern study area near Kerby and Cave Junction, Oregon. Map shows the exposed width of the Galice, as well as the major fault boundaries with the Josephine peridotite and the Applegate Group. "JO" refers to pillow lava and breccia of the upper portion of the Josephine ophiolite.
Figure 1.7  Nearly vertical graywacke beds exposed along the west fork of the Illinois River near Cave Junction, Oregon. View is to the south-southwest.

Figure 1.8  Well-aligned load casts along the base of an overturned graywacke bed. Road-cut along West Road near Pomeroy Dam (see fig. 1.6).
Smaller stream exposures are covered with alluvium from landslides and debris from recent "clear-cutting" logging operations.

The structural geology of the northern Galice in southwestern Oregon (fig. 1.6) has not been studied in much detail by previous workers. Geological maps including this area have been compiled by Wells and others (1949), Ramp (1979, 1986) and Page and others (1981). A few comments concerning the structure of the Galice were made by Diller and Kay (1924).

A distinct advantage of studying the northern Galice is the preservation of sedimentary structures which indicate the direction of stratigraphic younging. Among the most common younging indicators are graded beds, flute and groove casts and peculiar well-aligned load casts (fig. 1.8). Original sedimentary textures and structures have not been obscured by metamorphic overprinting or intense foliation development. The widespread occurrence of younging indicators in graywacke beds facilitates recognition of large-scale overturned folds.

A second area where structural data was collected is in Del Norte County, California, near the Oregon border (fig. 1.9). Continuous, across-strike exposures occur along stream-cuts in steep valleys of the Smith River drainage system. Traverses across the Galice were made along Oregon Mountain Road, the Siskiyou and Middle forks of the Smith River and Knopti Creek (fig. 1.9).
Figure 1.9 Reference map for study locations in northern California. A-A' is the location of the vertical cross-section discussed in chapter 4.
Field work leading to this thesis was conducted for six weeks during June and July, 1986. A structural map (Plate 1) was prepared from detailed observations and data collected along logging roads in the area immediately east of Kerby, Oregon. Structural data for the map was recorded on 1985 aerial photographs from the Bureau of Land Management (USDI). Enlarged quadrants of the USGS 15' Cave Junction topographic quadrangle provided the base map. Field data from traverses in the northern California field area were recorded on the USGS 7.5' Hurdygurdy Butte, Devils Punchbowl, and Broken Rib Mountain topographic quadrangle maps.

1.6 Objectives

This project was originally undertaken to approach several structural problems in the Galice Formation noted by previous workers. These are discussed below:

1. Whether or not the Galice was deformed in a partially lithified state.--- There is evidence that deep water sediments are strongly deformed in an unconsolidated state at many convergent margins (Cowan, 1982; Westbrook et al., 1982; Byrne, 1984). Chapter 3 is a discussion of early structures in the Galice and their possible significance.

2. An apparent scarcity of $F_1$ folds, and the ubiquitous parallelism between bedding and the $S_1$ cleavage.--- These problems are emphasized by Harper (1980a) and Wyld (1985).
An important objective of this study is to determine whether the strata constitute limbs of large, overturned folds, or whether they are largely upright and only rarely folded. This problem has not been fully resolved in previous structural studies of the Galice Formation.

3. The direction of Nevadan thrusting and vergence.-- Studies of Nevadan vergence from folds and tectonic lineations in the Galice and along the basal thrust of the Josephine ophiolite have produced conflicting interpretations (Harper and Wright, 1984; Roure, 1984; Cannat and Boudier, 1985; Gray, 1985).

4. The extent and relative timing of deformations which post-date the main-phase Nevadan deformation (D₁).-- Are F₂ folds a continuation of the main-phase Nevadan episode, or are they "post-Nevadan" structures? A solution depends on how "Nevadan" structures are defined. Some aspects of what is termed "late Nevadan" deformation are examined in chapter 6.

A more general question to be addressed is, "Do structures in the Galice Formation reflect the kinematics of underthrusting and emplacement of the Josephine/Galice thrust sheet during the Nevadan orogeny?" In rocks deformed by regional thrusting or shearing, it is important to determine whether penetrative fabrics record coaxial or non-coaxial strain histories. In chapter 5, syntectonic fibers in pressure shadows of pyrite framboids are examined and discussed in light of the above questions. Fibers have
proven to be a useful tool for structural geologists in the
determination of strains accumulated during tectonic
transport and emplacement (Durney and Ramsay, 1973; Beutner
et al., 1983; Sample and Fisher, 1986). Fiber growth in
Galice slate clearly reflects a north-south variation in
finite strain during Nevadan cleavage development.
Chapter 2: THE NEVADAN OROGENY

2.1 Regional Extent

The Nevadan orogeny was first defined as plutonism, metamorphism and deformation affecting Cordilleran rocks from Alaska to Mexico (Blackwelder, 1914). Blackwelder (1914) assigned a Jurassic age to the Nevadan. Subsequently, three phases of Middle and Late Jurassic arc-type plutonism have been interpreted as Nevadan events (Lanphere et al., 1968). The Nevadan was more recently restricted to Late Jurassic deformation and metamorphism in the Sierra Nevada foothills (Bateman and Clark, 1974); a Late Jurassic age for the Nevadan in the Klamath Mountains is consistent with recent age data from syn- and post-Nevadan plutons (Harper et al., 1986). The Nevadan is best documented in the western Klamath Mountains (Harper and Wright, 1984) and in the foothills of the Sierra Nevada range (Schweickert et al., 1984; Tobisch et al., 1987).

2.2 The Western Klamath Mountains

The Nevadan orogeny mainly affected rocks in the western Jurassic belt of the Klamath province. However, further to the east, the Rattlesnake Creek Terrane (RCT) in the upper plate of the Orleans thrust probably experienced
Nevadan deformation (Snoke, 1977). Gray (1985) suggested that melanges in the RCT were "remixed" during the Nevadan. Nevertheless, Nevadan-age structures may be indistinguishable from the deformational and metamorphic effects of a recently documented Middle Jurassic (ca. 170 Ma) orogenic event (Snoke, 1977; Fahan and Wright, 1983; Gorman, 1984).

The regional-scale effect of the Nevadan in the Klamath Mountains, shown in figure 2.1 (from Wyld and Wright, 1988), was westward thrusting of the Late Jurassic Josephine ophiolite and Galice Formation over the coeval volcanoplutonic Rogue-Chetco island arc subterrane (Harper and Wright, 1984; Wyld and Wright, 1988). Thrusting of the Josephine ophiolite and peridotite over the Late Jurassic arc occurred along the Madstone thrust (Dick, 1976). To the east, the ophiolite and overlying Galice flysch were underthrust at a low angle beneath the western Paleozoic and Triassic belt (WPTB) along the Orleans fault (Klein, 1977; Jachens et al., 1986). These regional thrusts are shown in figure 1.2. Intense deformation in the Galice, and locally in the upper portion of the Josephine ophiolite, accompanied Nevadan thrusting. Large flattening strains (chapter 5) are evident from flattened pillow lavas in the southern Josephine ophiolite and foliated pebble conglomerates in the Galice. Small-scale thrusting, folding and penetrative cleavage development are prevalent in the Galice Formation.
Figure 2.1 Illustration of Middle to Late Jurassic tectonism in the western Klamath Mountains.  A. Middle Jurassic arc (western Hayfork terrane, WHT) is built on continental margin (Rattlesnake Creek terrane, RCT).  B. Formation of the Josephine ophiolite and Late Jurassic arc following rifting of the older arc and its basement.  C. Following cessation of the Late Jurassic arc, imbrication of the arc, ophiolite and remnant arc occurred during the Nevadan orogeny.
overlying the Josephine ophiolite, as well as the type-Galice and underlying Rogue Formation.

All lithologies in the western Jurassic belt experienced regional prehnite-pumpellyite to lower greenschist facies metamorphism during the Nevadan Orogeny (Harper, 1980a). Metamorphic grade increases from north to south. The transition between the prehnite-pumpellyite and lower greenschist facies occurs about 25 kilometers south of the Oregon border, roughly at the latitude of Crescent City, California (Harper and Wright, 1984).

The age of the Nevadan is well-constrained as Late Jurassic. An upper age bracket is the Oxfordian-Kimmeridgian Galice Formation, the youngest formation involved in the orogeny. Middle Jurassic plutons intruding the western Paleozoic and Triassic belt are older than the cross-cutting Orleans thrust (Gray, 1985; Jachens et al., 1986). Small calc-alkaline plutons intruding the Galice and underlying ophiolite have yielded Ar$^{40}$/Ar$^{39}$ dates of 148 +/- 1 Ma and 150 +/- 1 Ma, respectively, and are interpreted as syn-emplacement plutons (Harper et al., 1986). 150 million-year-old dikes and sills (Saleeby et al., 1982) intruding the Galice (fig. 2.2) are typically deformed and metamorphosed along with the host rock. Two plutons intruding the Orleans thrust provided Pb/U ages of 150 +/- 2 Ma and 146 +/- 4 Ma (Harper et al., 1986). Plutons in the western Klamaths that are younger than 147 Ma are largely unaffected by Nevadan deformation and metamorphism.
Figure 2.2 Three meter wide felsic sill intruding vertical slates of the Galice Formation. Exposure along Middle Fork Smith River at intersection with Knopti Creek (fig. 1.9). Sills in the Galice are folded with their sedimentary host rocks.
Other important Nevadan age-constraints are the previously discussed outliers of early Cretaceous shallow marine sedimentary rocks which unconformably overlie the Galice and older terranes in the Klamaths. These rocks have been correlated with various formations of the Great Valley sequence in California, and may be as old as latest Tithonian, or about 145 Ma (Imlay et al., 1959). The above age relationships suggest that Nevadan thrust emplacement and peak metamorphism occurred in the Klamaths between 151 Ma and 147 Ma. Whole-rock metamorphic K/Ar ages of 148 +/- 2, 151 +/- 4, and 153 +/- 4 Ma from metagraywackes in the southern Galice are consistent with these age limits (Lanphere et al., 1978).

2.3 Nevadan Orogenesis in the Sierra Nevada

Much parallelism exists between Nevadan orogenesis in the western Klamaths and the western foothills of the Sierra Nevada range in northern and central California. The Late Jurassic Mariposa and Logtown Ridge formations, which may be southern extensions of the Galice and Rogue, were strongly deformed and metamorphosed in the Late Jurassic. The Smartville ophiolite, with published magmatic ages of 161-155 Ma (Saleeby and Moores, 1979; Saleeby, 1981), was emplaced during the Nevadan orogeny. A northwest structural trend in the northern Sierra Nevada is parallel to the structural trend in the southern Klamaths; however, the two
provinces are offset along a roughly east-west disjunction, the "Sierra-Klamath tear". An apparent difference between the provinces is that Nevadan deformation and low-grade metamorphism affected rocks as old as middle Paleozoic in the Sierra Nevada foothills (Varga, 1985).

Nevadan structural development was more complex in the foothills of the Sierra Nevada than in the western Klamaths. West-vergent high-angle reverse faults and upright folds with a steep slaty cleavage are interpreted as the dominant Nevadan structures (Bogen et al., 1985). However, Day and others (1985) document earlier east-directed thrust faults, and interpret them as Nevadan. The direction of Nevadan thrusting and emplacement is less clear in the Sierra than in the Klamath province.

Another important difference between the Klamaths and Sierra Nevada is the age of the Nevadan orogeny. The Nevadan may be older in the Sierra Nevada (Schweickert et al., 1984; Bogen, 1984). Stratigraphic ages and ages of cross-cutting plutons constrain the Nevadan as 158 Ma to 152 Ma in the northern Sierra Nevada (Schweickert et al., 1984).

2.4 Tectonic Models for the Nevadan Orogeny

The Nevadan orogeny has been suggested to have involved the collision of one or more island arcs with the Late Jurassic continental margin (Schweickert and Cowan, 1975; Schweickert, 1981; Schweickert et al., 1984; Day et al.,
Schweickert and Cowan (1975) envisioned the Nevadan as a collision of an arc/ back-arc basin/ remnant arc system with an early Mesozoic arc built on the continental margin. Their model requires the closure of an ocean basin at oppositely-dipping subduction zones, similar to modern arc convergence in the Molluca Sea, south of the Phillipine Islands (Hamilton, 1979; Silver and Moore, 1979). This model was recently simplified to involve collision of a single arc with the basement of an older continental arc (Schweickert et al., 1984). Collision during the Nevadan is proposed to be the cause of extensive shortening in pre-existing rocks along or near the Late Jurassic continental margin.

The Nevadan orogeny in the Klamath province was a short-lived event involving collapse of a Late Jurassic island arc, closure of a back-arc basin, west-directed thrust imbrication and accretion of the Late Jurassic arc and ophiolite to the continental margin (Harper and Wright, 1984). Harper and Wright (1984) proposed that this entire process took place proximal to the continental margin, within a single west-facing arc system.

It has been suggested that the Late Jurassic ophiolites in the Sierra and western Klamaths were generated by oblique spreading, roughly parallel to a truncated continental margin (Harper et al., 1985). This model is consistent with the east-west orientation of sheeted dikes in the Josephine ophiolite (Harper, 1980a), and evidence for major north-
south paleo-fracture zones beneath the Great Valley and in the Sierra and western Klamaths (Harper et al., 1985). The Late Jurassic paleotectonic setting of the Sierra-Klamath orogen is illustrated in figure 2.3. Nevadan convergence at a high angle to the continental margin may have resulted from an abrupt change in relative plate motion between the Farallon and North American plates. Large magnitudes of crustal shortening in the Klamaths and Sierra Nevada orthogonal to the continental margin favor high-angle convergence rather than a strike-slip dominated transpressive regime, as suggested by Saleeby (1981).

The expression of the Nevadan orogeny apparently differs significantly in the Klamaths and Sierra Nevada. Ingersoll and Schweickert (1986) note several important discrepancies:

1. There may be a significant age difference as noted above.
2. The Mariposa Formation, unlike the Galice, is interpreted as synorogenic flysch that was deposited in an active trench (Bogen, 1984).
3. Different litho-tectonic terranes were accreted in the Sierra region during the Nevadan orogeny. For example, the Nevadan apparently involved accretion of two island arc systems with intervening ophiolite and marine basinal sedimentary/volcanoclastic rocks (Schweickert and Cowan, 1985).
Figure 2.3 Late Jurassic paleotectonic setting of northern California showing inferred orientations of spreading centers, subduction zones and transform faults with respect to the continental margin. (after Harper and others, 1985, fig. 5)
4. It is highly uncertain whether Middle to Late Jurassic terranes in the western Sierra Nevada were obducted onto, or thrust beneath, the continental margin.

Strict parallelism between the Late Jurassic tectonic history of the Sierra Nevada and the Klamath Mountains remains controversial. Clear definition of a "Sierra-Klamath link" awaits better geochronologic control on accretionary and deformational events, particularly in the northern Sierra Nevada.

This thesis examines the deformational effects of regional convergence during the Nevadan orogeny. Although the focus is on the structure of a single formation, the deformational succession in the Galice has an important bearing on models for Late Jurassic accretion of island arc and ophiolite terranes in the Klamath Mountain province (Dick, 1976; Harper and Wright, 1984; Gray, 1985; Wyld and Wright, 1988).
Chapter 3: PRE- AND EARLY TECTONIC STRUCTURES

This chapter includes a discussion of those structures in the Galice Formation that appear to pre-date Nevadan folding, cleavage development and thrusting. These structures formed while the sediments were unlithified or incompletely lithified. Various lines of evidence in support of this conclusion are presented throughout this chapter. Although most of the structures are considered "pre-tectonic", the possibility of an early, pre-$D_1$ phase of tectonic deformation is also considered.

3.1 Penecontemporaneous Deformation

In the Galice Formation, as in most other tectonically deformed sedimentary sequences, it is important to make a distinction between hard-rock and soft-sediment or "penecontemporaneous" structures. Mistakenly interpreting soft-sediment folds as tectonic in origin introduces obvious errors into structural analysis of folding. For this reason, geologists have introduced several criteria, based on field observations, which help in this distinction (Rettger, 1935; Nevin, 1949; Potter and Pettijohn, 1963; Hobbs et al., 1976). From a lengthy list, Hobbs and others (1976) cite three criteria which, when used singly, provide
conclusive evidence for either soft-sediment or hard-rock folding:

(1) Existence of undeformed animal burrows in folded rocks implies soft-sediment folding.

(2) Bending of a metamorphic foliation around fold closures is evidence for hard-rock folding.

(3) A systematic relationship between strain of fossils and their position in the folds (e.g. Geiser, 1974) provides conclusive evidence for hard-rock folding.

Penecontemporaneous folds are defined as those which form in a sedimentary environment at approximately the same time as deposition. Thus, their development is thought to precede lithification. There could, however, be a complete gradation between folds formed in un lithified sediment and those formed in hard rock (Williams et al., 1969; Hobbs et al., 1976). The common conception that soft-sediment folds are non-tectonic may also be questionable (Potter and Pettijohn, 1963).

Penecontemporaneous folds are sometimes equated with "slump folds"—those which presumably form by sliding of a layer down an inclined surface with eventual crumpling at the base. Another type of penecontemporaneous structure is convolute lamination. This chaotic feature results from seismic disturbance of unconsolidated bedding laminae. However similar disruption may be produced as currents flow over the sediments (Potter and Pettijohn, 1963).
Slump features are often restricted to a single bed, or group of beds, within a particular stratigraphic horizon (Williams et al., 1969; Helwig, 1970). Folds may be truncated by undeformed beds parallel to a planar erosional surface (Potter and Pettijohn, 1963). Slump folds are commonly found in highly disturbed packets of strata, separated from undeformed beds by "planes of sliding" (Helwig, 1970). Groups of chaotic folds in which neighboring folds have different orientations and geometries have been cited as evidence for penecontemporaneous deformation (Rettger, 1935).

**Soft-sediment Folds**

Folds which satisfy some of the criteria mentioned above are found in the Galice Formation, particularly in the structurally complex outcrops along the Illinois River (fig. 1.7). Figure 3.1 is interpreted as a slump fold in which thinly-laminated shale is sharply truncated against part of a graywacke bed. The fold is intrafolial, as it is contained between two graywacke beds, and appears to be restricted to a specific bedding horizon.

