Pergamon

Journal of Structural Geology, Vol. 20, No. 9/10, pp. 1345 to 1363, 1998

PII: S0191-8141(98)00072-8
0191-8141/98/$ - see front matter

Mid-crustal magmatic sheets in the Cascades Mountains, Washington: implications for magma ascent

SCOTT R. PATERSON

Department of Earth Sciences, University of Southern California, Los Angeles, CA 90089-0740, U.S.A.
E-mail: paterson@usc.edu

and

ROBERT B. MILLER

Department of Geology, San Jose State University, San Jose, CA 95192-0102, U.S.A.

(*Received* 21 *July* 1997; *accepted in revised form* 20 *April* 1998)

Abstract—Diking, diapirism, ascent along faults, and ascent during heterogeneous ductile flow have all been championed as the most important means of magma ascent in the crust. We suggest that these mechanisms are end-members in a complex spectrum of ascent processes. In an attempt to evaluate which combination of ascent processes formed sheet-like bodies in the mid-crustal (20–25 km) Entiat pluton, Washington, we examined the tip regions of these sheets. The sheets have length/width ratios ranging from ~6 to > 75, with increasing ratios strongly correlated to decreasing sheet tip radii (from 850 to 100 m) and decreasing ratios of tip diameter/sheet width (from 0.66 to 0.33). Thus, these bodies have geometries falling between those of dikes and those associated with elliptical diapirs. The sheets are not associated with faults or fracture zones extending from their tips. Instead, sheet walls are oriented parallel to the axial planes of upright, syn-emplacement folds. In sheets with high length/width ratios, magmatic foliations in sheets are folded or parallel to axial planes of host rock folds. With decreasing length/width ratios, margin parallel foliations in both sheet and host rock are increasingly common. Our studies indicate that the sheets are emplaced at high angles to σ_1, not σ_3 as proposed in elastic dike models, and are always associated with complex, viscoelastic flow of host rock. These observations rule out elastic dike and fault models, and instead favor diapiric rise of magma sheets during viscoelastic behavior of host rock.

INTRODUCTION

A topic of much interest in the recent geological literature is the means by which magmas ascend through the lithosphere (Brown *et al.*, 1995; Collins and Sawyer, 1996; Petford, 1996; Weinberg, 1996). The mechanisms by which this is accomplished will have a relatively large impact on the growth and evolution of orogenic belts. For example, thermal–mechanical models of orogens show that the thermal structure and rates at which mass is redistributed in the crust strongly influence both surface and subsurface processes in these orogens (e.g. Ruppel and Hodges, 1994), sometimes long after these processes occurred (e.g. Koons, 1987). Magma transport is particularly important because the rates of this process (from meters per year to meters per second) will transfer mass, and thus heat, much more rapidly than tectonic processes (millimeters to centimeters per year (e.g. compare Paterson and Tobisch, 1992 to Ruppel and Hodges, 1994)).

Some progress, using theoretical and experimental studies, has been made on understanding possible ascent mechanisms (Lister and Kerr, 1991; Weinberg and Podladchikov, 1994). But ultimately it is important to examine the rock record to determine which of the possible ascent mechanisms operates in nature and to explore the interplay between different mechanisms. For example, mid-crustal (~20–24 km) sheet-like magma bodies exposed in the Cascades Mountains, Washington (Fig. 1) at first glance appear to support magma ascent in dikes or along faults. In a few cases, we have located the tips or lateral edges of these bodies and have examined both host rock and internal characteristics of these tips. These sheets are not associated with faults, have length/width ratios intermediate between typical dikes and diapirs, have narrow blunt tips, and intrude along the axial planes of active folds. After first exploring the possible interplay between magma ascent mechanisms, we use the characteristics of tips of intrusive sheets in the Cascades to argue that their ascent was not controlled by faulting or diking but by a combination of processes operating in a rheologically complex, actively deforming crust.

Magma ascent

Four end-member models—diking, ‘hot Stokes’ diapirism, ascent along faults, and ascent during heterogeneous ductile flow of host rock—have been championed as likely magma ascent mechanisms