Figure 3.2A is a hand sample of other intrafolial folds contained within a single bed. The sample's position in the bed, as well as the relative attitude of the regional $S_1$ cleavage, is shown in figure 3.2B. Bedding is highly disrupted, particularly in the core of the synform. Dark
Figure 3.1 Slump fold in shale beds and graywacke along West Fork Illinois River. The shale beds are truncated against the graywacke bed at the arrows. Younging direction from adjacent beds is "upwards" in the photo. Lens cap is 5 cm wide.
Figure 3.2  A. Hand sample of internally disrupted soft-sediment folds of bedding laminae. B. Sketch showing relative positions of samples shown in figures 3.2A, 3.3 and 3.4 in a single bed from an exposure approximately 6 km northeast of Kerby, Oregon. Bedding and cleavage ($S_0$, $S_1$) are indicated. View is to the north.
pelitic laminae are chaotically dispersed and cross-cut more quartzose layers. This cross-cutting relationship is more obvious in a thin-section of the antiform (fig. 3.3). Figure 3.3 is cut normal to the fold axis and axial plane. The foliation in the pelitic zones is parallel to, and morphologically resembles, the $S_1$ cleavage observed at the sample locality. An adjacent sample from the same bed (fig. 3.4) shows no folding, but contains irregular, discontinuous pelitic laminae which appear "stylolitic". The stylolitic spikes are, in fact, $S_1$ cleavage folia. These relationships are illustrated in figure 3.2B. The Nevadan cleavage transects the axial planes of the folds; thus, the folds are interpreted as pre-Nevadan.

Axial-planar cleavages are known to occur with penecontemporaneous fold structures (Williams et al., 1969; Newell et al., 1972; Woodcock, 1976). However, no preferred orientation of mineral grains or cleavage folia were found parallel to the axial planes of soft-sediment folds in the Galice Formation. Furthermore, a regional tectonic cleavage, $S_1$, transects the early folds, similar to the findings of Helwig (1970) from penecontemporaneous folds in Newfoundland. More recently, however, the chaotic zones of folding studied by Helwig (1970) have been interpreted as tectonic zones of detachment (P.F. Williams, fide W.S.F. Kidd).

Dispersed pelitic material and contorted beds (fig. 3.3) probably suggest that the rocks were not completely
Figure 3.3 Thin-section showing detail of the antiform in figure 3.2A. The section is cut normal to the fold axis and axial plane. S1, oriented upper left to lower right, apparently transects the axial plane. Photograph is about 2 cm wide.
Figure 3.4 Hand sample taken from the same bed as the folds in figure 3.2A (see fig. 3.2B for relative position). Bedding, subparallel to the top and bottom edges, is internally disrupted. S1 is present as "pseudostylolitic" spikes in thin, discontinuous pelitic laminae.
lithified at the time of folding. Furthermore, the folds are strictly local, and small-scale intrafolial folds, with wavelengths on the order of a few centimeters, were not observed in any other graywacke beds. Internal disruption of bedding laminae could have resulted from local sediment liquefaction and injection.

3.2 High Fluid Pressure and Early Deformation Mechanisms

Overpressuring occurs in thick sedimentary sequences when the pore-fluid ratio ($\lambda$) is greater than about 0.46 (Hubbert and Rubey, 1959). The $\lambda$ ratio, not to be confused with a principal strain, is defined as $P_f/\sigma_n$, where $P_f$ is the pore-fluid pressure, and $\sigma_n$ is the total normal stress exerted by the rock (+ fluid) column. Hubbert and Rubey (1959) first recognized that overpressuring can greatly reduce basal sliding friction along large thrust detachments, allowing horizontal movement of thick rock masses over great distances. This concept has been incorporated into many recent models for development of foreland fold-and-thrust belts and accretionary complexes (Cowan and Silling, 1978; Davis et al., 1983). Chapple (1978) predicted that structural thickening in a thrust belt would halt basal sliding if the pore-fluid pressure across the surface did not significantly exceed the hydrostatic pressure. Pore-fluid pressure has the effect of weakening the rock (reducing basal shear stress) by decreasing the
effective confining pressure. The Galice Formation had the following characteristics which, according to Hubbert and Rubey (1959), would have favored high fluid pressures: (1) Compactible pelitic beds serving as impermeable caps, interlayered with permeable sandstone (fluid reservoirs). (2) A total sedimentary thickness in excess of 2,000 meters. While the Galice may not have been this thick immediately after deposition, it was probably structurally thickened by thrusting prior to complete dewatering. (3) Periods of rapid sedimentary or tectonic loading. Massive graywacke and pebble conglomerate units probably indicate high depositional rates during short time intervals.

The onset of underthrusting of the Josephine ophiolite and overlying Galice flysch probably occurred while pore-fluid pressures were reasonably high. Geochronological evidence suggests that the Galice was deformed shortly after deposition (Harper et al., 1986). Calc-alkaline dikes, as young as 150 Ma, intruded poorly lithified sediments in the Galice (Harper, 1980a).

Late Jurassic underthrusting of the Galice and its ophiolitic basement is broadly analogous to underthrusting beneath an overriding plate at low-angle subduction zones. A modern analogue is the Lesser Antilles subduction complex. Rapidly deposited turbidites, transported from Venezuelan river deltas, are presently being subducted at a very low angle beneath older rocks of the Barbados accretionary
complex (Karig and Sharman, 1975). A zone of overpressured sediments occurs beneath a basal decollement of the the overriding plate (fig. 3.5). Fluid is interpreted to be migrating in advance of the wedge causing overpressuring and formation of decollements and mud volcanoes (Westbrook and Smith, 1983). Seismic profiles suggest that deformation within the underthrust sediment pile is restricted to sliding along bedding-parallel detachment faults connected by a small number of ramps (Biju-Duval et al., 1982).

The Galice Formation probably behaved similarly during early stages of Nevadan underthrusting. Layer-parallel sliding along overpressured horizons connected by ramps is suggested to be the dominant deformational mechanism. Thrusting relatively impermeable rocks (the diverse assemblages of the western Triassic and Paleozoic belt) over the Galice Formation could have had the effect of increasing its pore-fluid pressure, further retarding dewatering of the sediments. Early thrust-related structures are not recognized in the Galice, although they may be confused with later thrusts. The following section is a discussion of pre-tectonic structures which formed in response to high pore-fluid pressures.

3.3 Sediment Dewatering

The concept that regional slaty cleavage forms during tectonic dewatering of poorly lithified sediment was first
Figure 3.5 A speculative modern analogue for early underthrusting of the Galice Formation beneath the Late Jurassic continental margin. Sketch shows low-angle thrusting of a thick turbidite sequence beneath the older accretionary prism forming the Barbados ridge in the Caribbean. (after Westbrook and Smith, 1983) See text for details.
suggested by Maxwell (1962). This paper has sparked perhaps the largest controversy among structural geologists studying slaty cleavage development. Maxwell (1962) proposed that slaty cleavage is produced by preferred alignment of detrital mica grains during unidirectional tectonic dewatering of sediments. The strongest premise of Maxwell's (1962) argument is that clastic (sand) dikes are always strictly parallel to slaty cleavage, thus the two are coeval and genetically related. A host of geologists have supported this theory, citing evidence from various slate belts (see Geiser, 1975, and Beutner, 1980, for extensive reference lists).

Harsh criticism of the "dewatering hypothesis" has risen from findings that clastic dikes are generally not parallel to cleavage when viewed in three dimensions (Geiser, 1975; Beutner, 1980). Clastic dikes are soft-sediment structures, and are commonly folded by strains associated with cleavage development (Boulter, 1974; Geiser, 1975; Gregg, 1979). Near-parallelism of some dikes and cleavage can result from rotation of the dikes by tectonic strains (Groshong, 1976; Borradaile, 1977). Furthermore, Beutner and others (1977) showed that slaty cleavage and related folds deform laminated fault-veins which formed during hard-rock folding. A causal relationship between clastic dikes and slaty cleavage is further invalidated in light of detailed studies on the origin of slaty cleavage (e.g. Holeywell and Tullis, 1975; Groshong, 1976; Beutner,
1978; Wright and Platt, 1982; and many others). Therefore, strong evidence suggests that clastic dikes in pelitic flysch sequences were injected prior to complete lithification, before regional penetrative cleavage development.

**Clastic Dikes and Liquefaction**

Sandstone dikes are relatively uncommon in the northern Galice; only three outcrop-scale dikes were observed. Clastic dikes mark locations of fissures that filled with sediment either by forceful injection of a sediment/water mixture (Smith and Rast, 1954), or by penecontemporaneous sediment in-filling (Smith, 1952). The variable morphologies and modes of occurrence of clastic dikes has produced controversy over their origin (see discussions in Powell, 1969; Dionne and Shilts, 1974). In the Galice, as in many other flysch belts, clastic dikes appear identical in composition to other graywacke beds.

All dikes, when viewed in two-dimensional exposures, are oblique to the $S_1$ cleavage (figs. 3.6, 3.7). One of the dikes is folded (fig. 3.6); the axial trace of the folds is approximately parallel to the trace of $S_1$. The dike in figure 3.6 begins as a clastic sill and cuts up-section as a dike. Clastic dikes are known to be folded by differential sedimentary compaction between the dike and its host sediment (Donovan and Foster, 1972; Boulter, 1974; Boulter,
Figure 3.6 Photograph and sketch of a folded graywacke dike/sill. The dike cuts up-section into pelitic beds near the hammer. Younging direction is to the right. Exposure near Reeves Creek Road.
Figure 3.7 30 centimeter-wide clastic dike transecting vertical bedding and the bedding-parallel cleavage in an exposure along Reeves Creek Road. As in most large wacke beds in this area, the dike is unaffected by Sl.
1983). However, the orientations of folds in the dike shown in figure 3.6 suggest that the dike was probably folded during $S_1$-related shortening. The dikes observed near Kerby, Oregon are "isotropic", in that they do not possess primary or secondary foliations.

Small-scale disruption of bedding is frequently encountered in thinly-laminated rocks. Figure 3.8 shows obliteration of bedding laminae, presumably where sediment liquefaction occurred. The disrupted regions in figure 3.8 have a greater pelitic content than the adjacent graywacke layers. This suggests local mixing of thin, pelitic laminae with the more quartz-rich sandstone layers. $S_1$ clearly overprints the disrupted zones in figure 3.8. These may be soft-sediment injection features formed by local migration of a sediment/fluid mixture or "liquefaction". However, because of the small size and weakly-defined boundaries of the structures in figure 3.8, they may be have formed by bioturbation.

Other soft-sediment features commonly found in the Galice Formation are isolated round blobs or "pseudonodules" of coarse-grained graywacke in pelitic beds. They are interpreted as sedimentary load features which separated from the bases of wacke beds prior to lithification. The blobs possess a weak foliation that is parallel to $S_1$ in the pelitic beds; therefore, separation and isolation of the pseudonodules preceded $S_1$ development.
Figure 3.8 Local obliteration of bedding (horizontal in photo) by sediment liquefaction and injection or bioturbation (?) in interlaminated shale, siltstone and graywacke. S1 is developed in pelitic bed at top of photo. Younging direction is upwards. Scale bar=1 cm.
3.4 Melange Zones

Narrow zones of intense stratal disruption are typical in some exposures in the Galice Formation. The best examples are in the horizontal outcrops along the West Fork Illinois River, approximately 0.5 kilometers north of Pomeroy Dam (see ref. map- fig. 1.6). In these zones, subrounded blocks of graywacke are surrounded by shale (fig. 3.9). Tight small-scale folds are sometimes found in the shale matrix. The zones are 2-5 meters wide, and are bounded on each side by coherent strata. Individual graywacke beds in the zones are difficult to reconstruct from the isolated blocks. The term "melange" has frequently been used in descriptions of similar block-in-matrix geometries (Hsu, 1968; Silver and Beutner, 1980).

The zones of stratal disruption in the Galice Formation are comparable to "type 1" melange (Cowan, 1985) observed in the Franciscan assemblage of western California and Oregon. Type 1 melange consists of dismembered beds of sandstone in a pelitic matrix ("broken formation" of Hsu, 1968). Sandstone blocks tend to be crudely aligned parallel to original bedding, and the melange contains no exotic fragments. Blocks are either completely isolated or show pinch-and-swell structure. Type 1 melange generally occurs in tabular zones; they are grossly parallel to, and bounded by, less deformed or undeformed beds. Type 1 melange has
Figure 3.9 Tracing from a photograph of a horizontal exposure of subrounded graywacke blocks in a shaly matrix (type 1 melange of Cowan, 1985). The foliation in the matrix (vertical in this outcrop) appears to be an original bedding-plane fissility that was disrupted during melange formation. Exposure located along West Fork Illinois River, 0.5 km north of Pomeroy Dam.
been attributed to in-situ disruption of poorly-lithified sediments by extensional processes (Cowan, 1985).

The absence of extensional faults or veins in the Galice melange shown in figure 3.9 suggests that it formed at a time when the sediments were insufficiently lithified for brittle fracture. In the northernmost Galice Formation, such ductility cannot be attributed to extension under metamorphic conditions. The strong fissility in the shaly matrix is interpreted as a disrupted equivalent of primary bedding.

Melange in the Galice Formation, a largely ductile feature, is sometimes distinguishable from layers disrupted by fault-accommodated extension (fig. 3.10). It is difficult to determine the relative timing of these brittle extensional structures in the northern Galice. They may be related to regional post-Nevadan normal faulting. Alternatively, they may have formed soon after lithification.

Some possible explanations for bedding-parallel extension (see figure 3.11 A-D), assuming an early stage of development, include the following:

1. Gravitationally-induced lateral spreading. Beds are pulled apart radially during rapid episodes of load-induced dewatering. While pelitic layers may be rapidly compacted, the presence of pore-fluid in graywacke beds delays compaction, and serves to weaken the layers.
Figure 3.10 Layer-parallel extension by boudinage and extensional faulting in an exposure along West Fork Illinois River. Note well-defined faults in laminated pelites. Length of photo is approximately 1 meter.
Figure 3.11 Schematic sketches showing four possible settings for early boudinage and melange development. A. Lateral spreading from sedimentary and/or tectonic loading. B. Down-slope gravitational sliding. C. Extension above oceanic normal faults in ophiolitic basement. D. Layer-parallel shearing. See text for discussion.
(2) **Gravitational sliding** or slumping. Contrasts in density and fluid content cause disruption of layers by differential down-slope movement.

(3) **Normal faults** in ophiolitic basement. Oceanic normal faults in the Josephine ophiolite cause local extension of overlying sediments. Extension in the Galice possibly occurred as a result of gravitational instability (draping effect) above large normal faults. If the faults did not appreciably cut through the Galice, extension of some beds may have resulted from compactional variations directly above fault offsets in the ophiolitic basement.

(4) **Layer-parallel shearing.** Early tectonic bedding-parallel sliding results in boudinage of some beds.

These concepts are schematically illustrated in figure 3.11 A-D. It is uncertain whether local extension of bedding occurred during an early stage of underthrusting of the Josephine ophiolite and Galice Formation. Fisher and Byrne (1987) describe similar extensional structures in an ancient sequence of underthrusted turbidites exposed on Kodiak Island, Alaska, which formed by layer-parallel shear (fig. 3.11 D) combined with vertical loading (fig. 3.11 A). Nevertheless, pre-tectonic, syn-sedimentary origins for early extension in the Galice (figs. 3.11 A,B) are equally plausible. Extension of horizontal strata by load-induced radial spreading is not favored in the Galice, as there is an apparent room-problem unless compressional structures develop simultaneously.
3.5 Discussion

Structures that are unequivocally soft-sedimentary (folds, disrupted bedding) are not widespread in the Galice Formation. They were only observed in the northern field area (fig. 1.6). The "hard rock" nature of D₄ structures (discussed in chapter 4) suggests that the Galice was completely lithified at the time of Nevadan deformation.

It is conceivable, however, that initial Nevadan shortening occurred while the sediments were incompletely lithified. Pre-cleavage injection structures like clastic dikes (figs. 3.6, 3.7) suggest fluid/sediment mobility and locally high pore-fluid pressures. Early underthrusting of the Galice beneath the western Paleozoic and Triassic belt may have involved layer-parallel sliding along overpressured decollements in the Galice. Based on evidence from multi-channel seismic profiles, sliding along layer-parallel decollements was interpreted as an early deformation mechanism in a young turbidite sequence underthrust beneath the Barbados accretionary complex (Biju-Duval et al., 1982; Westbrook and Smith, 1983).

The Galice Formation largely consists of coherent strata; melange (fig. 3.9) is rare and restricted to bedding-parallel zones. Early deformational structures observed in accretionary complexes such as "web" structure, shear foliations defined by microfaults and mappable melange
zones (Byrne, 1984) are absent in the Galice Formation. This suggests that initial shortening in the Galice was not accompanied by intense shearing. Indeed, as emphasized in chapter 6, shearing occurred late in the deformation history.
Chapter 4: MAIN-PHASE NEVADAN STRUCTURES

4.1 $F_1$ Folds

$F_1$ folds in the Galice Formation have been defined as those possessing a weak to strong axial-planar $S_1$ cleavage (Snoke, 1977; Harper, 1980a). They are clearly the first generation of widely-distributed folds. Snoke (1977) termed $F_1$ folds "metamorphic folds" to distinguish them from a second generation of folds, $F_2$, which lack a penetrative axial-planar slaty cleavage. The $S_1$ foliation is folded in $F_2$ hinge zones, while it is always axial-planar to $F_1$ folds. Most $F_1$ folds are observed at the outcrop scale, although a few larger-scale folds were identified from bedding-cleavage relationships.

$F_1$ in the Northern Study Area

$F_1$ folding is widespread in exposures of the northern Galice in the vicinity of Kerby and Cave Junction, Oregon (fig. 1.6). The folds are, however, localized in small groups or fold-pairs. Fold styles vary widely, as evident from figures 4.1 and 4.2. In the Cave Junction area, $F_1$ folds range from open to very tight. The open folds (fig. 4.1) are parallel folds, geometrically resembling class 1C folds of Ramsay (1967). They have rounded profiles and

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Figure 4.1 Open Fl anticline exposed along East Fork Illinois River near Cave Junction, Oregon. S1 forms a weak pencil structure in the shale beds and shows weakly divergent fanning. Pencil long axes are parallel to the fold axis. Note that the anticline is cut by a normal fault to the right of the hammer. Load casts indicate that the steeply-dipping graywacke beds are overturned. Arrows show the younging direction.
Figure 4.2 (top) Tight, angular F1 folds in a steep roadcut, 3 kilometers south of Lake Selmac, Oregon. Folds are overturned to the northwest. Hammer in center of photograph for scale.

Figure 4.3 (bottom) Slightly overturned F1 syncline exposed along Siskiyou Fork Smith River, 4 kilometers west of the Preston Peak thrust. Note ideal similar fold geometry. Parasitic folds are seen to the immediate left of the hammer.
frequently exhibit weakly-developed pencil structure in pelitic beds (fig. 4.1). In contrast, the majority of \( F_1 \) folds in the northern field area are tight or isoclinal, asymmetric, and have angular profiles with interlimb angles of less than 70° (fig. 4.2). \( S_1 \) is weakly to moderately developed in folded pelitic beds. Graywacke beds do not normally possess the \( S_1 \) cleavage in the northernmost Galice Formation, but rather have spaced fractures perpendicular to bedding. Moderate hinge thickening occurs in most pelitic beds in \( F_1 \) fold hinges.

**\( F_1 \) in the Southern Study Area**

South of Oregon Mountain Road in northern California (see fig. 1.9), \( F_1 \) folds are predominately tight to isoclinal. In profile, they range from class 2 to class 3 similar folds (Ramsay, 1967), showing substantial hinge thickening and limb attenuation in most beds. Strong \( S_1 \) development, combined with an absence of bedding markers, frequently obscures the presence of tight \( F_1 \) folds in Galice slates. \( F_1 \) fold closures are rarely observed in exposures south of Oregon Mountain Road (Harper, 1980a; Wyld, 1985; this study). Figure 4.3 is one of few \( F_1 \) fold closures observed in the southern field area.

\( F_1 \) folds are typically overturned or show asymmetry to the west or northwest. Western limbs of the folds are normally shorter, and dip steeper, than the eastern limbs.
At least 70% of outcrop-scale $F_1$ folds in the Galice are northwest facing, in the sense of Shackleton (1958). All folds face upwards; that is, no $F_1$ axial planes have been rotated past horizontal by refolding or other modes of rotation. "Upside down" folds could conceivably exist, as several nearly recumbent $F_1$ folds were observed in both northern and southern field areas.

Stereographic projections of poles to bedding (fig. 4.4) show that overturned beds are evenly distributed among upright beds and are much less common. With few exceptions, beds have northeast-southwest strikes and dip southeast. Rotation of beds around $F_1$ fold axes (fig. 4.6B) did not cause extensive dispersal of points into eastern quadrants. The near coincidence of maxima of poles to upright and overturned beds (fig. 4.5) suggests that large-scale $F_1$ folds are, in a large part, tight to isoclinal with southeast-dipping axial surfaces. Like bedding, $S_1$ cleavage planes generally dip to the southeast, further suggesting tight, asymmetric $F_1$ folding (fig. 4.6A). Post-$F_1$ rotation is evident from north-south and northeast-southwest scatter in poles to bedding (fig. 4.4), poles to $S_1$ (fig. 4.9A) and $F_1$ fold axes (fig. 4.6B).

$F_1$ Fold Axes

Figure 4.6B is a stereographic projection of $F_1$ fold axes measured from bedding-cleavage intersections ($L_{O\times 1}$),
Figure 4.5 Contoured maximum concentrations of poles to upright (dashed) and overturned (solid) beds. Data from northern and southern field areas are combined. Two separate maxima occur in both upright and overturned beds. Total number of points contoured is 271–230 are upright, 41 are overturned.
and directly from outcrop-scale $F_1$ hinges. The axes are widely dispersed in the northeast and southwest quadrants. A weak concentration of points has a south-southwest trend, and an average plunge of $30^\circ$. Figure 4.6B is a combination of $F_1$ fold axes from the northern and southern field areas. No distinct groupings of points were found in data from either field area. Groupings of axes from certain regions or "domains" are not apparent from $F_1$ data.

Northeast-southwest dispersal of $F_1$ fold axes may be the result of rotation by northwest-trending post-Nevadan folds; however, there is no field evidence from this study to support this suggestion. Furthermore, $F_2$ fold axes, discussed in chapter 6, have similar orientations to $F_1$ axes, so that refolding of $F_1$ axes is improbable. Small $F_1$ folds at a single locality sometimes have axes which plunge northeast and southwest in the absence of later folds, suggesting primary non-cylindrical geometries.

Rotation by northwest-trending high-angle faults is an equally plausible explanation for dispersal of $F_1$ axes. Tilting or flexure along post-Nevadan faults could have caused dispersal of pre-existing Nevadan structures. A tight grouping of early fold axes is not expected over a large, multi-deformed region. $F_1$ axes were undoubtedly reoriented by Post-Nevadan faulting in the northern area shown in Plate 1. The orientations of $F_1$ axes are, however, somewhat consistent across small, unfaulted sections of the Galice. From figure 4.6B, it can be presumed that $F_1$ folds
Figure 4.6 A. Lower hemisphere projection of poles to S1. B. Projection of F1 fold axes. Data from both field areas are combined on each stereonet.
and directly from outcrop-scale $F_1$ hinges. The axes are widely dispersed in the northeast and southwest quadrants. A weak concentration of points has a south-southwest trend, and an average plunge of $30^\circ$. Figure 4.6B is a combination of $F_1$ fold axes from the northern and southern field areas. No distinct groupings of points were found in data from either field area. Groupings of axes from certain regions or "domains" are not apparent from $F_1$ data.

Northeast-southwest dispersal of $F_1$ fold axes may be the result of rotation by northwest-trending post-Nevadan folds; however, there is no field evidence from this study to support this suggestion. Furthermore, fold axes, discussed in chapter 6, have similar orientations to $F_1$ axes, so that refolding of $F_1$ fold axes is improbable. Small $F_1$ folds at a single locality sometimes have axes which plunge northeast and southwest in the absence of later folds, suggesting primary non-cylindrical geometries. Alternatively, the northeast-southwest dispersal of $F_1$ axes may reflect reorientation of the axes during thrusting (e.g., Vollmer and Bosworth, 1984).

Rotation by northwest-trending high-angle faults is an equally plausible explanation for dispersal of $F_1$ axes. Tilting or flexure along post-Nevadan faults could have caused dispersal of pre-existing Nevadan structures. A tight grouping of early fold axes is not expected over a large, multi-deformed region. $F_1$ axes were undoubtedly reoriented by faulting in the northern area mapped in Plate
1. The orientations of \( F_1 \) axes are, however, somewhat consistent across small, unfaulted sections of the Galice. From figure 4.6B, it can be presumed that \( F_1 \) folds formed by east-west or southeast-northwest compression, and were dispersed by subsequent deformation. Similar conclusions were reached in previous studies of the Galice Formation (Snoke, 1977; Harper, 1980a; Gray, 1985; Wyld, 1985).

**Large-scale \( F_1 \) Folds**

Across-strike traverses along logging road exposures in the area east of Kerby, Oregon, show the existence of tight to isoclinal \( F_1 \) folds with wavelengths of more than 0.5 kilometers (Plate 1, in back flap). These first-order \( F_1 \) folds were identified from bedding \( (S_0) \)-cleavage \( (S_1) \) relationships and sedimentary younging information. The folds cannot generally be traced through their hinge zones due to breaks in the exposure and intense faulting. Pelitic rocks, particularly in faulted sections of logging road exposures, have usually weathered to reddish clay-rich saprolite. Large \( F_1 \) folds are, therefore, partially exposed in "pieces" between highly-weathered fault zones and breaks in the exposure. Location of \( F_1 \) hinge zones is thus dependent upon recognition of abrupt transitions between continuous sequences of upright and overturned beds.

Large-scale \( F_1 \) folds mapped in this study are overturned to the north or northwest, as shown in the map
and cross-sections in Plate 1 and figure 4.7. Abrupt reversals in the younging direction of uniformly-dipping stratigraphic sequences is evidence that the folds are isoclinal. An example is found along McMullin Creek Road (see fig. 1.6) where a younging reversal is coincident with a narrow zone of brecciation (fig. 4.8). An isoclinal $F_1$ fold was probably sheared-out in its hinge zone. Plate 1 shows the ubiquitous occurrence of fault zones along overturned limbs of large $F_1$ folds. These zones are usually 10-20 meters wide, and are interpreted as Nevadan thrusts, along which nappe-like north or northwest-directed movement occurred.

Large-scale $F_1$ folds were also observed in continuous across-strike exposures along Siskiyou Fork Smith River in northern California. Figure 4.7 is a structural map and cross-section of a six-kilometer traverse across much of the exposed width of the Galice Formation (see fig. 1.9 for location of section). Two large-scale $F_1$ folds were located from bedding-cleavage relationships and a few preserved graded beds. Snode (1977) mapped a large overturned $F_1$ anticline near the Preston Peak thrust, immediately east of the cross-section in figure 4.7. Part of the overturned limb of this fold is exposed in the southeastern portion (right side) of figure 4.7.
Figure 4.7 (following 3 pages) Structural map and cross-section constructed from data collected along Siskiyou Fork Smith River in northern California. Refer to figure 1.9 for location of cross-section (A-A). Cross-sectional view is N25E. Note major reverse faults and large-scale F1 folds. A small, fault-bounded sliver of Josephine ophiolite pillow lavas and overlying hemipelagic chert are exposed at the 2.3 km mark in the cross-section.
CROSS-SECTION SYMBOLS

--- Bedding

--- Nevadan cleavage

 Stratigraphic younging direction

--- Fault

--- Brittle fault zone

\[\text{Silicic dikes and sills}\]

--- Hemipelagic chert

--- Pillow lava (Josephine ophiolite)
Figure 4.8 Sketch of the hinge zone of an isoclinal overturned anticline. A narrow zone of brecciation with a few subhorizontal beds coincides with an abrupt reversal in the younging direction, shown with arrows. Logging road exposure located 8 kilometers east of Kerby, Oregon. View is due-north.
Fold Mechanisms

$F_1$ folds in the Galice probably formed by buckling in some combination with flexural-slip along bedding surfaces. Stiff graywacke beds experienced extension in outer arcs, and compression in inner arcs of some $F_1$ fold hinges (fig. 4.9). Pelitic beds were shortened by pressure-solution (Chapter 5) sub-perpendicular to the $S_1$ cleavage. Weak cohesion between alternating graywacke and shale layers favored folding by flexural-slip and the resultant chevron fold geometries (Ghosh, 1968). In the northernmost Galice, tightening of $F_1$ fold limbs beyond approximately $60^\circ$ was aided by fracturing in fold cores, presumably when flexural-slip was no longer operable.

In contrast, cleavage-related shortening contributed significantly to $F_1$ fold tightening in northern California. Folding occurred under more ductile conditions in the southern Galice. Unlike $F_1$ folds in southern Oregon, parasitic folds are observed along $F_1$ limbs; note the small folds of thin laminae near the hammer in figure 4.3. The kinematics of Nevadan cleavage development and bulk shortening are addressed in chapter 5.
Figure 4.9 Photo and sketch of an F1 fold hinge from Reeves Creek Road. Note excellent preservation of cross-laminae in the graywacke bed. Buckling of the wacke bed was accommodated by "normal faulting" along peculiar dendritic zones and inner-arc "thrusting". A very weak S1 cleavage is observed in hand sample as irregular axial-planar fractures in the shale bed.
4.2 The $S_1$ Cleavage

$S_1$ is a regionally penetrative foliation which ranges from a slaty cleavage or schistosity in the southern Galice Formation to a very weak, discontinuous cleavage in the northern Galice. In exposures near Cave Junction, Oregon, $S_1$ is only apparent in hinge zones of $F_1$ folds, where it is seen as irregular, crudely axial-planar parting surfaces in shale beds. In thin-section, $S_1$ is defined by the subparallel alignment of thin, opaque cleavage lamellae or pressure-solution seams (fig. 4.10). The cleavage lamellae are locally accompanied by a weak grain-shape fabric. In the northern study area, $S_1$ is sometimes a stronger anastomosing cleavage, or locally, a true slaty cleavage.

A strong slaty cleavage is the dominant fabric in pelitic rocks south of the Oregon-California border. Galice slates are typically dark gray to black and possess a strong planar fissility. Cleavage planes are commonly polished and contain one or several sets of slickenside lineations. A reasonably well-developed mineral lineation can be observed on some cleavage planes. In this and subsequent chapters, this mineral lineation will be referred to as "$L_1"$, not to be confused with the $L_{0X_1}$, the bedding-$S_1$ intersection lineation.
Figure 4.10 (top) Discrete opaque folia define S1 in the northern Galice. Sample from an F1 hinge zone near Reeves Creek Road. Length of photograph equals 2 centimeters.

Figure 4.11 (bottom) Strict parallelism of S1 and a well-defined bedding horizon. Sample from Siskiyou Fork Smith River, 5 km southeast of US 199. Length of photo= 6mm.
The Problem of Bedding-Parallel Cleavage

Galice slates and metagraywackes generally possess a weak to strong foliation parallel to bedding, particularly where \( F_1 \) fold hinges are not observed. Figure 4.11 shows thin-section-scale parallelism between the \( S_1 \) slaty cleavage and a sharply-defined bedding horizon. This section is oriented normal to \( S_1 \) and parallel to the \( L_1 \) lineation.

No explanations have been offered by previous workers for this widespread phenomenon. \( S_1 \) is commonly parallel to bedding in \( F_1 \) fold limbs and approximately normal to bedding in the immediate hinge zone. This pattern reflects strong fanning of \( S_1 \) across \( F_1 \) folds. \( S_1 \) folia form a strongly-divergent cleavage fan across the shale bed in figure 4.9; the folia are parallel to bedding on the limbs, gradually becoming axial-planar to the fold (parallel to mesoscopic fractures sketched in figure 4.9) in the immediate hinge zone. Therefore, it can be argued that the occurrence of \( S_1 \) parallel to bedding is not restricted to the limbs of isoclinal folds. Note that divergent fanning was also observed in the core of the anticline shown in figure 4.1.

With the exception of areas having \( F_1 \) folds, \( S_1 \) is consistently parallel to bedding over large regions. This is most evident in the exposures along Oregon Mountain Road, near the Oregon-California border, although it is a regional phenomenon (observed in both field areas and further to the south). \( S_1 \) is parallel to bedding over several kilometers
of exposure, without any evidence of $F_1$ folding. Bedding and $S_1$ are gently warped by large $F_3$ folds in this area.

Similar regionally-penetrative bedding-parallel tectonic foliations have been attributed to large-scale isoclinal folding (Holst, 1985) and to small-scale transposition of thinly-bedded rocks (Hobbs et al., 1976). While there is some evidence for regional isoclinal folding (discussed above in section 4.1) and transposition in the Galice Formation (Snoke, 1977; this study), a strong bedding-parallel cleavage exists even in areas where these structures are clearly absent.

An alternative explanation to isoclinal folding or transposition is that bedding-parallel fabrics are inherited from an original depositional fabric (Williams, 1972; Maltman, 1981). A strong sedimentary fabric may lead to subsequent static recrystallization parallel to bedding. Maltman (1981) proposed that some depositional fabrics are intensified during burial metamorphism or diagenesis in the absence of tectonic stress. This process would normally require a lengthy period of very deep burial and minimal alteration of the fabric by later tectonic foliation development. Alternatively, strong alignment of detrital grains parallel to bedding may be produced during rapid compaction related to dewatering, as suggested by Altermann (1973).

A strong depositional (+compactional?) bedding-parallel fabric is preserved in some Galice pelites where $S_1$ is
weakly developed. This fabric is probably not tectonic in origin, as the well-aligned detrital grains are truncated by $S_1$ cleavage lamellae where $S_1$ is oblique to bedding. In contrast to the observation of Williams (1972) in similar low-grade rocks at Bermagui, Australia, there is no evidence in the Galice for metamorphic mineral growth parallel to an earlier detrital bedding fabric.

In the northern Galice, a bedding-parallel $S_1$ pressure-solution cleavage appears to have been imposed on an earlier depositional fabric. It remains unclear why the regionally-penetrative tectonic foliation ($S_1$) preferentially developed parallel to bedding. In the southern Galice, where penetrative strains were much higher, transposition or isoclinal folding may be more likely origins of bedding-parallel cleavage.

**Microstructure of Slaty Cleavage**

$S_1$ in the southern field area (fig. 1.9) is a domainal slaty cleavage, possessing lenticular domains separated by cleavage lamellae (terminology of White and Knipe, 1978; Weber, 1981). Figure 4.12 is a photomicrograph of the typical domainal slaty cleavage. Lenticular domains consist of monocrystalline quartz, chlorite-mica stacks and framboidal pyrite grains flanked by fibrous pressure shadows. The darker cleavage lamellae are 2-4 microns wide and contain phyllosilicates with fine carbonaceous and
Figure 4.12 (top) Photomicrograph showing dark cleavage lamellae and lenticular domains which define slaty cleavage. Lenticular domains are 10-20 microns wide. Section cut normal to S1 and parallel to L1 (XZ Section). From Middle Fork Smith River, 4 kilometers northeast of Patrick Creek Resort. Figure 4.13 (bottom) Sketch of a bedding-parallel framboidal pyrite layer in slate. Bedding and S1 are horizontal.
pyritic material. Minor carbonate is also interspersed between the cleavage lamellae. Pyrite framboids sometimes occur in long lenses, where they have been disaggregated into constituent pyrite spherules. In many cases, they form long pods or thin bedding-parallel layers (fig. 4.13). In these layers, each spherule is connected to other pyrites by fibrous quartz, or occasionally, white mica. Framboidal pyrite is commonly a major constituent of the rock.

A weakly to moderately-developed lineation (L₁) is observed in thin-sections cut parallel to the slaty cleavage. L₁ is defined by weak alignment of elongate "patches" of quartz, micas and carbonate with patches of pyrite and carbonaceous material. This crude alignment of light and dark-colored (in plane light) "domains" gives the S₁ planes a streaky appearance. Fibrous overgrowths and pressure shadows of pyrites and detrital grains also help to define the lineation. Note that dimensional alignment of individual grains does not contribute significantly to this lineation.

Most Galice pelites contain thin, discontinuous laminae which are distinctively darker than the surrounding rock. They are clearly transected by S₁ in slates and appear to have been folded passively during cleavage-related shortening, showing hinge thickening and limb attenuation (fig. 4.14). They are normally parallel to bedding, where bedding and S₁ are distinguishable. These features are probably organic-rich feeding trails of bottom-dwelling
Figure 4.14 (top) Dark fold-like forms presumed to be trace fossils. Sample from Knopti Creek, near its intersection with Middle Fork Smith River. Length of photo is 6mm.

Figure 4.15 (bottom) Foliated metagraywacke from an exposure along Knopti Creek (near sample locality of fig.4.14). Cleavage folia separate domains of detrital clasts. Average spacing of the folia is 50 microns. Detrital clasts are mostly quartz, chert and plagioclase.
organisms. Sinuous trails are observed on bedding-parallel cleavage surfaces of some slates.

$S_1$ in Metagraywacke

The $S_1$ cleavage in metagraywacke is a domainal fabric (fig. 4.15). It is similar to the "rough cleavage" described by Gray (1978) and interpreted as the psammitic equivalent of slaty cleavage in interbedded pelites. Thin, opaque cleavage folia surround domains containing large grains of detrital origin which average 50 microns in width. Mica and chlorite occur as secondary overgrowths and fibrous beards on the detrital grain margins. Subgrain development and undulatory extinction observed in detrital quartz grains are evidence for intracrystalline deformation; these features, however, may have been inherited from the original source-rock.