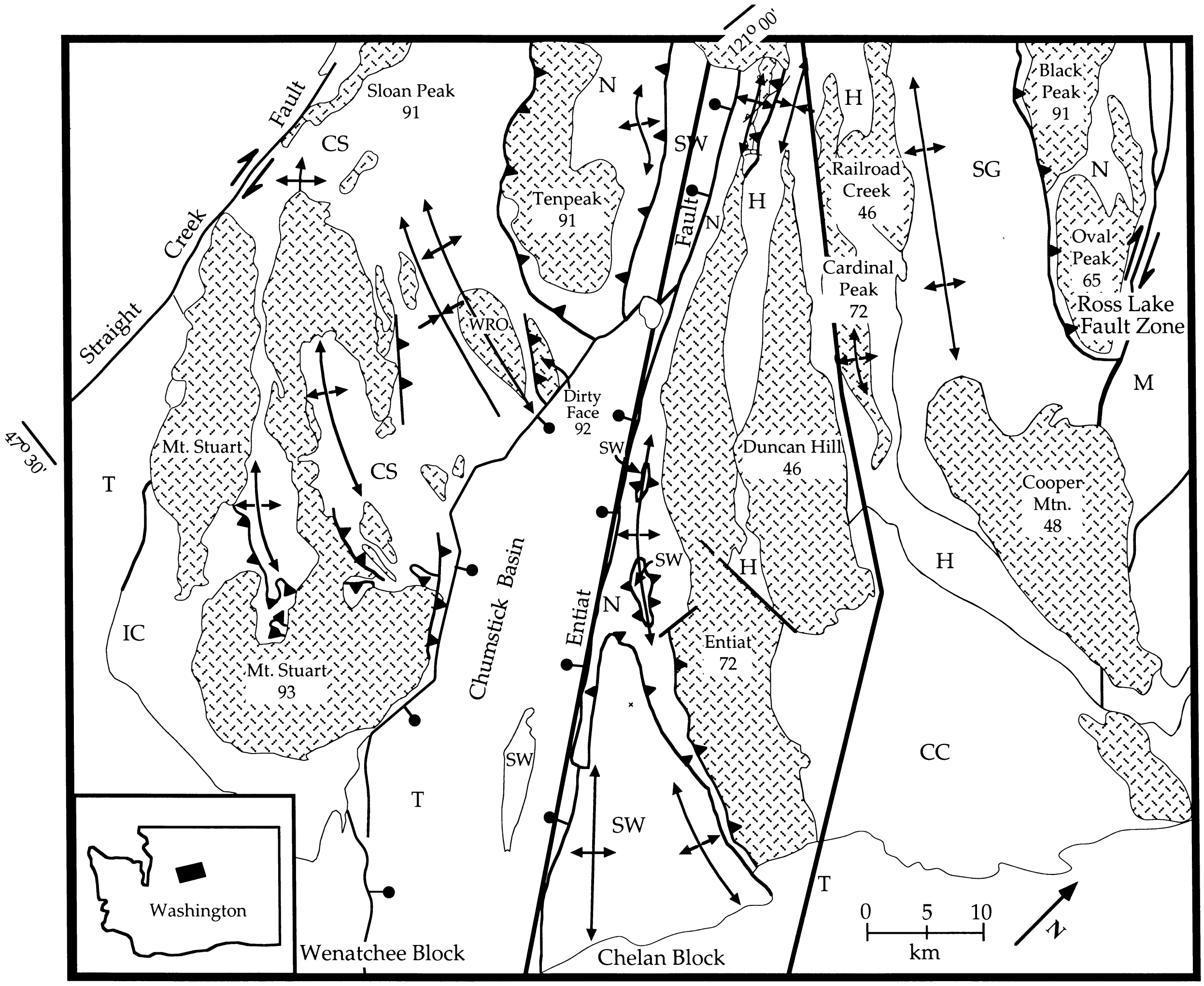

Fig. 1. Regional map of southern Cascades Core, Washington emphasizing Late Cretaceous to early Tertiary structures. Inset shows location in Washington State. Cross hatched pattern = plutons (names and ages (my) inset). Country rock units as follows: CC = Chelan Complex; CS = Chiwaukum Schist; H = Holden Unit; IC = Ingalls Complex; M = Methow Basin; N = Napeequa unit; SW = Swakane gneiss; T = onlapping Tertiary units mainly Columbia River basalts to south and southeast and continental basin sediments in the center of the diagram. Straight Creek, Entiat, and Ross Lake fault zones = oblique slip Tertiary faults that divide the Cascades core into Wenatchee and Chelan structural blocks. Within these blocks are shown the main Late Cretaceous thrust faults and axial traces of regional-scale folds. Smaller-scale folding is widespread in all pre-Tertiary country rock units and locally developed in many of the plutons. Box shows location of Fig. 3.

(Clemens and Mawer, 1992; Brown *et al.*, 1995). While a fifth mechanism, porous flow, has also been proposed (e.g. McKenzie, 1984; Stevenson and Scott, 1987), most workers generally agree that magma cannot ascend long distances in the crust in a geologically reasonable time frame by this mechanism (e.g. Clemens and Mawer, 1992; Brown *et al.*, 1995).

Each of the first four mechanisms results in unique magma–host rock characteristics. Dike models assume that the host rock behaves elastically and that low viscosity magma is transported along extensional fractures resulting in magma bodies with high length to width ratios ($L/W > 1000$) (Shaw, 1980; Lister and Kerr, 1991). Hot Stokes diapir models assume that a roughly spherical or tear-drop shaped batch of magma ($L/W < 3$) rises buoyantly through the crust. Host rock, often assumed to be a homogeneous Newtonian material with a temperature-dependent viscosity, is heated by the magma, thus allowing it to be displaced by ductile flow around the body and down towards the magma generation site (e.g. Marsh, 1982; Schmeling *et al.*, 1988). Weinberg and Podladchikov (1994) noted that host rocks will follow power-law behavior where viscosity is a function of both temperature and strain rate. Since host rock temperature and strain rate are a function of distance from the pluton margin, power-law behavior will result in narrow aureoles, greater ascent distances, and faster ascent rates (Weinberg and Podladchikov, 1994).

Recently, much attention has focused on the ascent of magmas along faults (e.g. D'Lemos *et al.*, 1992; Berger *et al.*, 1996) and/or emplacement in local extensional zones in these faults (e.g. Hutton, 1988). In spite of its increasing popularity, ascent of magma along faults is a poorly defined process. Most authors do not state whether the fault is simply acting as a zone of weakness (the actual means by which magmas ascend and displace host rock being left unstated), or that fault movement is somehow assisting magma ascent, or that magma ascent is occurring by flow into localized extensional regions which then migrate along the fault. Paterson and Fowler (1993) noted that these models require the appropriate location, geometries, and rates of regional faulting, and that sheeted bodies are the most likely outcome at mid- to shallow crustal levels.