Well-foliated metagraywackes and pebble conglomerates possess a strong lineation on cleavage surfaces. In these coarse-grained rocks, the lineation is largely defined by preferred dimensional alignment of quartz, phyllosilicates, lithic clasts, patches of fine-grained quartzose matrix and fibrous overgrowths of detrital grains. In thin-sections cut normal to this lineation, $S_1$ lamellae anastomose around flattened detrital grains. In contrast, the lamellae are more planar in sections cut parallel to the lineation and normal to $S_1$; lensoidal detrital grains are elongated
parallel to the trace of $S_1$, and are connected by fibrous overgrowths and a fine-grained recrystallized quartzose matrix. This lineation is interpreted as the equivalent of the $L_1$ mineral or "stretching" lineation in slates.

**Quartz-Calcite Veins**

Veins consisting of recrystallized quartz and calcite are locally present in slates. The veins occur in several sets which intersect $S_1$ at various angles. The larger veins (0.5–2.0 mm wide) have central cores of strongly-twinned calcite and polygonal quartz grains. When viewed normal to $S_1$ and parallel to $L_1$ (XZ sections), marginal zones consist of finer-grained quartz and possess a foliation parallel to the trace of $S_1$ in the wall-rock. Since the outer quartz "fringes" do not appear to be fibrous overgrowths, they are interpreted as outer zones of large composite veins which attained a foliation during $S_1$ development (i.e., they are pre-cleavage or early syn-cleavage). In addition to being partially transected by $S_1$, the veins are frequently buckled and offset parallel to $S_1$ (figs. 4.16, 4.17). Pre-cleavage quartz veins are extensively shortened, particularly when they were originally at high angles to $S_1$ (fig. 4.18). Intracrystalline deformational features in the veins include subgrain development and sutured grain boundaries.

Although many of the veins were probably deformed during $S_1$-related shortening, initial vein-opening may have
Figure 4.16 Quartz vein (vertical) deformed by buckling and offset along pressure-solution seams. The deformed vein cuts older calcite veins which lie at a low angle to S1 (horizontal in photo). Section cut normal to the vein and S1, and parallel to L1. Sample from a slate exposure along Middle Fork Smith River, 4 kilometers northeast of Patrick Creek Resort. Horizontal dimension of photomicrograph= 1.5mm.
Figure 4.17 Quartz vein offset along a residual zone of dissolution in slate. Tracing from a photomicrograph with S1 horizontal. Enhanced dissolution occurred along the darker zones, presumed to be remnants of trace fossils (see fig. 4.14). Sample from same exposure as figure 4.16 along Middle Fork Smith River. Section cut normal to S1 and parallel to L1.
Figure 4.18 Thin quartz veins shortened by folding. Tracings from magnified images of a thin-section. Sample from same exposure as figures 4.16 and 4.17, also cut normal to S1 and parallel to L1.
been during an early stage of $S_1$ development. The largest veins, commonly perpendicular ($\pm 10^\circ$) to $S_1$, are consistently normal to the $L_1$ lineation on cleavage surfaces. This relationship is especially apparent in a slate exposure along the Middle Fork Smith River and US 199, about 4 kilometers northeast of the Patrick Creek Resort. The orthogonal relationship between the large veins and $L_1$ at this outcrop (fig. 4.19) suggest that the two may be coeval; the extension parallel to $L_1$ was probably responsible for vein dilation. This conclusion is, however, somewhat tenuous, as a temporal relationship between the veins and $L_1$ is not firmly established.

$F_1$ Axes and $L_1$

The three fold axes plotted in figure 4.19 from a slate exposure along Middle Fork Smith River lie at various angles to the mean $L_1$ orientation. This is consistent with variable dihedral angles between $F_1$ axes and $L_1$ measured from other Galice slates in this study. Ramsay and Huber (1983) note that local variations between fold axis orientations and co-generational X-direction lineations are quite common in slate belts. However, few measurable $F_1$ axes and younger slickensides on $S_1$ planes contributed to a paucity of $L_1$ data obtained in this study.

Other workers have found $L_1$ to be generally parallel to $F_1$ fold axes, especially in metagraywackes (Hartper, 1980a;
- stretching lineation ($L_1$)
- trace of extension vein on $S_1$
$\Delta F_1$

Figure 4.19 Stereographic projection of structural elements measured from a slate exposure along Middle Fork Smith River in northern California, 4 kilometers northeast of Patrick Creek. $L_1$ lineations lie at 90° to extensional vein traces on $S_1$ planes. See text for discussion.
Wyld, 1984). Large non-coaxial strains during during a progressive shear deformation may cause rotation of pre-existing fold axes into parallelism with an X-direction lineation (Bryant and Reed, 1969). However, large penetrative ductile shear strains are not recorded by post-$F_1$ structures in the prehnite-pumpellyite facies rocks examined in this thesis. Furthermore, it is unusual that extensive elongation, as large as 300% (determined from syntectonic fibers grown parallel to $L_1$, chapter 5) would occur parallel to the north-south Nevadan structural trend. $F_1/L_1$ relationships provide an interesting structural problem in the southern Galice that could be approached in a well-focused future study.

4.3 Transposition

Local transposition of beds into parallelism with $S_1$ was a result of tight folding, combined with cleavage-parallel extension. Transposition is especially apparent in thin, alternating beds of slate and fine-grained metagraywacke. Snoke (1977) also described transposition in thinly-bedded "laminites" (interlayered siltstone metagraywackes and slates) along the Siskiyou Fork Smith River. One line of evidence for transposition is the frequent pinching-out of thin graywacke beds in the laminites. Fold closures are generally missing completely.
Figure 4.20 is a rare example where isolated fold hinges can be observed in transposed laminites.

A second example of transposition is shown in figure 4.21. This photograph shows isolated graywacke lenses (light color) which are elongated parallel to $S_1$. Wispy trails of graywacke protrude from the larger beds in a direction parallel to the trace of $S_1$. It is uncertain whether this sort of transposed layering was a result of isoclinal $F_1$ folding, as is evident from the laminites in figure 4.20. Pressure solution along spaced cleavage planes probably contributed to the offset and ultimate isolation of individual segments of graywacke beds.

All examples of transposition in the Galice Formation occur in thinly-bedded rocks with strong slaty cleavage. No evidence of transposition was observed in exposures north of the California-Oregon border. It is thus intrinsically related to strong $S_1$ development. A case can be made that Nevadan transposition occurred on a regional scale by isoclinal folding. Reversals in the younging direction occur in the absence of fold closures, as isoclinal hinge zones are narrow and may be obscured by faulting (fig. 4.7).

The main-phase Nevadan deformation involved substantial extension parallel to $S_1$. Cleavage-parallel extension is necessary for the separation of graywacke beds and isolation of fold hinges in transposed rocks (fig. 4.20). Sills showing boudinage parallel to $S_1$ observed near Patrick Creek (fig. 1.9) are further evidence for large Nevadan extensions
Figure 4.20 (top) Transposition of bedding into parallelism with S1. Tracing of silty beds from a photograph showing isoclinal folding and lensing-out of the beds. Non-shaded portions are black slate. Several large pyrite frambooids with quartz pressure shadows are also shown. In this outcrop, S1 dips gently to the southeast. Exposure along Middle Fork Smith River, 3 kilometers northeast of Patrick Creek.

Figure 4.21 (bottom) Transposition of graywacke beds (light color) into parallelism with S1 in slate. Note offsets and lensoidal appearance of transposed graywacke beds. Large offsets are faults. Siskiyou Fork Smith River, 3 kilometers west of the Preston Peak thrust.
(Harper, 1980a). Measurement of extension parallel to \( L_1 \) using syntectonic fibers in Galice slates will be discussed in the next chapter.

4.4 The Preston Peak Thrust

The Orleans fault is a regional thrust fault along which the western Paleozoic and Triassic belt was displaced westward over the western Jurassic belt (see figure 1.1). The fault juxtaposes the Galice Formation with older rocks along nearly the entire north-south length of the Klamath province. It is considered to be a Nevadan-age thrust in light of cross-cutting relationships with several well-dated plutons and regional structural relationships (Klein, 1977; Snode, 1977; Wyld, 1985; Harper et al., 1986). I will use the name assigned by Snode (1977), the Preston Peak thrust, as a local name for the Orleans thrust in the northern Klamath Mountains.

In the field areas of this study, the Galice Formation is faulted against either: 1) Metasedimentary and metavolcanic rocks of the western Paleozoic and Triassic belt (mapped as the Applegate Group in Oregon by Wells and others, 1949), or 2) A serpentinite-matrix melange forming the lowest exposed portion of the Preston Peak ophiolite (Snode, 1977). These rocks have been correlated with the Rattlesnake Creek terrane (Irwin, 1966) to the south (Gorman, 1985). The Preston Peak thrust exposes rocks of
different ages and structural levels in the upper plate. In the area near Cave Junction, Oregon, the upper plate primarily consists of strongly foliated siliceous argillites of the Applegate Group. Irregular bodies of greenstone and serpentinite melange are found sporadically along the thrust. Inclusions in the melange include red chert, metagabbro, metasedimentary rocks and large blocks of massive marble (e.g. "lime Rock" shown on Plate 1). In northern California, the Galice Formation is thrust beneath serpentinite-matrix melange containing tectonic inclusions of amphibolite and, locally, blocks of the Galice Formation.

The Preston Peak thrust has been interpreted as a high-angle reverse fault along its exposure in southwestern Oregon (Wells et al., 1949; Ramp, 1979). Although it strikes roughly north-south along the entire extent of the Klamaths, the trace of the fault curves abruptly toward the west near Cave Junction, where it passes underneath Illinois River alluvium for several kilometers. This unusual curvature was probably produced by large-scale post-Nevadan folding or oroclinal flexure, and is not an artifact of a low-angle fault trace curving around topography (based on mapping by Wells et al., 1949; Ramp, 1979; and field observations from this study). Structural trends in the Galice such as bedding, cleavage and F_1 fold axes are parallel to the curved trace of the thrust (Plate 1). The Preston Peak thrust is cut by numerous small high-angle
faults, which were previously interpreted as transcurrent tear faults (Snoke, 1977; Gray, 1985).

A large re-entrant, in which the Galice is exposed in northern California (fig. 1.1) suggests at least 110 kilometers of displacement along the Orleans fault (Jachens et al., 1986). Where unaffected by post-Nevadan folding, the thrust is relatively flat, dipping 10° to 30° to the east (Wyld, 1985; Jachens et al., 1986). Nevadan displacement along this thrust produced approximately 40% shortening across the Klamath province (Jachens et al., 1986). Ages of plutons intruding the fault suggest that thrusting ceased between 150 Ma and 148 Ma (Harper et al., 1986; Wyld, 1985).

Rocks in the upper plate of the Preston Peak thrust were deformed to some extent during the Nevadan orogeny. Nevadan thrusting reportedly caused remixing of serpentinite melange in the Rattlesnake Creek terrane of the upper plate (Snoke, 1977; Gorman, 1985; Gray, 1985). An east-dipping shear foliation in ultramafic rocks of the Preston Peak complex was interpreted as Nevadan (Snoke, 1977). Siliceous argillites and metaglomerates of the Applegate Group in Oregon possess a penetrative anastomosing foliation. Interestingly, this foliation is often subparallel to the S1 cleavage in the Galice Formation. Clasts in pebbly mudstones are strongly flattened, the least competent being smeared-out in the plane of foliation. Polished foliation surfaces and the locally phacoidal nature of the foliation,
is evidence that the rocks were strongly sheared. It is unclear, however, whether the deformation suffered in the upper plate of the Preston Peak thrust was Nevadan or earlier. These older rocks were probably deformed during a Middle Jurassic deformational event (Fahan and Wright, 1983). Brittle shearing in the upper plate was certainly an important Nevadan effect (Snoke, 1977; this study).

4.5 Nevadan Thrusting and Vergence

Nevadan thrusts in the Galice commonly dip east or southeast and are subparallel to bedding and the typically bedding-parallel cleavage. In the northern Galice, Nevadan thrusts are brittle structures, generally occurring parallel or slightly oblique to bedding (fig. 4.22). With few exceptions, main-phase Nevadan structures pre-date Nevadan thrusts. Truncated $F_1$ folds are widespread in the Galice (see figs. 6.9, 6.10, 6.11).

The relative age of thrusting in the Galice Formation is considered "late Nevadan", as it occurred after the cessation of the main-phase deformation and regional Nevadan metamorphism. No changes in the attitude or intensity of $S_1$ were observed in the vicinity of Nevadan thrusts. This observation contradicts the findings of some previous workers. For example, Kays (1967) documented mylonitization and chlorite-zone recrystallization along Nevadan thrusts in the type-Galice Formation. However, he also notes the
(following page)

Figure 4.22 (top) Nevadan thrust intersecting bedding (and bedding-parallel S1 cleavage) at a low angle. View is to the southwest. Exposure along Knopti Creek Road, approximately 1 kilometer west of the Preston Peak thrust. Note that bedding is discontinuous and lensoidal above and below the hammer, possibly indicating smaller splay thrusts. Down-dip slickensides were observed along the underside of the hanging wall.

Figure 4.23 (bottom) Small-scale thrusts offsetting silty beds. White zones are sigmoidal quartz veins. View is north. Sample from a slate exposure, 4 kilometers north-east of Patrick Creek. See text for additional details.
occurrence of drag folds and other evidence of brittle detachment along Nevadan thrusts. In the southernmost Galice, the dip of $S_1$ steepens progressively away from the Orleans thrust (Wyld, 1985). In the southern extent of the Galice, Nevadan thrusting and $S_1$ development may have been related aspects of the same event.

Some thin section-scale thrusts (fig. 4.23) were at least partially coeval with $S_1$ development. The thrusts in figure 4.23 are generally parallel to $S_1$, the graywacke beds showing offset and thinning between the small thrusts. The faults are marked in places by crudely-sigmoidal polygonized quartz veins. The $S_1$ fabric is intensified against portions of the veins and some veins are cut by cleavage lamellae. Furthermore, the relatively rigid veins and offset wacke beds created triangular-shaped strain shadows where the $S_1$ fabric is substantially weaker. In this particular case, the thrusts may be considered "main-phase" or syn-$S_1$ structures.

In the present study, no systematic variations of main-phase Nevadan structures were observed near the Preston Peak thrust. Thin-sections of slates from within a few meters of the thrust are typical of most Galice slates, showing no evidence for unusually high strains.

The typical southeastward dip of $S_1$ (see poles to $S_1$ in figure 4.6A) suggests dominantly northwestward vergence during the Nevadan orogeny. Nevadan folds and thrusts are also northwest-vergent. Westward thrusting along the
regional-scale Orleans thrust is suggested by increased westward vergence of \( F_1 \) and \( F_2 \) folds and associated cleavages in the Galice as the thrust is approached (Gray, 1985; Wyld, 1985). This structural vergence is evident from the cross-sections in figure 4.7 and Plate 1. The possibility of dominantly eastward Nevadan thrusting (Roure, 1984) is not supported by data in this thesis or most previous studies of the Galice (Snoke, 1977; Harper, 1980a; Wyld, 1985; Gray, 1985; Jachens et al., 1986).

4.6 Summary

The Galice Formation is a uniformly-dipping (fig. 4.4) package of deep water turbidites that were deformed into west and northwest-vergent tight to isoclinal folds during the Nevadan orogeny. Approximately 15\% of all measured beds in the Galice are overturned, although some overturned beds were certainly overlooked in rocks lacking younging indicators. Large-scale \( F_1 \) folds are probably controlled by the most competent stratigraphic intervals—those dominated by massive graywacke beds. Smaller outcrop-scale \( F_1 \) folds are preferentially developed in thinly-bedded rocks.

The three previous sections of this chapter have been devoted to description of structures which formed during a \( D_1 \) or "main-phase" Nevadan deformation. \( F_1 \) folds, \( S_1 \) cleavage and transposition are clearly associated with an initial stage of shortening in the Galice Formation.
Characteristics of the main-phase structures are summarized in figure 4.24. Development of these structures probably coincided with Nevadan regional metamorphism; however, the exact timing of peak metamorphic conditions with respect to development of $S_1$, $L_1$ and syntectonic veins is uncertain.

A strong bedding-parallel cleavage is found over large regions of the Galice where there is no evidence for extensive $F_1$ folding. Although this may result from complete transposition in strongly deformed slates and laminites, the problem remains enigmatic in rocks of the northern Galice where transposition is absent due to low amounts of penetrative Nevadan strain. I suggest that the $S_1$ bedding-parallel cleavage may have formed parallel to an earlier depositional or compactional bedding-plane fabric.
<table>
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<th>Nevadan Structure</th>
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<th>SOUTH</th>
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<tr>
<td>Sl</td>
<td>Very weak to moderately de-</td>
<td>A strong slaty or pyllitic cleavage in pelites, a</td>
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<td></td>
<td>veloped pressure-solution</td>
<td>moderately developed schistosity in metagraywacke</td>
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<td>cleavage, locally slaty in</td>
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<td>existent in graywacke</td>
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<tr>
<td>Fl</td>
<td>Open to tight folds with</td>
<td>Generally tight-to-isoclinal similar folds</td>
</tr>
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<td></td>
<td>minor hinge thickening</td>
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<tr>
<td>Transposition of</td>
<td>None</td>
<td>Occurs in thinly-bedded slates and fine-grained</td>
</tr>
<tr>
<td>Bedding</td>
<td></td>
<td>metagraywackes</td>
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Figure 4.24 Characteristics of main-phase (D1) Nevadan structures in the northern and southern field areas (figures 1.6 and 1.9, respectively).
Chapter 5: NEVADAN CLEAVAGE DEVELOPMENT

5.1 Pressure-solution Origin of Slaty Cleavage

The importance of the process of pressure-solution during slaty cleavage development has been emphasized in a number of detailed microstructural studies (e.g. Groshong, 1975; Holeywell and Tullis, 1975; White and Knipe, 1977; Beutner, 1978). Grain-shape fabric analyses of slates (Holeywell and Tullis, 1975; Beutner, 1978) demonstrate that shortening normal to slaty cleavage may be dominantly by corrosion, diffusion-transfer and neocrystallization of quartz and phyllosilicate minerals. Dissolution of these minerals results in the concentration of less mobile components of the rock, typically carbonaceous material, pyrite and fine micas in cleavage films or lamellae.

Large detrital quartz grains and chlorite-mica aggregates are abruptly truncated by sub-opaque cleavage lamellae or "selvages" in pelitic rocks affected by pressure-solution. In contrast, similar grains that are not enclosed between the lamellae are relatively equant (Groshong, 1975). Dissolved components migrate either by diffusional transfer, or in solution, to sites of active deposition; generally, this will be in "strain shadows" within lenticular domains, or in nearby syn-cleavage veins.
Some new grains may, however, grow within the cleavage lamellae (Beutner, 1978).