A recent suggestion is that during regional deformation, magma moves along an interconnected network of low stress sites, such as those occurring in pressure shadows or tension gashes (Brown *et al.*, 1995; Collins and Sawyer, 1996). Since most low stress sites are small, the result is a migmatitic terrain with small interconnected to discontinuous batches of magma. Magma ascends by migration of low stress sites or by migration of melts along very irregular, temporally changing paths, both of which decrease the rate of magma ascent and increase the likelihood of magma freezing. Therefore, this ascent mechanism is most appropriate for the lower crust where host rock temperatures are relatively high.

These four mechanisms assume rather simplified magma–host rock behavior and represent end-members in what is potentially a complex spectrum of magma ascent processes (Fig. 2). They may dominate magma ascent under certain circumstances, but we believe that ascent in natural systems commonly involves two or more of these processes (e.g. Rubin, 1993). Figure 2 is our attempt to summarize graphically the transitions between these end-member models. Text inside the box summarizes the dominant behavior of magma during each ascent mechanism; whereas, text outside the box summarizes the dominant behavior of host rock. Text in the four corner boxes indicates the type of intrusion that results from each set of processes. This diagram also emphasizes the natural transitions between different ascent mechanisms of both magma and host rock behavior. For example, Rubin (1993) examined the interplay between viscous flow and elastic deformation of host rock (i.e. top horizontal axis in Fig. 2) in an attempt to evaluate under what conditions magma is more likely to ascend in diapirs or dikes. He concluded that viscosity

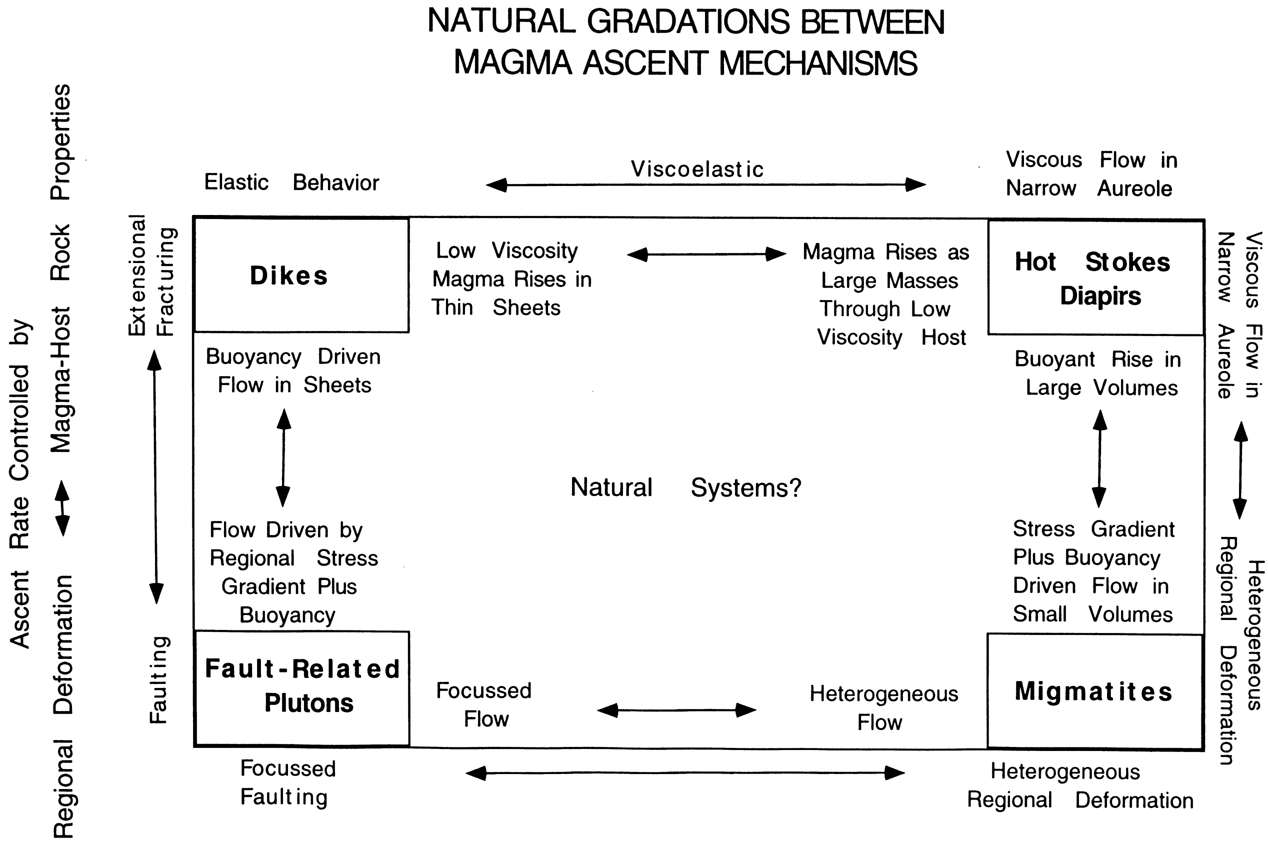

Fig. 2. Figure emphasizing the natural transitions between different ascent mechanisms. The igneous bodies resulting from different proposed ascent mechanisms are shown in small boxes in each corner of the large rectangle. The main characteristics of magma behavior (inside the large rectangle) and of host rock behavior (outside the large rectangle) are noted for each end-member mechanism. Thus the diagram can be usefully viewed as a magma chamber (large rectangle) or as a box emphasizing transitions (arrows along sides and diagonals) between different ascent mechanisms. See text for discussion of these transitions.