Pressure-solution is usually recognized by beds, veins and fossils which are offset along opaque cleavage lamellae or "seams". Beutner (1978) argued that originally rectangular detrital chlorites in the Martinsburg Slate (New Jersey) were aligned parallel to slaty cleavage, and attained lozenge and rhombohedral shapes by selective corrosion of each grain normal to cleavage. This mechanism is shown in figure 5.1B. Pressure-solution, combined with minor intracrystalline deformation, can account for the bulk shortening and preferred grain alignment related to slaty cleavage development (Beutner, 1978). Mechanical rotation of large platy detrital grains is demonstrably minimal in commercial slates such as the Martinsburg Formation (Holeywell and Tullis, 1975; Beutner, 1978; Morris, 1981).

**Pressure-solution in the Galice Formation**

Pressure-solution selvages consisting of opaque carbonaceous material and microcrystalline pyrite are typically intensified at bedding interfaces between slate and metagraywacke in the Galice Formation (fig. 4.10). Dissolution along the selvages causes beds to pinch-out (fig. 4.10) and leads to apparent offsets of pre-cleavage veins parallel to $S_1$ (figs. 4.16, 4.17). At interfaces between slate and wacke beds, pressure-solution selvages
Figure 5.1 A. Tracing of a large chlorite stack from the Galice showing long axis alignment parallel to S1, and traces of the 001 mineral cleavage planes. Inferred original outline of the grain is dashed. B. Variously-shaped chlorites result from corrosion and removal of material outside of dashed lines (after Beutner, 1978, fig. 4).
anastomose around lenses of wacke creating a transposed appearance at thin-section and outcrop scales (fig. 4.22). The dark folia forming the selvages bend markedly around impeding structures like graywacke blobs and large pyrites, suggesting strong mechanical influence by these structures.

Dark cleavage-parallel laminae, interpreted as deformed trails of bottom-dwelling organisms, appear to be sites where dissolution was particularly high. These features were isoclinally folded, probably by volume reduction along the "limbs" and dilational volume-gain in the "hinges". The final geometry of these mechanically passive features resembles a transposed foliation with isolated fold hinges and discontinuous dark laminae, which normally lie at low angles to $S_1$. Discontinuity of the laminae may also reflect original discontinuity of the fossils, as well as those laminae which curve in and out of the plane of the thin-section. Two probable original constituents of these zones, quartz and calcite, were almost entirely removed, leaving only fine micas, opaque material and minor quartz. Cleavage lamellae, defined by the residual material, are thicker and more closely-spaced in these zones, suggesting locally enhanced pressure-dissolution.

Large phyllosilicate grain aggregates or "chlorite-mica stacks" (Voll, 1960) are common in Galice slates. Their shapes resemble those recognized by Beutner (1978): diamonds, rhombohedrons and lozenge-shaped grains (fig. 5.1B). Various origins have been proposed for chlorite-mica
aggregates, such as growth as metamorphic porphyroblasts (Roy, 1978), early mimetic growth parallel to a bedding fabric (Craig et al., 1982) and deposition as detrital grains (Beutner, 1978). The latter suggestion is favored for two reasons. First, chlorite stacks are distinct, large rectangular grains which have 001 mineral cleavage traces subparallel to bedding in uncleaved pelites (Beutner, 1978; this study). Secondly, the aggregates are clearly pre-deformational and pre-metamorphic; their retangular detrital shapes were modified by pressure-solution (Beutner, 1978), and the stacks were not recrystallized during regional low-grade metamorphism.

Figure 5.2 is a photomicrograph of a parallelogram-shaped mica stack in a Galice slate. The 001 mineral cleavage trace is commonly oriented at a high angle to S_. For this reason, chlorite-mica stacks have been termed "cross-micas".

Figure 5.1A is a large chlorite stack in a foliated graywacke siltstone. Intensified opaque cleavage folia along the upper and lower surfaces of the stack indicate selective corrosion along these surfaces. Dashed lines mark the probable original rectangular outline of the chlorite stack. At least 60% shortening by pressure dissolution probably occurred in the vicinity of this grain. Elongated grains with 001 normal to S_ indicate significantly larger amounts of shortening normal to S_.
Figure 5.2 Photomicrograph and sketch of a parallelogram-shaped mica stack in slate. Sample from an exposure along Middle Fork Smith River, 4 kilometers northeast of Patrick Creek. Photo taken under crossed-nicols.
The phyllosilicates in the stacks are sometimes buckled or kinked when 001 planes are at high angles to $S_1$. This suggests the possibility of a small amount of shortening by intragranular deformation. However, kinking is not found in most of the stacks, and there is no microstructural evidence for preferential development of cleavage lamellae along microfolds or kink bands (e.g. White and Knipe, 1978).

In northern Galice pelites with a weaker $S_1$ cleavage, the 001 traces of fine-grained micas are moderately to strongly oriented parallel to bedding. $S_1$ cleavage folia truncate some of the detrital grains; however, a strong bedding-parallel preferred orientation persists.

Figure 5.3 is a different type of bedding-parallel fabric which is crenulated by $F_1$ microfolds. Examination of this thin-section, which is from an $F_1$ hinge zone, reveals buckled and fractured pyrite stringers which were originally parallel to bedding. $S_1$ pressure-solution selvages transect the bedding fabric along the limbs of the microfolds. The mechanical separation of the pyrite stringers from the matrix and other stringers produced dilational openings which were progressively filled with fibrous chlorite and white mica (fig. 5.3). The fibers are strongly curved, an artifact of fiber growth during buckling and rotation of the pyrite stringers. Assuming the fibrous zones represent all of the extended portions of the pyritic layer in figure 5.3, the layer is estimated to have experienced 400-500% bulk extension parallel to $S_1$. In light of the relatively weak
Figure 5.3 (top) Crenulated bedding-parallel pyritic layer in an F1 hinge zone. Bedding is horizontal; S1 is vertical. Sample from Siskiyou Fork Smith River, 3 kilometers east of US 199. Length of photo is 1 cm. (bottom) Extension parallel to S1 caused dilation and fiber growth in zones between crinkled pyrite stringers. Fibrous zones consist of white mica and chlorite (blue fibers under crossed-nicols).
development of $S_1$ in this layer, extension was presumably by local volume-gain.

5.2 Syntectonic Fibers and Pressure Shadows

Syntectonic fibers in pressure shadows adjacent to rigid objects such as pebbles, fossil fragments and pyrite or magnetite grains are excellent indicators of deformation paths and incremental strains in low to medium grade metasedimentary rocks (Choukroune, 1971; Elliott, 1972; Durney and Ramsay, 1973; Wickham, 1973; Gray and Durney, 1979; Ramsay and Huber, 1983). The concept that fibrous minerals in pressure shadows trace displacement paths in a homogeneously deforming continuum has prompted development of techniques of directly measuring incremental strain (all above references plus Beutner and Diegel, 1985; Ellis, 1986). Incremental strain calculations involve measurement of the lengths and orientations of tiny fiber increments with respect to a reference line in the plane of section. Small lines or chords connecting these increments should mimic the curvature of the fibers. Some important pitfalls and sources of error in using syntectonic fibers to deduce strain paths are noted by several authors (e.g. White and Wilson, 1978; Gray and Durney, 1979; Beutner and Diegel, 1985; Ellis, 1986). They are briefly discussed below:

(1) Curvature of fibers must be established as occurring by growth, rather than by intracrystalline
deformation. Curvature by fiber growth is verified when the entire curved length of the fiber goes extinct at once. Extinction which sweeps along a fiber is common, and therefore caution is needed in interpretation of incremental strain data.

(2) Fibers must not curve out of the plane of section. As Ellis (1986) notes, non-Euclidean thin-sections have yet to be invented!

(3) Beutner and Diegel (1985) suggest that only rigid, essentially spherical objects should be considered in fiber studies. Euhedral or irregularly-shaped objects may undergo shape-induced rotations producing misleading fiber curvature patterns. Furthermore, in metamorphic rocks, euhedral minerals, such as magnetites, can grow during metamorphism, and may not record the total incremental strain history.

(4) Fibers must show a consistent, recognizable sense of growth—either towards or away from the object. For objects such as pyrite or magnetite grains, fibers always grow from the matrix toward the object (antitaxial growth), the latest growth increment occurring at the fiber-object interface (Durney and Ramsay, 1973). Evidence for antitaxial fiber growth in pressure shadows of pyrite/magnetite grains is presented in section 5.3.

Perhaps the most ideal rigid objects supporting fiber growth are pyrite "framboids". Framboids are subspherical, raspberry-like aggregates of smaller microcrystalline spherules. The individual spherules are normally on the
order of 1-4 microns in diameter, while the larger
framboidal aggregates range from 12 microns to several
millimeters in diameter. Pyrite framboids are products of
early diagenesis of sediments associated with activity of
sulfate-reducing bacteria (Berner, 1970). They form under
anaerobic conditions during diagenesis of highly
carbonaceous sediments (Berner, 1970). Framboids are most
common in carbonaceous pelitic rocks, although they are
present in some argillaceous limestones and graywackes.

Pyrite framboids in slates are generally flanked by
fibrous minerals in pressure shadows on opposite sides of
the pyrite. Fibers in slates consist of either quartz,
white mica, chlorite, or some combination of the three.
Pressure shadows represent sheltered, low-stress zones of
mineralization on opposite sides of the framboid. Unlike
passively deformed reduction spots, fibers in pressure
shadows of pyrite framboids probably do not record early
shortening by sedimentary compaction.

Figure 5.4 illustrates framboid-matrix interactions
which lead to pressure shadow development (Beutner and
Diegel, 1985). Figure 5.4A shows material flowlines or
particle paths for shape changes produced by a homogeneous
coaxial flattening, with the least principal stretch ($\sqrt{\lambda_3}$)
vertical. Figure 5.4B shows how flattening of the circle
centered at "0" will produce a strain ellipse, S, and
reciprocal strain ellipse, R, which contain the flowlines.
A deformation by mechanical flowage will produce particle
Figure 5.4 Framboid-matrix interactions during a deformation by pure flattening, where the maximum shortening direction is vertical and the maximum extension is horizontal. (after Beutner and Diegel, 1985, fig.4) See text for details.

Figure 5.5 Schematic model showing three successive stages of fiber development (after Beutner and others, 1983, fig.16) Three generations of structures and associated fibers in pelites of the Hamburg sequence (Pennsylvania) record passage of a thrust sheet through a flat-ramp-flat thrust system. See text for discussion.
paths which are deflected around the spherical object (fig. 5.4C). Figure 5.4D represents a purely solutational mechanism by which material is dissolved in a quadrant of compression and reprecipitated in an extensional quadrant.

Crystal fibers grow incrementally in the extensional quadrant as the matrix is, in effect, pulled away from the framboid. Fibers grow progressively in the maximum direction of incremental extension (Durney and Ramsay, 1973). Pressure-dissolution and redeposition is the dominant mechanism contributing to fiber growth. However, mechanical flowage around large framboids can produce exceptionally long pressure shadows which misrepresent the bulk strain (Gray and Durney, 1979; Beutner and Diegel, 1985). Ideal framboid diameters are within the range of 25-75 microns (Beutner and Diegel, 1985).

Non-coaxial Fibers

One of the primary assumptions made in fiber analysis is that curved fibers record a non-coaxial strain history (Elliott, 1972; Durney and Ramsay, 1973). Non-coaxial fibers showing progressive reorientation can form under two general strain conditions. First, curved fibers can grow during bulk simple shear, or alternatively, during bulk simple shear in a spinning reference frame (Beutner and Diegel, 1985). Fiber growth under these conditions could occur in shear zones or on a rotating fold limb within a
shear zone, respectively. Secondly, curved fibers may develop by growth during two phases of flattening, separated by a rigid rotation. This second type of strain history is reflected by fiber curvature patterns in short, overturned limbs of folds in the Martinsburg slate (Beutner and Diegel, 1985). Rotation of the reference frame (bedding, e.g., Wickham and Anthony, 1977; Beutner and Diegel, 1985) with respect to relatively fixed axes of principal strain may explain sharply curved fibers associated with a cleavage that apparently formed by pure flattening.

Alternatively, central fibers in a pressure shadow which remain straight and parallel to cleavage record a coaxial strain history. Straight fibers imply that the direction of maximum incremental extension remained parallel to cleavage throughout cleavage development. Coaxial fibers form on fold limbs that did not experience substantial rigid rotations or shearing during cleavage development (Beutner and Diegel, 1985).

Syntectonic fibers uniquely record penetrative strains associated with the movements of thrust sheets (Beutner et al., 1983; Sample and Fisher, 1986). Figure 5.5 shows the progressive development of three discrete sets of fibers and associated structures within a single thrust slice of the Hamburg sequence in eastern Pennsylvania. Three cleavages with associated fiber sets formed in a sequence of deformational stages, each marked by unique strain conditions. The fiber geometries and associated structures
(fig. 5.5) record strains predicted as a thrust sheet encounters a closing bend, climbs a ramp and passes over an opening bend onto another flat (Beutner et al., 1983).

5.3 Fibers in the Galice Formation

Framboidal pyrite is a major constituent of most Galice shales and slates. The framboids are sometimes pulled apart into their constituent spherules, forming long quartz-pyrite lenses. In Galice slates, excellent fibrous pressure shadows are observed adjacent to spherical framboids. Most framboids have smooth surfaces, a result of the small diameters (1-2 microns) of the constituent spherules. Framboids within the 25-75 micron preferred range (Beutner and Diegel, 1985) are most common. The framboids in this range normally intersect at least several closely-spaced cleavage lamellae, thus recording an average local strain. Smaller individual spherules lie within cleavage lamellae; their pressure shadows probably misrepresent the bulk strain (Gray and Durney, 1979; Beutner and Diegel, 1985).

Pressure shadows in the Galice Formation consist of fibrous quartz, white mica, or intergrown combinations of the two. The distal ends of the fibers begin as overgrowths on grains in the matrix; however, the fiber-matrix interface usually marks a distinct break in optical continuity. In plane light, pressure shadows are easily discernible from the adjacent matrix, as they are conspicuously white. When
viewed under crossed-nicols, however, the fibers in the pressure shadows appear to merge with grains in the matrix.

Fibrous pressure shadows can have various geometries with respect to the pyrite framboid. The four geometries sketched in figure 5.6 are indicative of different strain conditions or deformation mechanisms. Figure 5.6A shows a pressure shadow with equal fiber lengths that preserves the outline of the spherical object ("rigid fiber" model of Ramsay and Huber, 1983). This geometry is indicative of unidirectional extension with negligible shortening in all other directions within the plane of section. Shadows resembling a strain ellipsoid (fig. 5.6B) form under moderate differential strain conditions, where the deformation mechanism is primarily pressure-solution (see fig. 5.4A,B). Figure 5.6C and D are attenuated pressure shadows. Distal portions of the shadows are compressed, no longer preserving the shape of original framboid-matrix boundary. Attenuated pressure shadows may form by dissolution of the most distal fibers, which are removed from the sheltering effect of the host object. Alternatively, they may indicate the influence of mechanical flow during the deformation (Beutner and Diegel, 1985; see fig. 5.4C).

Attenuated pressure shadows with triangular outlines are the most common geometries of pressure shadows in XZ sections of Galice slates. Even though some fibers pinch-out in the distal portion of the pressure shadow, a central
Figure 5.6 Geometries of pressure shadows adjacent to rigid spherical objects. A. Rigid fiber model, B. Shadows defining a strain ellipse, C. Triangular, attenuated pressure shadows, D. Strongly attenuated pressure shadows.
fiber can usually be traced from the framboid to where it ends against, or passes into matrix.

Fiber growth is interpreted as antitaxial in pressure shadows of prites in Galice slates. The oldest material is at the distal end of the shadow; the newest fiber material is located at the pyrite-fiber interface. Several observations are consistent with antitaxial growth. First, the fibers are free of inclusions of matrix material, typically associated with syntaxially-grown fibers (Durney and Ramsay, 1973). Secondly, individual spherules are sometimes "plucked" off the framboid and included in distal portions of the pressure shadow. Finally, there are no solutional interfaces between the fibers and matrix that would suggest syntaxial growth (Ramsay and Huber, 1983).

**Coaxial Fibers**

One of the most important characteristics of fibers viewed in XZ sections of slates in the Galice Formation is that nearly all of them are straight, showing only slight asymptotic convergence from the framboid interface to the distal end of the pressure shadow (fig. 5.7). A few rare shadows with curved fibers probably reflect perturbational effects, as the fibers were forced to grow around large quartz grains or mica stacks. Consistent patterns of fiber curvature were not discovered in any thin-sections.
Figure 5.7 (top) Framboid with fibers adjacent to a large quartz-calcite vein. Length of photomicrograph is 1.1 mm. (bottom) Slightly ovoid framboïd with typical triangular-shaped pressure shadows of fibrous white mica. Framboid diameter is 70 microns. Photos taken in plane light. Both samples are xz sections from an exposure along Middle Fork Smith River (same location as figure 5.2).
In accordance with the findings of Beutner and Diegel (1985), fibers in sections parallel to \( S_1 \) (XY sections) are generally straight, parallel to the \( L_1 \) lineation and preserve the outline of the framboid. Since there is no evidence for shortening or extension normal to \( L_1 \) (Y direction), plane strain conditions are presumed to have prevailed during fiber growth. Additionally, the relatively strong \( L_1 \) stretching lineation observed in both hand samples and thin-sections is evidence that Galice slates were not in the flattening field of the Flinn (1962) diagram.

In one slate sample from near Kerby, Oregon, fibers in pressure shadows of large framboiads viewed in the XY plane show optically-continuous curvature through an angle of approximately 30° (fig. 5.8). The fibers are sharply reoriented into parallelism with \( L_1 \) about half-way along their lengths; the youngest fiber increments (adjacent to the pyrite) are parallel to \( L_1 \). Fiber curvature in the XY plane was not observed in other slates exposed further to the south. It may be significant that the exposure containing this sample, a quarry northeast of Kerby, Oregon, is an unusual outcrop of strongly foliated slate. The rocks at this locality presumably experienced larger flattening strain than the weakly-foliated or non-foliated rocks exposed nearby. Fiber curvature was clearly related to a reorientation of the maximum direction of elongation (+\( L_1 \)) during penetrative deformation.
Figure 5.8 Large framboid in a section cut parallel to S1 (XY section). L1 is approximately horizontal in this photo. Note that the pressure shadows preserve the shape of the framboid. Horizontal dimension of the framboid is 4 mm. The sample is from an exposure along Reeves Creek Road, 3 kilometers southwest of Lake Selmac, Oregon. Photo taken with crossed nicols.
Selective Sampling

One of the primary objectives of studying fibers in the Galice was systematic sampling at localities likely to have experienced non-coaxial strains. Overturned $F_1$ fold limbs and areas adjacent to Nevadan thrusts are potential sites of non-coaxial penetrative strains. Figure 5.9 shows fiber patterns from the overturned limb of an $F_1$ fold in the northern field area. Curved distal portions of the fibers indicate an early reorientation (clockwise as shown) of the maximum extension direction with respect to $S_1$. This curvature is consistent with reorientation by rigid rotation of the overturned limb, as suggested for fiber curvature observed in the Martinsburg slate (Beutner and Diegel, 1985).