contrasts between magma and host rock of between 11 and 14 orders of magnitude are required before viscous deformation of the country rock is negligible (i.e. magma ascends strictly by elastic diking). Rubin further noted that if both fracturing and viscous flow of host rock occur during magma ascent, then (1) magma bodies with intermediate L/W ratios may form, (2) ascent rates will fall between that for dikes and diapirs, and (3) magma bodies may rise as sheets in orientations other than that predicted by elastic dike models.

Clearly, transitions also occur between the other ascent mechanisms. A transition between dikes and fault-related plutons (left side of Fig. 2) will occur as the amount of fracture parallel displacement increases, such as that described along Proterozoic dikes by Cadman *et al.* (1990). Rates of magma ascent will again be intermediate between those predicted for dike and fault models with orientations of magma bodies increasingly controlled by stresses along the fault. Faults or extensional fractures may form low stress sites, indicating that a complete gradation exists between fault/dike models in which magma flow and host rock deformation are localized, and migmatite models in which host rock deformation is widespread and magma flow dispersed (e.g. Collins and Sawyer, 1996). One can also imagine a transition between magma ascent in migmatites and in hot Stokes diapirs (right side of Fig. 2), where increasingly large and increasingly centralized magma batches ascend buoyantly along low stress pathways during ductile flow of host rock. Finally, a transition between hot Stokes diapirs and fault-related plutons can occur when the style and location of ascent of hot Stokes diapirs are increasingly influenced by the location and slip rates of active faults (a possible example is the Bergell pluton described by Berger *et al.*, 1996).

Figure 2 displays several overall changes between the different ascent mechanisms. For example, host rock deformation changes from elastic behavior in the upper left to heterogeneous viscous flow in the lower right. Faulting is most important in the lower left and least important in the upper right corner. Systems in which ascent rates are largely controlled by magma–host rock characteristics fall at the top of the diagram, whereas those largely controlled by regional deformation fall at the bottom. Figure 2 is still an oversimplification: other end-member processes are possible. For example, there is much evidence that the rise of diapirs with small L/W ratios commonly involve temporally and spatially changing host rock behavior and not just simple viscous flow in narrow aureoles (e.g. Buddington, 1959; Paterson *et al.*, 1996). Nevertheless, we feel that this diagram is useful in emphasizing the transitions between ascent mechanisms and thus the need to determine, when studying natural systems, the relative importance of, and interplay between, different mechanisms rather than searching for a single end-member mechanism.

One useful means of evaluating natural systems is to examine the former propagating margins or 'tips' of igneous bodies, that is the host rock occupying the region into which magma was moving. This region potentially preserves a great deal of information about host rock transfer processes during magma ascent. For example, examination of dike tips at shallow crustal levels has provided much information about how dikes form and propagate (e.g. Pollard *et al.*, 1975; Baer and Beyth, 1990). These authors noted that a process zone forms at dike tips in which extensive fracturing and some crystal-plasticity occurs. In contrast the 'tip' of a simple diapir is its roof zone, and for hot Stokes diapirism the roof should preserve a ductile aureole with margin parallel foliations associated with large flattening strains, and fabrics indicating flow of material around the diapir (Marsh, 1982; Schmeling *et al.*, 1988). The 'tips' of fault-assisted intrusions presumably are roof zones of the growing chambers and/or positions where faults first intersect the lateral margins of ascending igneous bodies. For our purposes, we will simply note that fault models all require a fault along the margins of the pluton with at least as much displacement as the size of the pluton.

To our knowledge, tips of igneous bodies in migmatites have not been described in detail, presumably because of the complex and varied nature of melt lenses and the uncertain and often irregular direction(s) of magma transport (e.g. Brown *et al.*, 1995). However, recent studies of migmatites indicate that a variety of melt-migration processes operate including fracturing/diking, migration into and along local shear planes, and migration into a variety of low stress sites that form and move through rock by visco-elastic flow (Brown *et al.*, 1995; Collins and Sawyer, 1996). Thus, tips of melt bodies in migmatites may share characteristics of all previously described tips differing largely in scale.

After an introduction to the geology of Cascades Mountains, Washington, we describe sheets and sheet tips associated with the Entiat pluton and compare them to the above descriptions.

DESCRIPTION OF CASCADES SHEET TIPS

Introduction to Cascades core

The Cascades core is a fault-bounded tectonic domain (Fig. 1). The dominantly Eocene, dextral Straight Creek–Fraser River fault separates the core from low-grade Paleozoic and Mesozoic oceanic and arc-type rocks to the west. The Cretaceous and Paleogene Ross Lake fault zone separates the core from Mesozoic strata of the Methow basin to the east. The major internal structure of the Cascades is the brittle, Eocene Entiat fault, which divides the core into the Wenatchee block to the southwest and the Chelan block to the northeast. The sheet-like intrusive bodies discussed below are part of the Entiat pluton in the Chelan block (Figs 1 & 3).