An overturned $F_1$ fold limb in more slaty rocks along Middle Fork Smith River unexpectedly contained coaxial fibers, showing no curvature. It is apparent that more samples from overturned limbs of $F_1$ folds are required to make any conclusions concerning strain states across the folds.

As previously noted, Nevadan thrusting did not involve intensification or reorientation of the $S_1$ cleavage in the northern Galice Formation. This is supported by observations of pressure shadows in slates adjacent to Nevadan thrusts, and particularly, the Preston Peak thrust. Fibers in these slates record coaxial strains, similar to
Figure 5.9 Thin-section sketch of fibers and sample locality on the overturned limb of a northwest-vergent F1 fold. Exposure near Reeves Creek Road. See text for details.

$$\sqrt{\lambda_1} = 1.45$$

Figure 5.10 Determination of the maximum principal stretch, $\sqrt{\lambda_1}$, as the tip-to-tip length of the pressure shadows plus pyrite diameter, divided by the pyrite diameter.
most other slates. Therefore, I suggest that shear strains related to Nevadan thrusting had no effect on S1 development or fiber growth.

5.4 Finite Strain

The technique of measuring finite strain from fibers in pressure shadows of spherical objects is rather simple, and does not involve digitization or reiterated multiplication of strain matrices, normally required in incremental strain analysis. The maximum principal stretch, $\sqrt[\lambda_1]$, or $1+e$, is equal to the tip-to-tip length of pressure shadows + framboid divided by the diameter of the framboid, as measured between the central fibers of the shadows (fig. 5.10). Assuming plane strain with no syn-deformational volume change ($\sqrt[\lambda_2]$ equals approximately 1.0), the least principal stretch, $\sqrt[\lambda_3]$, is simply $1/\sqrt[\lambda_1]$.

Fibers were measured from thin-sections of Galice slates cut normal to $S_1$ and parallel to $L_1$ (XZ sections), and parallel to $S_1$ cleavage planes (XY sections). Framboids with no pressure shadows, or very short ones, were observed in sections cut normal to $S_1$ and normal to $L_1$ (YZ sections). These minor fringes may represent off-center cuts through framboids (Beutner and Diegel, 1985). It is therefore assumed that there was no fiber growth parallel to the Y direction. The geometry of pressure shadows in XY sections, resembling the geometry shown in figure 5.6A, probably
indicates no lengthening or shortening parallel to the Y direction (i.e. plane strain during cleavage development). Pancake or disc-shaped pressure fringes (e.g. Williams, 1972b) were not observed in Galice slates.

Pressure shadows used for finite strain measurement were selected according to the following criteria:

(1) Spherical or subspherical framboids that intersect at least several cleavage lamellae were chosen; those larger than approximately 90 microns were omitted.

(2) Pressure shadows which possess a relatively straight central fiber which terminates against matrix at the distal end of the pressure shadow are optimal. This criterion is satisfied in nearly all pressure shadows observed in Galice slates.

(3) Fibers that are bent or deflected around other objects were not considered. Some fibers were forced to grow around large quartz grains, mica stacks or other framboids, resulting in curvature of the pressure shadow. In other cases, the distal portions of unusually short pressure shadows are juxtaposed against large grains. This phenomenon probably occurred by solution-removal of the intervening matrix, and possibly the oldest portion of the shadow. Measurement of these pressure shadows could potentially yield low finite strain values.
Pyrite Stringers

A sample taken along Reeves Creek Road contains long pyrite "stringers" that were stretched, the dilational gaps being filled with fibrous quartz (fig. 5.11). Since they are meandering features, and essentially parallel to bedding, they are interpreted as pyritized equivalents of trace fossils. They are constructed of smaller pyrite spherules, much like the microspherules in spherical frambooids.

Quartz fibers filling the gaps in the stringers are curved when viewed in XY sections, and show characteristics of antitaxial growth (fig. 5.11). The pyrite stringers (fig. 5.11) are from the same sample as frambooids with curved fibers discussed above (see fig. 5.8). Medial inclusions of pyrite and clots of carbonate are found in the vein-like openings. Fortuitously, the long axes of several stringers lie within 10° of the weak L₁ lineation (see fig. 5.11), allowing direct measurement of $\sqrt{\lambda_1}$. Stringers at moderate to high angles to L₁ are not pulled apart, but rather possess fibrous pressure fringes along their lengths. Magnitudes of extension from the stringers are consistent with measurements of pressure shadows of pyrite frambooids in the same section (e.g. fig. 5.8).

Values of $\sqrt{\lambda_1}$ measured in XZ and XY sections are shown in the table and histograms of figure 5.12 for thin-sections from the north and south field areas. Pressure shadows in
Figure 5.11 (left) Tracing from an Sl-parallel thin-section showing stretched frambooidal pyrite stringers and the orientation of Ll. (right) Close-up view of a gap filled with sigmoidally-curved quartz fibers with medial clots of carbonate and pyrite. From same sample as figure 5.8.
Figure 5.12 (top) Table showing $\sqrt{\lambda_1}$ values from fibers in slates from north and south of Oregon Mountain Road (see fig. 1.9). "n" is the number of frambooids measured, "ave.(XZ), (XY)" are the mean $\sqrt{\lambda_1}$ values, and "range" shows the maximum and minimum values of $\sqrt{\lambda_1}$.

(bottom) Histogram showing #pressure shadows measured vs. extensional values from the northern and southern field areas. Measurements from XZ and XY sections are combined. "M" is the mean value.
Galice slates from the area south of Oregon Mountain Road show substantially higher values of $\sqrt{\lambda_1}$. The lack of measured pressure shadows from the northern field area reflects a paucity of pelitic rocks possessing slaty cleavage. Framboids in the typical weakly-foliated or non-foliated pelites of the northern field area rarely support pressure shadows. Furthermore, thin-sections cut parallel to $S_1$ in slates from the northern and southern Galice only rarely intersect framboid-rich layers. Framboids are usually scattered along discrete bedding/cleavage-parallel layers. Lower mean values from XY section may reflect this paucity of data. Nevertheless, the ranges of $\sqrt{\lambda_1}$ values in XY sections fall within ranges of $\sqrt{\lambda_1}$ in XZ sections. The data are rather dispersed throughout the ranges shown in the histograms in figure 5.12.

Assuming plane strain, apparent from XY sections of most Galice slates, the average value of $\sqrt{\lambda_1}$ from the southern area (combined data from both XZ and XY sections) corresponds to a finite strain ellipsoid with dimensions--3.1: 1.0: 0.3 and about 68% shortening normal to $S_1$. The lowest values in the range correspond to 55% shortening. Fibers from slates in the vicinity of Kerby, Oregon give an average shortening value of only 33%. These magnitudes of finite strain are schematically represented in figure 5.13.

Low values of $\sqrt{\lambda_1}$ in slates, such as those determined from fibers in the northern field area, can be attributed to tectonic volume-loss during cleavage development. Large
Figure 5.13 Deformation of a cube with passive spherical markers. X, Y and Z axes and magnitudes of principal stretches in those directions are shown for an undeformed cube, and cubes deformed by finite strains suffered in the north and south field areas (figures 1.6 and 1.9, respectively). Conditions of plane strain with no volume loss are assumed. Note that rocks in both field areas only experienced prehnite-pumpellyite facies metamorphism.
flattening strains normally expected during cleavage formation (a minimum of 55%-60% based on finite strain analyses--Wood, 1974; Beutner, 1978) were not fully compensated by extension in the X direction of northern Galice slates. Wright and Platt (1982) suggest that $\sqrt[3]{\lambda_1}$ is 10% or less in Martinsburg slate that experienced 50% tectonic volume-loss. Similarly, 29%-42% tectonic volume-loss was calculated from a slate in the Hamburg sequence (Pennsylvania), having experienced an extension of $\sqrt[3]{\lambda_1}=1.41$ (Beutner and Charles, 1985). Since some slates in the northern Galice yield similar extensions ($\sqrt[3]{\lambda_1}=1.41$), 29%-42% may be a reasonable estimate of local volume-loss during $S_1$ development.

In northern Galice slates, relatively short fibers and very few veins suggest that some proportion of material was removed, rather than being deposited locally in pressure shadows and veins. These slates may have experienced low fluid pressures and open-system behavior during cleavage formation (Beutner and Charles, 1985), possibly related to their inferred higher structural position (relative to rocks to the south) during main-phase Nevadan deformation and regional metamorphism.

5.4 Strain Inhomogeneity

Outcrop to regional-scale variations in finite strain are reflected in the intensity of folding and cleavage
development in the Galice Formation. In this section, finite strains from fibers are compared with those estimated from mesoscopic Nevadan structures.

**Pencil Structure**

The weak, anastomosing $S_1$ cleavage in the northern Galice is locally observed in the form of "pencil structure". Elongated linear fragments of shale or siltstone weather out of exposures where the rock is gently folded, possesses a weak $S_1$ cleavage and has a strong bedding-parallel fissility (Reks and Gray, 1981). Pencil structure is found in the cores of very gentle folds of bedding, the "pencils" having long axes parallel to the fold axis. Reks and Gray (1981) show that 9-26% shortening normal to cleavage is required for the development of pencil structure. Thus, the lowest amounts of penetrative strain in Galice pelites are probably associated with pencil structure. Massive graywackes without $S_1$ or any other signs of penetrative strain similarly experienced very low amounts of shortening by ductile processes.

**$F_1$ Folding**

Angular $F_1$ folds in the northern field area record as much as 62% shortening normal to the axial planes. This representative value was determined from a series of angular
folds, two of which are shown in figure 4.2. An original (unfolded) length of a graywacke bed was compared with a deformed length measured orthogonal to the axial plane. Folding was apparently by buckling and flexural-slip in graywackes and interbedded graywackes and shales. There is a clear north-to-south transition in the Galice from folding almost entirely by mechanical processes, to folding primarily accommodated by $S_1$-related shortening.

Nevadan strain in the northern Galice is strongly localized. For instance, the tightness and styles of $F_1$ folds, as well as the degree of $S_1$ development, are widely variable. Localized high-strain zones are evident in areas of particularly intense $F_1$ folding and strong cleavage development. Furthermore, large-scale $F_1$ folds (see Plate 1) may record Nevadan strains that are not apparent from outcrop-scale structures.

5.6 Interpretations

Syntectonic fibers in pressure shadows of pyrite framboids in Galice slates record a north-south finite strain gradient during the main-phase Nevadan deformation. Average shortening normal to $S_1$ in the northern and southern field areas are 33% and 68%, respectively. However, it is apparent from figure 5.12 that finite strain is quite variable in both field areas. Low amounts of penetrative finite strain in the northern field area are evident from
weak $S_1$ development and the local occurrence of pencil structure associated with gentle $F_1$ folds. However, relatively low extensions from fibers in slates in the northern field area could be misleading, as they may reflect substantial tectonic volume-loss.

Nevadan shortening in excess of 70% is possible further to the south, in light of strongly flattened pebble conglomerates in the Galice and deformed pillow lavas in the Josephine ophiolite. A comparison of pebbly wackes from Kerby, Oregon with those from Big Flat, California (20 kilometers south of the map area in figure 1.9) reflects extreme north-south variation in penetrative ductile strain (fig. 5.14). As previously noted, metamorphic grade also increases to the south (Harper, 1980a).

$S_1$ development occurred during a coaxial stage of flattening under plane strain conditions, dominantly by pressure-solution, combined with metamorphic recrystallization in rocks in the southern Galice. Although syntectonic fibers record coaxial deformation during $S_1$ development, the possibility of "pre-fiber" non-coaxial deformation cannot be excluded. Rigid rotations of bedding with respect to $S_1$ certainly occurred on limbs of overturned, asymmetric $F_1$ folds, but these large rotations are not recorded by syntectonic fibers observed in this study. Furthermore, there is no evidence from fibers that non-coaxial penetrative strains were experienced in slates near Nevadan thrusts. As will be discussed in the following
Figure 5.14  Comparison of pebbly wackes from Kerby, Oregon (top sample) and the Big Flat area in the southern Galice Formation (lower sample). The large grains are mostly chert, with lesser volcanic and argillaceous clasts. The local trace of bedding is horizontal in both samples.
chapter, Nevadan thrusts are largely post-kinematic, and were superimposed on earlier, main-phase structures.
Chapter 6: LATE NEVADAN DEFORMATION

Nevadan thrusting, which was briefly discussed in chapter 4, clearly occurred after Nevadan metamorphism and the main-phase Nevadan deformation in the northern Galice Formation. This chapter contains a more detailed discussion of the complex geometries resulting from Nevadan thrusting. Two important effects of thrusting were the truncation of $F_1$ folds and the development of angular discordance between sets of beds. $F_2$ folds, and their possible relation to Nevadan thrusting, are described in the first section.

6.1 $F_2$ Folding

A second generation of Nevadan folds, $F_2$, have been defined on the basis of the following characteristics:

1. They fold the $S_1$ cleavage.
2. They locally possess an axial-planar crenulation cleavage ($S_2$).
3. They commonly have kink or chevron fold geometries.

$F_2$ folds in the northern Galice are outcrop-scale structures having wavelengths of less than three meters. They frequently have angular geometries with sharp hinges (fig. 6.1A,C,D), although open, concentric folds are equally abundant (fig. 6.1B). $F_2$ folds are typically found where $S_1$
Figure 6.1 Outcrop sketches of P2 folds in exposures along Knopti Creek (fig. 1.9). See text for details.

Figure 6.2 The S2 crenulation cleavage defined by microfolds and kink bands, locally with discrete opaque folia along their shorter limbs. Length of photo is approximately 1cm. Sample from Knopti Creek, 2 kilometers east of the Middle Fork Smith River intersection.
is parallel to bedding. Apparently, strong anisotropy was a favorable condition for their development. Unlike \( F_1 \) folds, \( F_2 \) folds are normally upright, although somewhat asymmetric. Their variable styles, as well as the local nature of the \( S_2 \) crenulation cleavage in the northern Galice, suggest that the best criterion for identifying \( F_2 \) folds is that they fold \( S_1 \).

In northern California, the two fold generations can be distinguished by fold tightness, the degree of hinge thickening and the nature of the axial-planar cleavage. However, in the northernmost Galice Formation, there is a potential problem in distinguishing \( F_2 \) folds from \( F_1 \) folds, because the latter commonly have a very weak axial-planar \( S_1 \) cleavage. In the Kerby-Cave Junction area in Oregon, \( F_1 \) and \( F_2 \) fold styles (angularity, tightness) show significant overlap, and evidence of structural overprinting (e.g. Williams, 1970) is not always evident. A strong bedding-parallel fissility can generally be traced through the hinges of both \( F_1 \) and \( F_2 \) folds. It was essentially bedding that was folded during both \( F_1 \) and \( F_2 \) episodes in the northernmost Galice. Distinction between \( F_1 \) and \( F_2 \) folds in the area near Cave Junction was based on whether pelitic beds break along axial-planar fractures when struck with a hammer, particularly in water-polished exposures.
The $S_2$ Cleavage

$S_2$ is a crenulation cleavage that is axial-planar to $F_2$ folds. It is uncommon in the northern Galice Formation, only being observed at two locations in this study—near McMullin Creek Road (fig. 1.6) and along Knopti Creek (fig. 1.9). $S_2$ has both discrete and zonal crenulation cleavage morphologies (Gray, 1977). In slates, $S_1$ is normally kinked in a stair-stepping fashion, with the local development of discrete $S_2$ cleavage folia along short limbs of the kinks or microfolds (fig. 6.2). In strongly-foliated metagraywackes and phyllitic metasiltstones, $S_2$ occurs as spaced kink bands, frequently in conjugate sets.

Other than the short limbs of microfolds and discrete cleavage folia, no additional elements define the crenulation cleavage in the northern Galice. There is no evidence for recrystallization or mineral growth in potential strain shadows of rigid objects such as pyrite grains. Fibrous pressure shadows of pyrite frambooids that formed during the main-phase deformation are folded around $S_2$ microfolds and are locally truncated by discrete $S_2$ cleavage folia. Even though $S_2$ cleavage development appears to have involved shortening by pressure-solution, neocrystallization of quartz or phyllosilicates in dilational zones is not observed, possibly suggesting local volume loss.
$S_2$ is a stronger, regionally-penetrative cleavage in southern exposures of the Galice Formation. To the south of the area in figure 1.9, $S_2$ is most commonly a discrete crenulation cleavage with extensive segregation of mineral phases. Recrystallized micas are oriented parallel to $S_2$, suggesting that the cleavage formed during Nevadan regional metamorphism (Gray, 1985). Gray (1985) notes that $F_2$ folds are cross-cut by dikes, similar to those to the west dated at 150 Ma (Saleeby et al., 1982), constraining the age of $F_2$ folding as Nevadan.

$F_2$ Kinematics

Gray (1985) observed a progressive transition from gentle, upright $F_2$ folds to tighter, west-verging folds near the Orleans thrust (fig. 6.3). This implies that $F_2$ folding was contemporaneous with Nevadan thrusting; the higher strains near the thrust produced tighter, west-vergent folds. A west-to-east transition from upright to strongly inclined $F_2$ folds may correlate with increased bulk simple shear as the thrust is approached (Sanderson, 1979).

$F_1$ and $F_2$ fold axes have similar orientations, although variations up to $30^\circ$ in trend and plunge are observed (Harper, 1980a; Gray, 1985). At individual outcrops, $F_1$ and $F_2$ fold axes are commonly oblique to each other. Figure 6.4 is a plot of $F_2$ fold axes and poles to $S_2$ from this study. $F_2$ fold axis orientations are rather variable, although a
Figure 6.3 Schematic representation of the progressive rotation of F2 axial planes (locally defined by S2) towards the Orleans thrust in the southern Galice Formation. F2 folds become tighter and attain westward vergence near the thrust (based on data from Gray, 1985).
Figure 6.4 Stereographic projection of F2 fold axes (top). Data combined from both field areas (figs. 1.6 and 1.9). Similar to F1 fold axes, shown below, F2 axes generally plunge to the south. Poles to S2 from two localities are plotted with F2.
concentration of south-plunging $F_2$ axes is similar to the concentration of $F_1$ axes, shown below for comparison.

$F_2$ folds are found in association with thrusts and brittle shear zones. $F_2$ chevron folds tend to be confined to discrete, highly disrupted fault zones with boundaries roughly parallel to the local attitude of bedding. Some $F_2$ folds may have formed along bedding-parallel detachments, much like a rug is crumpled on a smooth floor. The kink folds sketched in figure 6.1D probably formed by sliding of the slate sequence (thinner beds) along the upper surface of the massive metagraywacke beds (stippled).

Other $F_2$ folds formed by flexure (fig. 6.5) along Nevadan thrusts (Wells et al., 1949; Kays, 1967; this study). The close association of $F_2$ folds with thrusting, and the nature of the $S_2$ crenulation cleavage discussed above, suggest that $F_2$ folding was a second generation Nevadan event ($D_2$). In the northern Galice, $F_2$ folds are strictly brittle structures which fold the $S_1$ cleavage, and locally refold $F_1$ isoclincs (Harper, 1980a).

6.2 Nevadan Fault Zones

In the northern Galice, there are 5-30 meter-wide zones of intense faulting, brecciation and $F_2$ folding. Such fault zones were observed in exposures along Knopti Creek, Siskiyou Fork Smith River (fig. 4.7) and in the area east of Kerby, Oregon (Plate 1). The zones are generally parallel
Figure 6.5 Drag folds showing sense of displacement of a Nevadan thrust. Exposure near Reeves Creek Road, 3 kilometers southwest of Lake Selmac. View is southwest.

Figure 6.6 Isoclinally-folded graywacke bed (outlined) in a brittle shear zone. The bed is graded; the younging direction is shown with arrows. The graywacke isocline is "rootless", as it is suspended in strongly fissile shale and thinner wacke beds.
to less deformed, uniformly dipping beds on either side. At
a few localities, a "scaly" shear foliation, defined by
slightly anastomosing fractures with polished surfaces, is
developed in pelitic rocks. Quartz veins are present in
some of the fault zones in northern California. High shear
strains are evident from strongly reclined isoclinal folds
and dismembered graywacke beds in the fault zones (fig.
6.6).

These fault zones are interpreted as major thrust zones
(or reverse fault zones if steeply dipping), based on
asymmetry of $F_2$ folds in the zones. The fault zones are
most frequently found in strata along overturned limbs of
large-scale $F_1$ folds (Plate 1), and possibly functioned as
basal thrusts for nappe-like movement of the folds. These
brittle fault zones may be similar to regional-scale north-
south trending reverse faults that have repeated the crustal
sequence of the Josephine ophiolite in northern California
(Harper, 1980a). Large reverse faults in the Oregon
Mountain area (northern part of figure 1.9) were interpreted
as younger than Nevadan thrust faults (Vail, 1977). I
suggest that the major fault zones observed in this study
are large Nevadan thrusts, based in their inferred
northwestward vergence and their close association with
large-scale $F_1$ folds and smaller $F_2$ folds.
6.3 Bedding and Fold Truncations

Truncations of sets of beds and F₁ folds are conspicuous structures in rocks of the northern Galice Formation, particularly in the exposures along the east and west forks of the Illinois River (fig. 1.6). A detailed structural cross-section through a well-exposed horizontal outcrop (fig. 6.7) helps to elucidate the geometries of truncated folds and beds. Figure 6.7 is a section drawn normal to F₁ fold axes which plunge approximately 40° southwest. Most notable from this section are high-angle discontinuities between portions of F₁ folds and sets of beds. In the central part of the section, dips of beds vary over a short interval without any sign of a fold closure. This suggests truncation of some beds at depth.

A detailed map view of the southeasternmost fold in the cross-section is presented in figure 6.8. This southwest-plunging anticline is dissected by a fault defined by a covered interval with a few exposed pieces of sheared shale and blocks of graywacke. The fault separates beds of different thickness and orientation (note representative strike-and-dip symbols). Similar truncations, presumed to be faults, are associated with a large number of F₁ folds in the northern Galice. Many (perhaps most) F₁ folds occur as "monoclines" or in small groups between thicker intervals of unfolded beds. At some localities, truncations of F₁ folds
(following page)

Figure 6.7 Detailed cross-section through a 20 meter section of subvertical beds in the Galice. Exposure along West Fork Illinois River near Cave Junction, Oregon. View is to the south-southwest. Arrows show the younging directions, dashed lines are S1 and wavy dashed lines show irregular fissility in sheared shale.
Figure 6.8 Map view of a truncated, southwest-plunging anticline exposed along West Fork Illinois River, 2 kilometers south of Pomeroy Dam. Fold is located at the left-hand edge of the cross-section in figure 6.7. See text for description.
and sets of beds are so common that they may be considered "penetrative" structures at the outcrop scale. The remaining sections of this chapter will be devoted to description of the morphologies of bedding and fold truncations and some possible models for their development.

Examples of Fold Truncations

Four line-drawings in figure 6.9 show the diversity of fold truncation geometries. Truncation surfaces occur at various angles to the limbs and axial planes of $F_1$ folds. Truncations are oriented parallel to the fold limbs (fig. 6.9A) or oblique to them (fig. 6.9B,C,D). Figures 6.9B and 6.10 are examples where the exposed trace of the truncating surface is parallel to the axial trace of the $F_1$ fold. Where younging directions could be determined (e.g. fig. 6.9A,B), beds on opposite sides of the truncating surface young in the same direction.

An unusual aspect of the truncations is that beds terminated by the truncation surface are not traceable on the opposite side of the discontinuity. Fold limbs are typically replaced by more massive beds (fig. 6.9A,B) with different orientations with respect to the truncated limbs. Fault surface features like breccia, gouge and slickenside striations are not present, although the surfaces of some graywacke beds are polished where truncations occur; for example, this is observed along the upper surface of the
Figure 6.9 Sketches of outcrops showing geometries of Fl fold truncations. Younging directions (forked arrows), fold axes (arrows with plunges) and strike-and-dips of beds are shown where determined. Sl is weakly developed in the hinge of the fold in B. A, B and C are from horizontal exposures along West Fork Illinois River. D is from a vertical quarry exposure along Holton Creek Road in Kerby, Oregon.
Figure 6.10 Steeply-plunging syncline that is thrust ed-out in its hinge zone, parallel to another set of beds. The outcrop surface dips slightly toward the viewer; the lower beds strike approximately north-south with north to the right. East Fork Illinois River near Cave Junction, Oregon.
massive wacke beds in figure 6.9B. Beds seldom curve into parallelism with the truncation surfaces, as is the case with drag folds in the Galice (fig. 6.5). Truncation surfaces are sharp and well-defined (fig. 6.11). Thin beds on the left-hand limb of the anticline in figure 6.11 were contorted and juxtaposed against noticeably wider beds of a different orientation.

Small folds that appear to have been thrust-out in their hinges parallel to slaty cleavage have been described from the Martinsburg slate in east-central New Jersey (Beutner, 1978), and from the southern Galice Formation (Gray, 1985). Wilson and DeHedouville (1985) found that many first generation folds in a sequence of low-grade argillaceous rocks in Australia contain faults and features resembling thrust duplexes which produce small-scale repetitions of bedding. Slightly larger outcrop-scale truncated folds have been observed in fold-and-thrust belts and ancient accretionary complexes exposed on land (G.D. Harper, W.S.F. Kidd and T. Byrne, pers. comm.).

**Pseudocross-bedding**

Peculiar obliquity between sets of beds, resembling large-scale cross-bedding, occurs in the northern Galice Formation in the absence of $F_1$ fold closures. Variably-oriented sets of beds are bounded by truncation surfaces. In an example from the "type-locality" of these structures
Figure 6.11  Photograph and sketch of a truncated F1 fold. Beds are contorted and pulled-apart adjacent to a well-defined truncation surface (parallel to pencil, marker). Several siltstone/sandstone beds are stippled in the sketch. Horizontal outcrop along West Fork Illinois River near Pomeroy Dam.
along East Fork Illinois River, truncation surfaces mark angular discordance between beds, and are represented in figure 6.12 as dark lines.

Figure 6.13 shows two examples where younging directions were determined from adjacent sets of beds separated by a truncation surface. In these examples, beds on opposite sides of the truncation young in the same direction, similar to younging observed in truncated folds. In figure 6.13B, both the horizontal beds (upper) and inclined beds (lower) are overturned; the younging direction is "downward". An important conclusion drawn from these observations is that the widespread phenomenon of "pseudocross-bedding" did not result from displacement of fold hinges (parallel to the axial plane) by faulting, because beds on both sides are observed to young in the same direction. Fault truncations parallel to one of the limbs would not result in complete elimination of the fold closure. Alternative causal mechanisms will be explored in section 6.4.

Pervasive bedding truncations are rarely described in the literature. Cross-bedding like structures are commonly observed in transposed sequences (Hobbs et al., 1976). They can be produced by attenuation of a fold limb and consequent juxtaposition with other differently-oriented limbs (Turner and Weiss, 1963). Alternatively, angular discontinuities can develop as small faults in the cores of tight folds in more brittle regimes. Fold closures may be obscured by
Figure 6.12 Photograph and sketch of an exposure along East Fork Illinois River near Cave Junction. Variably-oriented sets of beds form crossbedding-like structure. In the sketch, thin lines show traces of bedding and thicker lines mark the traces of truncation surfaces. Chisel and marking pen for scale.
Figure 6.13  A. (top) Angular discordance between a massive graywacke bed (right) and interbedded wackes and shales. Standard 12x18cm field book for scale. B. (bottom) Unconformity-like truncation of an inclined set of beds. Exposure 5 kilometers east of Kerby, Oregon. Coin is a US quarter. Arrows show younging of beds in both photographs.
propagation of the faults through the hinge zone (Sander, 1911).

In addition to transposition of layering, sedimentary explanations have been evoked to account for angular discordance between sets of beds. Williams and Nealon (1986) describe outcrop-scale packages of beds separated by truncation surfaces in interbedded Silurian sedimentary rocks in western Ireland which are remarkably similar to pseudocross-bedded strata in the Galice. They conclude that the bedding truncations developed from multiple tidal channel migrations, as the truncation surfaces are marked by convolute lamination and sedimentary breccia. This shallow water depositional model can clearly be ruled out in the Galice Formation. Large-scale sedimentary cross-bedding involving shale beds as well as graywackes is atypical of turbiditic flysch sequences; therefore, a deformational origin is inferred.

6.4 Models for Fold Truncations

$F_1$ folds in the Galice Formation commonly appear to have been thrust-out in their hinges parallel to the axial plane. Gray (1985) noted that some $F_1$ folds in the southern Galice were thrust to the west along $S_1$ planes and are now half-anticlines and half-synclines (fig. 6.14). The resultant geometry is similar to the truncated fold in figure 6.10. Gray's (1985) observations suggest that the
Figure 6.14 Line drawing depicting truncation of F1 fold hinges in the southern Galice parallel to S1. (modified after Gray, 1985) See text for discussion.

Figure 6.15 Differential displacement along an F1 axial plane resulting from refoldling around an F2 axis. The F1 hinge zone is bisected, producing crossbedding-like structure (after Ramsay, 1967, fig. 10-29). See text for details.
thrusts only develop along axial planes that intersect the hinges of $F_1$ folds. There are examples in the northern Galice, however, where $S_1$-parallel thrusts cut across a fold limb and do not intersect the hinge of the fold (fig. 6.9B).

Figure 6.15 shows how displacement along an $F_1$ axial plane can be caused by refolding. In figure 6.4, refolding of an angular $F_1$ fold is accompanied by variable amounts of flexural-slip along the $F_1$ axial plane. Note that a geometry resembling cross-bedding forms when an angular fold is truncated and offset along its axial plane. Refolding of $F_1$ folds around $F_2$ axes, however, is rarely observed in the northern Galice. Furthermore, it is unlikely that faults would preferentially develop parallel to the weak $S_1$ foliation in the northernmost Galice.

Cleavage-parallel thrusting, as suggested by Gray (1985) is more evident in strongly foliated rocks in northern California, where there is little or no anisotropy parallel to bedding (i.e., monotonous slate sequences). In thin-section, offsets along some $S_1$ planes lack pressure solution seams, and are instead marked by strongly deformed quartz-calcite veins with inclusions of wallrock material. Pervasive slickensides on slaty cleavage planes are further evidence for cleavage-parallel shearing in the southern field area (fig. 1.9).
Bedding-Parallel Thrusting

Thrusting parallel to bedding was more important in the northern Galice during late Nevadan deformation. Figure 6.16 is a schematic line drawing demonstrating how some fold truncations may have developed by bedding-parallel thrusting. A sequence of beds is folded into tight, kink-like folds with axial traces shown as dots in figure 6.16A. Incipient fractures are dashed lines drawn parallel to bedding in the upright limbs of the folds. Displacement parallel to bedding causes truncation of the folds oblique to their axial planes. Although angular discordance between beds (bedding truncations) is produced in figure 6.16B, fold closures are not completely eliminated.

This mechanism of fold truncation appears to have occurred in the northernmost Galice for three reasons. First, $F_1$ folds tend to have tight, chevron or kink fold geometries (see fig. 4.2). Secondly, individual $F_1$ folds, or small groups of them, are typically bounded by thick, unfolded homoclinal sequences of beds. Fractures are likely to cut across the $F_1$ folds parallel to this strongly preferred orientation of bedding. This is especially evident along West Fork Illinois River near Pomeroy Dam, where there are mutually-parallel "packages" of folded and unfolded beds that are presumably bounded by thrusts. Note that the truncation geometry in part B of figure 6.16 is similar to that of the truncated fold in figure 6.9A.
Figure 6.16 Line drawing showing the development of F1 fold truncations in the Galice by thrusting parallel to bedding.

Figure 6.17 Fold truncation produced by propagation of a thrust ramp. See text for discussion.
Thirdly, this truncation mechanism is favorable because many tight $F_1$ folds in the northern field area contain fractures and minor faults which cut through the hinge zone, parallel to one of the fold limbs. These fractures presumably developed in response to excessive fold tightening and provided surfaces of weakness for subsequent thrust displacement. Thrusting may have occurred preferentially along these fold-related fractures.

Fault-Propagation Folding

Another mechanism that could produce truncated folds is known as "fault-propagation folding". In figure 6.17 (bottom), strongly asymmetric folds accommodate shortening in front of an upward-propagating thrust. Fault-propagation folds are formed and subsequently truncated during the same thrusting event.

This is a rather elegant analogy for the formation and eventual truncation of angular, west-verging $F_1$ folds in the northernmost Galice Formation. In fold-and-thrust belts, fault-propagation folds are usually observed as isolated "rootless" fragments between thrusts (Woodward et al., 1985). This is similar to the geometries of large and small-scale $F_1$ folds in the Galice observed in the area east of Kerby, Oregon (Plate 1) and in outcrops along the Illinois River. Indeed, there may have been an important
causal and temporal relationship between \( F_1 \) folding and Nevadan thrusting.

6.5 The Development of Pseudocross-bedding

Younging directions determined from truncated beds (e.g. fig. 6.13 A and B) indicate that the origin of pseudocross-bedding did not involve the dissection of folds in their hinge zones. Crossbedding-like structure produced by thrusting-out folds would most commonly show opposite directions of younging on either side of the truncation surfaces. Where younging indicators are present in the Galice, truncated sets of beds show the same sense of younging (fig. 6.13). Pseudocross-bedding is normally found in thinly-bedded rocks, however, in which the younging direction is in many places indeterminate.

The development of small-scale ramp-flat thrust systems may explain the younging patterns observed in truncated beds. Comparisons can be made with a model for the progressive development of a thrust duplex. Figure 6.18A shows how thrust sheets are successively imbricated by movement along layer-parallel flats connected by ramps which cut through the stratigraphy (Boyer and Elliott, 1982). The layers are flexed by "fault-bend folding" (Suppe, 1983) over closing and opening bends in the thrust. The final geometry in figure 6.18A shows several angular discordances between layering in adjacent thrust sheets. Specifically,
Figure 6.18 Structural analogies for pseudocross-bedding from fold-and-thrust belts. A. The progressive development of a thrust duplex by successive stacking of thrust sheets. Truncations of layering in adjacent thrust sheets form where the layers are forced to conform to the geometry of the ramp by flexural folding. (after Boyer and Elliott, 1982, fig. 19) B. Drag folds are cut by imbricate thrusts (dashed), producing thrust-bounded "horses" (after Boyer and Elliott, 1982, fig. 9).
discordance occurs along portions of the thrust where the layers have been folded to conform to the geometry of the ramp.

Compression at ramps can lead to asymmetric folding of layers in the hanging wall and footwall by drag-related flexure (fig. 6.18B). Incipient imbricate thrusts, shown as dashed lines in figure 6.18B, propagate through the fold hinges producing a pair of thrust horses. Displacement of the horses would apparently result in complex discordance between layering, as well as truncated folds. Similar imbricate thrusts, parallel to the axial planes of medium to large-scale $F_1$ folds were observed in the Galice Formation, approximately 8 kilometers east of Kerby, Oregon (Plate 1).

Figure 6.19 is a possible example of pseudocross-bedding formed by small-scale thrust imbrication of thinly-bedded slates. The top set of beds (upper left) is gently flexed, creating an angular discordance which increases along a well-defined planar surface. Since no fold closures were observed in this outcrop and $S_1$ is parallel to bedding, it is suggested that the geometry near the top of the outcrop formed at the opening bend between a thrust ramp and flat. Considering the gentle flexure of the upper set of beds, the dihedral angle of intersection between the ramp and flat (angles "$B_1, B_2, \ldots"$ in figure 6.18A) was probably very small. To the right of the hammer in figure 6.19 is a thrust-bounded set of beds which may be analogous to a horse, a thrust slice bounded by imbricate thrusts. Quartz
Figure 6.19 Photograph and sketch of beds from an exposure of interbedded metasiltstone and slate along Siskiyou Fork Smith River, 3.5 kilometers southeast of US 199. View is to the southwest. Traces of bedding show complex angular discordance. See text for discussion.
veins which occur parallel to bedding/S₁, and in gaps between sets of beds, may also attest to a thrusting origin of bedding truncations. Since S₁ is truncated throughout this exposure, thrusting is interpreted as post-D₁.

Similar small-scale ramp-flat geometries have recently been described from the Witwatersrand quartzites in southern Africa (Roering and Smit, 1987). Bedding-parallel thrusts in these rocks are connected by narrow, upward-cutting shear zones serving as ramps. Development of small-scale bedding-parallel detachments, thrust duplexes and thrust-related folds in stiff quartzite layers contributed to as much as 66% bulk shortening in these rocks (Roering and Smit, 1987). In a second example, outcrop-scale ramp-flat systems formed by underplating of slates and graywackes in a subduction complex exposed on Kodiak Island, Alaska (Sample and Paterson, 1988).

6.6 Summary

The latest stage of Nevadan deformation involved F₂ folding, as well as extensive thrusting, which truncated F₁ folds and produced "pseudocross-bedding" structure. F₂ folding was not accompanied by regionally penetrative cleavage development in the field areas studied for this thesis, although S₂ is a penetrative foliation south of Big Flat, California (20 kilometers south of figure 1.9). Nevertheless, F₂ folds are considered to be "Nevadan", 
rather than a later event, because of their probable relationship to Nevadan thrusting and regional metamorphism (Gray, 1985; this study). Recent observations suggest that the $S_1$ and $S_2$ cleavages in the southern Galice formed during a progressive deformation that may be traced by curved quartz fibers in syntectonic veins (G.D. Harper, pers. comm.). However, in the northern Galice, $F_2$ folds clearly formed in conjunction with brittle shearing, under temperatures and hydrodynamic conditions that were not sufficient for penetrative cleavage development. I therefore suggest that late-stage structures in the northern Galice Formation formed during continued post-metamorphic underthrusting, possibly during uplift. The southern Galice presumably represents deeper buried (or thrusted) rocks which experienced larger penetrative strains.