The Chelan Mountains terrane makes up most of the Chelan block and consists of the Napeequa and Holden units (Figs 1 & 3). The Napeequa unit is composed of quartzite (metachert?), schist, and amphibolite (metabasalt), whereas the Holden is dominantly hornblende gneiss, hornblende–biotite schist, and amphibolite derived from intermediate volcanic and volcaniclastic rocks (Cater and Crowder, 1967; Cater and Wright, 1967; Tabor *et al.*, 1987a, b). The Holden unit has been dated radiometrically (U–Pb) as Late Triassic, whereas the Napeequa may be Mississippian to Jurassic based on correlations with unmetamorphosed strata. Both units, but particularly the Holden unit, have been intruded by numerous plutons now preserved as orthogneisses. In the southwest part of the Chelan block, the Napeequa unit overlies the Swakane terrane along a probable thrust contact of unknown vergence (Tabor *et al.*, 1987a, b; Hurlow, 1992). The latter terrane consists largely of Swakane Gneiss, mainly a biotite–plagioclase–quartz gneiss, which has a Precambrian provenance age (e.g. Mattinson, 1972).

The Holden, Napeequa, and Swakane units and associated orthogneisses are all intensely foliated with the foliation folded, usually coaxially, a minimum of twice around NW- to SE-trending, gently plunging axes. Each unit preserves early tight to isoclinal folds, which fold a foliation subparallel to lithological layering but have the dominant foliation parallel to their axial planes. These folds are refolded by gentle to tight upright, typically symmetric folds. Mineral lineations show a fair degree of scatter but also have NW–SE, gently plunging maxima (averages = 3/349 in Swakane, 4/333 in Napeequa, 2/347 in Holden) subparallel to fold axis maxima of the upright folds in each unit. Although folded, outside of narrow contact aureoles, the dominant foliation in the Chelan block tends to be moderate to gently NE-dipping with average dips of 21°NE in the Swakane, 38°NE in the Napeequa, and 8°NE in the Holden unit. All of these structures are associated with amphibolite facies mineral assemblages and calculated paleo-pressures of between 6 and 12 kbar (Tabor *et al.*, 1987a, b; Sawyko, 1994).

Entiat pluton

A series of highly elongate, NW–SE-striking plutons intrude these units (Fig. 1). One of these plutons, the 73 Ma Entiat pluton (Fig. 3), is >70 km long with a length to width ratio of >9. This pluton is largely tonalite with hornblende the dominant mafic phase. Trace element and REE chemistry have a subduction-related arc signature (Dawes, 1994). The presence of magmatic epidote in the pluton and calculated Al-in-hornblende barometric data (Fig. 3) indicate emplacement pressures between 6.0 and 7.2 kbar for the Entiat pluton and do not reveal any suggestion of tilting (Dawes, 1994).

The Entiat pluton consists of numerous sheets, centimeters to hundreds of meters wide, defined by compositional and textural differences (Cater and Crowder, 1967; Cater and Wright, 1967). Sheeting is particularly common in the northwest end and along the southwest margin of the pluton (Figs 3 & 4). Sheets typically dip steeply in the northwest end, moderately northeast along the southwest margin, and are locally folded in both regions by upright NW-trending magmatic folds. The numerous, thin sheets (meters to hundreds of meters) at the northwest end of the pluton give way to a few 0.5- to 2.5-km-wide sheets in the wider southeast portion of the pluton. However, these broad sheets display numerous schlieren layers and layers defined by modal and textural changes that may reflect more diffuse sheeting.

This pluton preserves weakly to strongly developed magmatic and locally subsolidus foliations and lineations (Cater, 1982; Tabor *et al.*, 1987a, b). Where magmatic foliations occur in the Entiat pluton they define complex patterns which overprint the sheets and internal layering. Foliation patterns define both meso- and macro-scale zones of upright, NW-trending magmatic folds and zones of margin parallel shear similar to and/or continuous with those in the nearby host rock.

The timing of emplacement of the Entiat pluton relative to regional deformation is critical to the discussion of sheet tips below. Regionally, the Entiat pluton intrudes across and thus postdates the thrust contacts between the Swakane, Napeequa, and Holden units and the early isoclinal folds in these units. However, the upright, NW–SE-trending folds are believed to be synchronous with emplacement for the following reasons: (1) magmatic folds of sheets and magmatic foliations with the same style and orientation as the upright, host rock folds occur throughout the pluton; (2) high-temperature (amphibolite facies) subsolidus fabrics occur along the margins of the Entiat pluton; (3) host rock schistosity in the limbs of upright folds in some places bends into parallelism with the pluton margin and in other places is continuous with magmatic foliation across the margin; (4) sills of the Entiat pluton in the host rock are folded by upright folds in some localities and in other places postdate these folds (e.g. Hurlow, 1992); and (5) K/Ar cooling ages from hornblende which define the main schistosity give a minimum age for this folding of between 67.1 and 71.5 Ma, only slightly younger than the 73 Ma age of the pluton (Tabor *et al.*, 1987b). Elsewhere, we have concluded that these folds formed

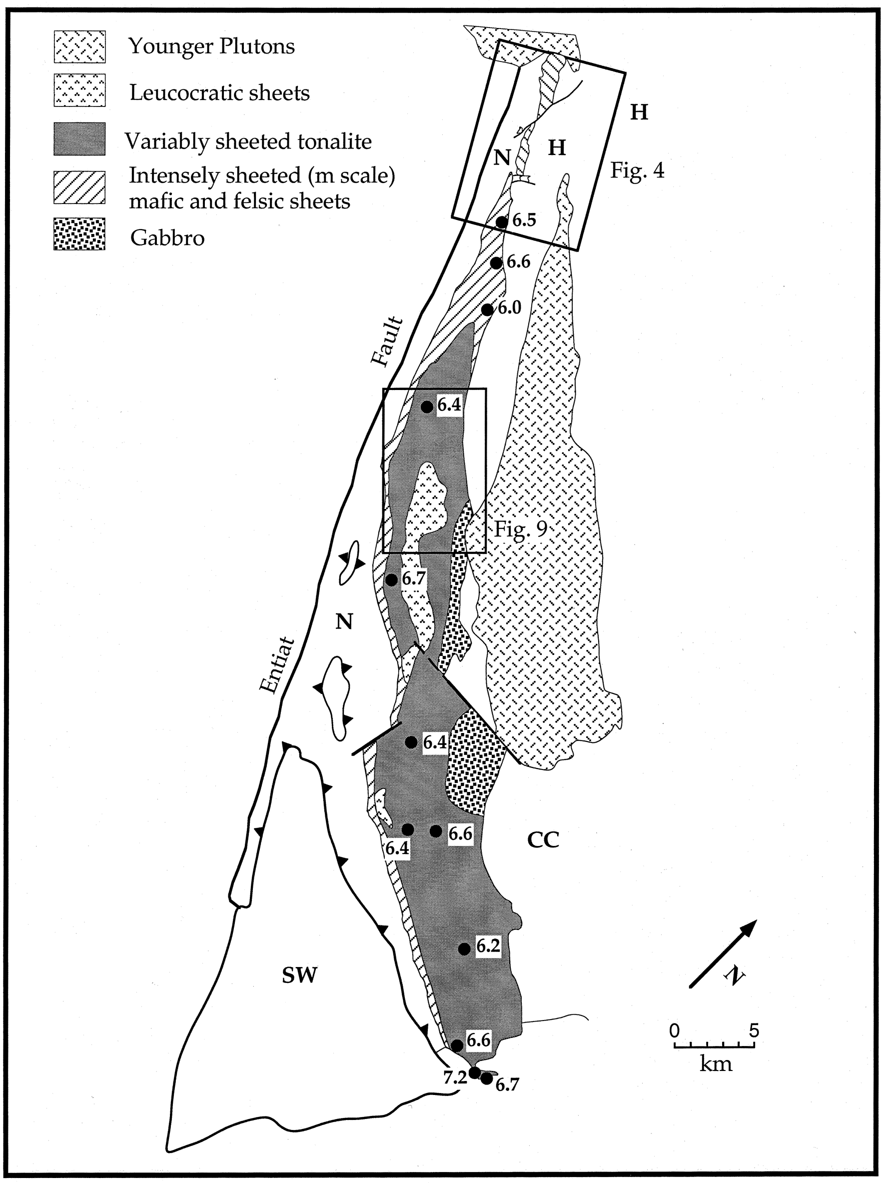

Fig. 3. Map of Entiat pluton showing the overall degree of sheeting, available Al-in-hornblende barometric data from Dawes (1994), and location of Figs 4 and 9. Note the strongly sheeted region at the northwest end, where elevations range between 1 and 2200 m, that is higher than elevations at the southeast end (from 250 to 1000 m). In the host rock: CC = Chelan Complex, H = Holden units, N = Napeequa, and SW = Swakane Gneiss.

during NE–SW contraction perpendicular to the pluton margin and during both subhorizontal and subvertical extension parallel to the pluton (Paterson and Miller, 1995). We return to this point after describing the sheets.

At the northwest end of the pluton, the main body feathers out into a series of narrow, elongate sheets, some of which intrude host rocks where they end in lateral edges or tips (Fig. 4). We have completed preliminary mapping of five of these sheet tips. One tip, the Marble Meadows tip (Fig. 3), occurs in the middle of the pluton where an $\sim 18 \times 3$ km, migmatitic, biotite alaskite to quartz–plagioclase–K-feldspar pegmatite sheet intrudes tonalites that make up the main part of the Entiat pluton (Tabor *et al.*, 1987b). The four remaining tips occur where sheets from the Entiat pluton intrude into and end in the surrounding host rock. The main characteristics of these five sheet tips are summarized in Table 1.

None of these sheets are parallel-sided bodies, nor are sheets and tips 100% exposed. But in order to better characterize each body, we have made estimates of three measurements: (1) a length to width ratio (L/W) measured in map view with the width representing an average measured in several places well away from its tip; (2) a radius of curvature at the tip (i.e. a measure of the tip bluntness); and (3) a tip diameter to sheet width ratio (D_T/W; i.e. a measure of how much taper occurs at the tip). Below we describe each sheet tip beginning with the sheets with the largest L/W ratio