Bedding- and cleavage-parallel thrusting, producing widespread bedding and fold truncations, accommodated late Nevadan (post main-phase) shortening in the northern Galice. The development of pseudocross-bedding probably involved small-scale thrust imbrication by displacement of beds along bedding-parallel detachments and cross-cutting ramps. A general conclusion from observations in this chapter is that outcrop-scale structures can be compared with regional structures involving large crustal sequences. It is unfortunate that the Galice Formation lacks suitable marker beds or lithologies which could aid in recognizing large Nevadan thrusts and quantifying displacement along them.
Chapter 7: POST-NEVADAN FOLDING AND FAULTING

7.1 $F_3$ Folds

Some folds in the Galice Formation are interpreted as $F_3$ folds. These folds are open folds or broad warps of thick sequences of beds that have wavelengths of 0.5 kilometers or more. They are interpreted as $F_3$ folds because they lack any associated cleavage; where observed together, smaller $F_2$ folds have differently oriented axes than $F_3$. $F_3$ folds range from symmetric to slightly asymmetric, and have subhorizontal fold axes. The hinge zones of $F_3$ folds normally coincide with the few places where the Galice is horizontal or very gently dipping.

Several $F_3$ folds were observed in the continuous exposures along Oregon Mountain Road (fig. 1.9). Figure 7.1 shows a limb and hinge zone of an $F_3$ fold with an estimated wavelength of approximately 1 kilometer. It is uncertain whether $F_3$ folds observed in the Galice Formation are the same generation as those folds which affect the crustal sequence of the Josephine ophiolite in northern California. Folds with axes plunging gently to the south-southeast repeat the crustal sequence and basal Galice several times (fig. 1.2), and have wavelengths on the order of tens of kilometers. These regional folds may be Nevadan, although
Figure 7.1 Exposure of a limb and hinge region of an F3 syncline along Oregon Mountain Road. View is to the north-northwest. Beds steepen progressively from subhorizontal (right) to dips of about 60° (left). To the right of the photo, horizontal beds in the hinge of the fold are truncated against vertical beds along a high-angle fault. Length of view is approximately 100 meters.
their north-northwest trending axes do not follow other Nevadan structural trends.

The Lower Cretaceous Horsetown Formation, exposed at Indian Hill, near O’Brien, Oregon (fig. 1.3), constrains at least some of the large-scale folding as younger than Barremian (ca. 119 Ma), the age of the youngest fossils found in this outlier (Imlay et al., 1959). These clastic sedimentary rocks were tilted (and presumably folded) about a northeast-trending fold axis. Three poles to bedding from basal conglomerates and overlying coarse sandstones are plotted in figure 7.2; this figure also shows how minimal post-Nevadan rotations (about 30°) around this axis could account for northwest-southeast dispersal of poles to S₁ in the Galice Formation. Axes of the regional-scale folds of the Josephine ophiolite (discussed above) lie at almost 90° to the inferred Indian Hill fold axis, suggesting that the two are unrelated.

Post-Nevadan folding or oroclinal bending probably produced the strong curve in the trace of the Preston Peak thrust near Cave Junction, Oregon (fig. 1.6, Plate 1). Structural trends in the Galice appear to reflect the bend in the thrust at this location (Plate 1). The thrust was significantly steepened in this area (Wells et al., 1949; Ramp, 1979; this study), in contrast to gently dipping southern portions of the thrust (Snoke, 1977; Wyld, 1985; Jachens et al., 1986).
Figure 7.2 Stereographic projection of poles to S1 in the Galice, poles to bedding in the Cretaceous Horsetown Formation and a presumed fold axis responsible for tilting the Cretaceous rocks. Cretaceous strata measured at Indian Hill, near O'Brien, Oregon (fig. 1.3). Arrows show paths of rotations around the Cretaceous fold axis, possibly accounting for dispersal of poles to S1 in the Galice Formation.
Post-Nevadan folds and some east-west and northwest-southeast trending high-angle faults in the western Jurassic belt may have resulted from Middle(?) Cretaceous accretion of the Franciscan complex to the continental margin (Kays, 1967; Harper, 1980a). The Josephine peridotite is thrust over the Cretaceous Dothan Formation (a member of the Franciscan assemblage) in southwest Oregon along the northern extension of the Coast Range thrust. However, neither the age of this thrust, nor the precise age of the Dothan Formation have been established.

7.2 High-angle Faulting and Regional Uplift

Some late-stage steeply dipping faults in the Galice Formation and Josephine ophiolite may post-date the Cretaceous folding event described above. An example is the northeast-trending Illinois Valley normal fault system (Wells et al., 1949). One of these regional normal faults juxtaposes the Galice Formation with the Josephine peridotite along most of the exposure of the Galice in Oregon (fig. 1.2). This normal fault dips steeply to the east, and probably has a throw of at least 5 kilometers. Smaller normal faults with similar trends have been mapped in the upper plate of the Preston Peak thrust (Snoke, 1977; Ramp, 1979). Normal faults displace Nevadan thrusts, reverse faults and large-scale folds in the Galice and Rogue
Formations (Kays, 1967; Vail, 1977) and may be as young as Oligocene (Wyld, 1985).

Gray (1985) mapped northeast-trending strike-slip faults which truncate F₁ and F₂ folds in the southern Galice Formation. He also mapped a series of right-lateral strike-slip faults in the upper plate of the Preston Peak thrust, some of which offset the thrust. Gray (1985) suggested that F₃ and F₄ folds which deform earlier Nevadan structures formed by oroclinal flexure along these faults.

Right-lateral strike-slip faults in the Galice Formation and upper plate of the Preston Peak thrust may radiate from the Condrey Mountain dome (Gray, 1985; Jachens et al., 1986). The Condrey Mountain dome provides a window through the low-angle Orleans thrust, exposing the Condrey Mountain schist, a possible eastward-transported equivalent of the Galice Formation (see fig. 1.1). Tilting of plutons in the western Paleozoic and Triassic belt was probably linked to doming of the Condrey Mountain schist during early to mid-Tertiary uplift of the Klamaths (Barnes et al., 1986). Displacement along major normal and strike-slip faults may have been related to Cenozoic regional uplift. Miocene bench gravels at the top of the "Klamath peneplain" in northern California attest to approximately 1 kilometer of post-Miocene uplift (Cater and Wells, 1953; Harper, 1980a).
8.1 Conclusions and Interpretations

Important conclusions drawn from this study of the Galice Formation include the following:

(1) As discussed in chapter 3, the Galice was not extensively deformed prior to lithification. Bedding-parallel extensional structures and "type 1" melange are strictly local features which may be related to either sedimentary or tectonic processes. Sliding along layer-parallel decollements and local ramps, however, may have occurred at an early stage of underthrusting, similar to the structural response of modern turbidites being thrust beneath the Barbados Ridge (Westbrook and Smith, 1983). There is little evidence that the Galice was disrupted in a soft-sediment state during tectonic emplacement by underthrusting.

(2) From bedding orientation data (fig. 4.4), it is apparent that the Galice was tightly folded on a regional scale during the main-phase Nevadan deformation. However, the low percentage of overturned beds determined from abundant younging indicators in the northern field area (15%) suggests that large-scale F₁ folds are sparse and interspersed in thick sequences of upright, unfolded strata.
Outcrop-scale $F_1$ folds are apparently sparse in rocks south of the Oregon-California border. Potential factors obscuring $F_1$ folds in the southern field area are transposition of bedding into parallelism with $S_1$ and cleavage-parallel thrusting.

(3) $S_1$ is dominantly parallel to bedding in the Galice Formation. To a large extent, this phenomenon must be a result of extensive isoclinal folding; however, this becomes difficult to reconcile with the scarcity of $F_1$ fold closures (Harper, 1980a; Wyld, 1985; this study) observed in the southern field area (fig. 1.9). In this study, the hinges of large-scale $F_1$ folds were difficult to locate with bedding-cleavage relationships alone, as $S_1$ is rarely at a high angle to bedding. Small-scale transposition of bedding into parallelism with $S_1$ is common, but did transposition in the Galice occur on a regional scale? The scarcity of large-scale $F_1$ hinge zones suggests that they were probably eliminated by post-$F_1$ thrusting (fig. 6.14).

(4) $F_1$ folds and Nevadan thrusts clearly show vergence to the west and northwest, normal to the strike of the Orleans fault. In the area near Cave Junction, Oregon, $F_1$ folds are either overturned to the northwest (fig. 4.2), or have northwest-dipping limbs that are shorter, and dip steeper than southeast-dipping limbs. In the southern study area, $F_1$ folds generally verge to the west-northwest (figs. 4.3, 4.7); slightly different vergence to the south possibly reflects the broadly arcuate geometry (convex toward the
Pacific) of the western Klamaths (see fig. 1.1). The dominantly eastward dips of upright beds, overturned beds (figs. 4.4, 4.5) and the $S_1$ cleavage (fig. 4.6A) reflect large-scale overturned $F_1$ folds with northwest vergence.

The orientations of outcrop-scale thrust faults and the regional Orleans thrust also suggest northwest vergence (figs. 4.22, 6.5). Gravity data indicate that low-density material extends uninterrupted for at least 110 kilometers beneath the Orleans fault (Jachens et al., 1986). Therefore, a major westward component of displacement along this fault is probable. Moreover, large-scale east-dipping reverse faults which repeat the Josephine ophiolite and Galice Formation (Harper, 1980a) are west-vergent, late Nevadan structures (see section 6.2). In conclusion, east-directed Nevadan folding and thrusting suggested in the northern Sierra Nevadan (Day et al., 1985) and the western Klamaths (Roure, 1984) is incompatible with structural observations in this thesis and previous studies (Snoke, 1977; Harper, 1980a; Gray, 1985), as well as interpretations of the sense of displacement along the Orleans thrust (Gray, 1983; Jachens et al., 1986).

(5) There is a strong north-south gradient in Nevadan penetrative strain which is coincident with a change in regional metamorphic grade. Syntectonic fibers in slate from the northern field area record an average of 33% shortening normal to $S_1$, whereas slates from the southern field area were shortened an average of 68% (fig. 5.12).
33% is a maximum estimate of cleavage-related shortening in the northernmost Galice, considering that $S_1$ is typically weakly-developed or absent, and pencil structure, formed after only 9-26% shortening (Reks and Gray, 1981), is commonly associated with gentle $F_1$ folds. Note that shortening values estimated from most $F_1$ folds in this area are much higher (see section 5.4). Plane strain with considerable volume-loss during $S_1$ development may account for low extension values measured from fibers in northern Galice slates. 68% tectonic shortening normal to $S_1$ in the southern field area is similar to shortening values determined from the dimensions of reduction spots (Wood, 1974), chlorite stacks (Beutner, 1978), and lengths of syntectonic fibers in other slate belts (Beutner and Diegel, 1985).

(6) Syntectonic fibers in pressure shadows of pyrite framboids record a coaxial extension history, consistent with pure flattening during formation of $S_1$. Non-coaxial strains, known to occur along rotated limbs of overturned folds (Wickham and Anthony, 1978; Beutner and Diegel, 1985) or in rocks deformed by syn-kinematic thrusting (Beutner et al., 1983; Sample and Fisher, 1986) are not recorded by fibers in slates examined in this study.

A coaxial strain history during cleavage development suggests one of three conclusions about the timing of cleavage formation and fiber growth. First, $S_1$ may have formed by pure flattening of upright $F_1$ folds which were
subsequently inclined during post-kinematic regional simple shear. A second possibility is that fiber growth was restricted to a late stage of flattening and did not record non-coaxial strains related to limb rotation (e.g., Beutner and Diegel, 1985). Thirdly, earliest-formed non-coaxial fiber increments may have been dissolved when the distal portions of pressure shadows were removed from the sheltering effect of the pyrite framboid. This third possibility is not preferred, as shorter pressure shadows also contain coaxial (straight) fibers.

(7) Most Nevadan thrusts in the Galice are younger than the main-phase Nevadan structures (F₁ and S₁). Coaxial syntectonic fibers in Galice slates adjacent to Nevadan thrusts and the Preston Peak (Orleans) thrust suggest that Nevadan thrusting was post-S₁. Moreover, F₁ folds are ubiquitously throrsted-out and S₁ is truncated by small thrusts in the area near Cave Junction and to the south in northern California (Gray, 1985; this study). Pseudocross-bedding and cleavage-parallel slickensides are also interpreted as post-main phase thrust-related structures.

Brittle-Ductile Transition

A general interpretation from the observations made in this thesis is that the northernmost Galice (area of Plate 1) was deformed at a higher structural level than rocks to the south. The Galice near Cave Junction, Oregon, shows
weak $S_1$ development and no visible signs of metamorphism, probably suggesting metamorphic temperatures of less than 200°C. South of Gasquet, California (fig. 1.2), metamorphic temperatures were in the range of 250-300°C (G.D. Harper, pers. comm.) and $S_1$ is a strong slaty cleavage or pyllitic foliation in metapelites and a schistosity in metagraywackes.

In the northern field area, brittle shortening mechanisms (chevron folding, thrusting) were important throughout the Nevadan deformation. Since there are no apparent north-south differences in the tightness of $F_1$ folds or the extent of thrusting, I suggest that total Nevadan shortening ($F_1$, $F_2$ and thrusting combined) was relatively constant over the north-south extent to the Galice.

The north-south gradient in metamorphism and penetrative strain may reflect continued underthrusting during differential regional uplift or tilting. Other factors that could have influenced this gradient are north-south variations in thickness of the sediments and the geometry and timing of closure of the Josephine basin.

Alternatively, underthrusting of the Josephine/Galice plate beneath the continental margin along the Orleans thrust may have had a small southward component of strike-slip, so that deeper structural levels are now exposed to the south. Southeastward underthrusting, slightly oblique
to the continental margin, may account for northwest vergence observed from both field areas in this study.

**Structural History**

Figure 8.1 is a diagrammatic summary of the structural history of the Galice Formation. Structures in the northern and southern field area of this study are sketched in their inferred sequence of development. Brittle-ductile contrasts between structures in the north and south are most notable. \( F_1 \) folds probably initiated by buckling and flexural-slip in rocks in the northernmost Galice. While these fold mechanisms probably operated to the south, penetrative shortening related to \( S_1 \) development may have occurred early in the \( F_1 \) fold history. Tight \( F_1 \) folds formed largely by flexural-slip and kink folding in the northern Galice, whereas fold thinning was dominated by ductile processes to the south. Fracturing occurred in tight, angular \( F_1 \) folds near Cave Junction, Oregon, presumably when flexural-slip was no longer operable. In contrast, isoclinal folding and transposition are evidence of flattening of \( F_1 \) folds in the southern Galice.

The post-main phase period was dominated by development of brittle thrust-related structures. The geometries resulting from bedding- and cleavage-parallel truncation of \( F_1 \) folds are sketched in figure 8.1. Imbrication of beds forming pseudocross-bedding structure was also important
Figure 8.1 Series of sketches showing the structural evolution of the Galice Formation in the northern and southern field areas of this study. Three stages in the structural history are recognized: a main-phase deformation, Nevadan thrusting and post-Nevadan deformation. Post-Nevadan faults (normal and strike-slip) were omitted. Pseudocross-bedding and F3 folds occur in both field areas; thus, they are placed in the center of the diagram.
South

1. Main-Phase
   - F1 Folding
     - S1
     - Isoclinal
     - Transposition
     - S1-Parallel

2. Nevadan Thrusting
   - S0-Parallel
   - Pseudocross-bedding

3. F3 Folds
   - F2
   - S1
   - S2

North

- S0
- Fracturing

1km
during late Nevadan thrusting. Angular discordance between sets of beds is common in both field areas. Therefore, pseudocross-bedding may have contributed significantly to late-stage Nevadan shortening in the Galice.

$F_2$ folds are interpreted as post-main phase structures which may be related to Nevadan thrusting. The timing of $F_2$ development with respect to thrust-related structures such as truncated $F_1$ folds is uncertain. However, since no truncated $F_2$ folds were observed, they are younger than, or coeval with the formation of bedding and fold truncations.

Broad, kilometer-scale $F_3$ folds are the youngest structures shown in figure 8.1; they are interpreted as Cretaceous or younger structures. These folds have horizontal, northwest-trending axes which are markedly different from the northeast Nevadan trends. Furthermore, the folds gently warp Lower Cretaceous sedimentary rocks. Their styles are atypical of folds in sediments deformed by underthrusting (e.g., Byrne, 1984; Sample and Paterson, 1988).

8.2 Further Research

Several structural problems in the Galice may be addressed in future studies. Careful analysis of the problems listed below may lead to a better understanding of the deformational kinematics in the Galice, and related implications for Nevadan orogenesis.
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(1) $L_1/F_1$ Relationships. Determination of any consistencies or variations in the attitude of $L_1$ could improve present understanding of the kinematics of the main-phase Nevadan deformation. It is unclear why $L_1$ is oriented parallel to $F_1$ fold axes (Harper, 1980a) and at a high angle to the inferred west-northwest Nevadan transport direction. A future investigation might focus on why $F_1$ fold axes in the northern Galice appear parallel to the $X$ direction (maximum extension direction) of the finite strain ellipsoid.

(2) Slickenside analysis. Measurement of abundant slickensides on cleavage planes and along Nevadan thrusts could provide additional constraints on the late Nevadan thrusting direction.

(3) Fiber analysis. Galice slates are excellent rocks for analysis of syntectonic fibers, as most samples contain rigid spherical pyrite frambooids. Perhaps the greatest potential lies in fibers in pelitic rocks from the southern Galice, where $S_2$ is a strong regionally-penetrative pressure-solution cleavage in slates. In these rocks, it may be possible to trace fibers related to $S_1$ extensions through some reorientation into parallelism with the $S_2$ cleavage. Such continuous fibers would verify that the two fabrics formed during a single progressive deformation.

From the present study of the northern Galice, it is apparent that no such $S_1$-$S_2$ continuity exists; in rocks containing $S_2$, $S_1$-related fibers in pressure shadows are
simply crenulated along with the \( S_1 \) fabric, and no fiber growth was observed parallel to \( S_2 \). In contrast, metamorphic conditions sufficient for cleavage development and fiber growth presumably persisted during \( F_2 \) folding in the southern Galice (Gray, 1985). Syntectonic fibers in the southern Galice may uniquely reflect non-coaxial strains associated with thrusts or shear zones which were active during regional metamorphism.
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