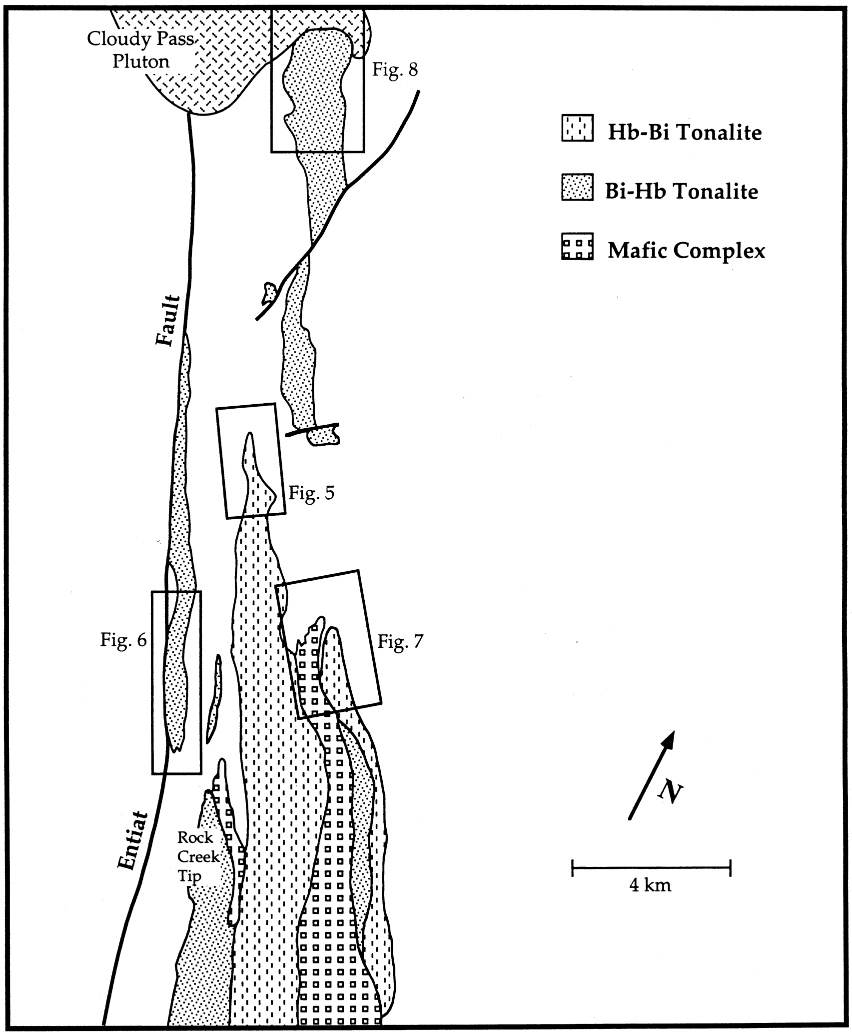

Fig. 4. Map showing sheeted nature of northwestern end of Entiat pluton and location of sheet tips discussed in this paper.

Table 1. Characteristics and structural measurements in sheet tip and host rocks. Note particularly the correlation between L/W ratios and D_T/W ratios. Also note the approximate parallelism between sheets and axial planes of host rock folds

Magma sheet characteristics							
Tip name	Orientation of sheet	Foliation (strike/dip)	Fold axis (Plunge/trend)	Mineral lineation	L/W	Tip radius (m)	D_T/W
Chipmunk Ck	331/85	Folded	9/292	ND	> 75	~100	~0.33
Old Gib	332/90	ND	ND	Shallow	~25	~100	~0.35
Pomas Pass	333/80	Folded	11/333	ND	> 14	~200	~0.50
Hart Lake	328/80	349/74	Not folded	Steep	> 8.0	~700	~0.64
Marble Mdw	321/85	Margin parallel	Not folded	Steep	~6.2	~850	~0.66

Host rock structural data			
Tip name	Fold axis (plunge/trend)	Axial plane (strike/dip)	Min. lineation (plunge/trend)
Chipmunk Ck	2/339	~339/70	ND
Old Gib	14/336	~336/85	15/348
Pomas Pass	2/338	~338/80	5/343
Hart Lake	17/148	~328/70	6/126
Marble Mdw	12/171	~345/80	Variable

(most sheet-like) and ending with the smallest L/W ratio.

Chipmunk Creek tip

At its northwest end, one of the main sheets in the Entiat pluton tapers in width from ~1.3 km to ~200 m over a distance of about 3 km, ending in a well exposed tip (Figs 4 & 5). This sheet has a minimum L/W ratio of 75, the uncertainty arising because the southeast end merges with other sheets in the main body of the Entiat pluton. The rather blunt tip has a radius of curvature of ~100 m and the sheet has a D_T/W ratio of ~0.33, that is the width at the tip is only one third of the average sheet width. The large L/W ratio and small D_T/W ratio make this the most 'dike like' of all the sheets examined (Table 1). Near its tip the sheet consists of hornblende–biotite tonalite. Some layering of and mingling between different phases is preserved in the tip. A few associated sheets of hornblendite intrude nearby host rock. Near the tip, the sheet strikes 331° and dips ~85°N. A well developed magmatic foliation occurs in the sheet and defines an antiformal structure, mimicking the orientation and style of the fold structure in the surrounding host rock (Fig. 5). Tight mesoscopic, upright, gently plunging folds of foliation also occur in the tip.

The host rock surrounding the tip is orthogneiss of unknown age called the Leroy Creek complex by Cater and Crowder (1967). Compositional layering in the orthogneiss is dominated by light colored, locally pegmatitic, biotite–quartz gneiss that grades into less

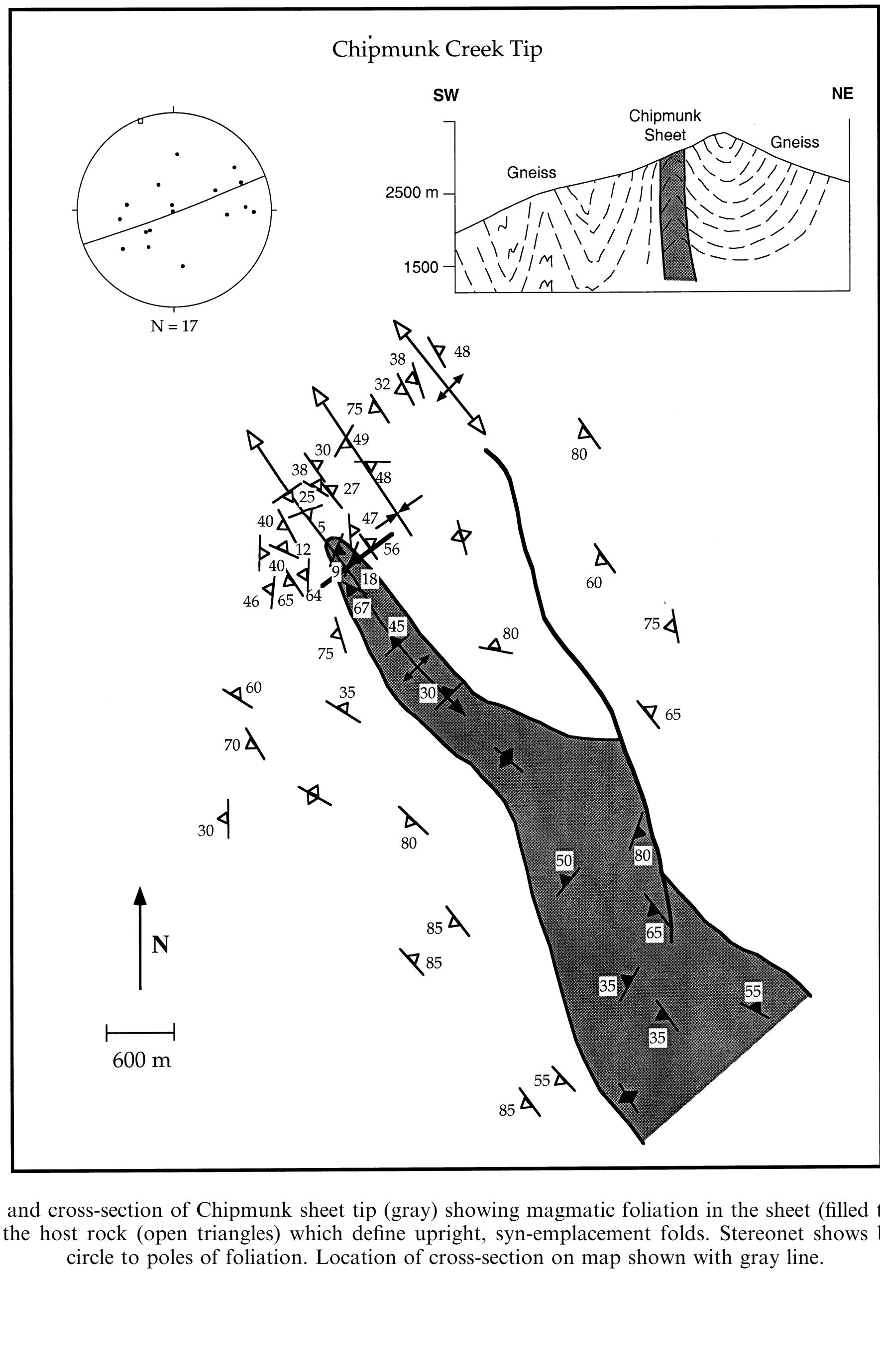

Fig. 5. Map and cross-section of Chipmunk sheet tip (gray) showing magmatic foliation in the sheet (filled triangles) and foliation in the host rock (open triangles) which define upright, syn-emplacement folds. Stereonet shows best fit, great circle to poles of foliation. Location of cross-section on map shown with gray line.

common hornblende gneiss. The gneisses are well foliated with the foliation subparallel to compositional layering. Fracturing and minor shearing during retrograde (chlorite zone) metamorphism overprints the above structures.

Both the high-temperature solid-state foliation and compositional layering in the Leroy Creek orthogneisses are folded, defining a slightly SW-vergent antiform with a steep axial plane (339/70) and fold axis that plunges 2° towards 339°. The sheet tip occurs in the hinge of this fold with the margins of the sheet oriented roughly parallel to the axial plane of the fold. We did not find any faults extending from the tip nor along the sides of the sheet, nor evidence of an increased amount of fracturing near the tip. The only new axial planar fabric is a sporadically developed weak fracture cleavage in leucocratic layers and an even less common crenulation cleavage in some biotite-rich zones. Since no penetrative axial planar fabric exists, the steeply dipping margins of the sheet must intrude across host rock anisotropy, that is the subhorizontal layering in the hinge of the fold. However, immediately adjacent to the sheet margins, the limbs of the fold locally steepen so that they become margin parallel, suggesting that intrusion caused some deflection during ductile flow of the host rock.

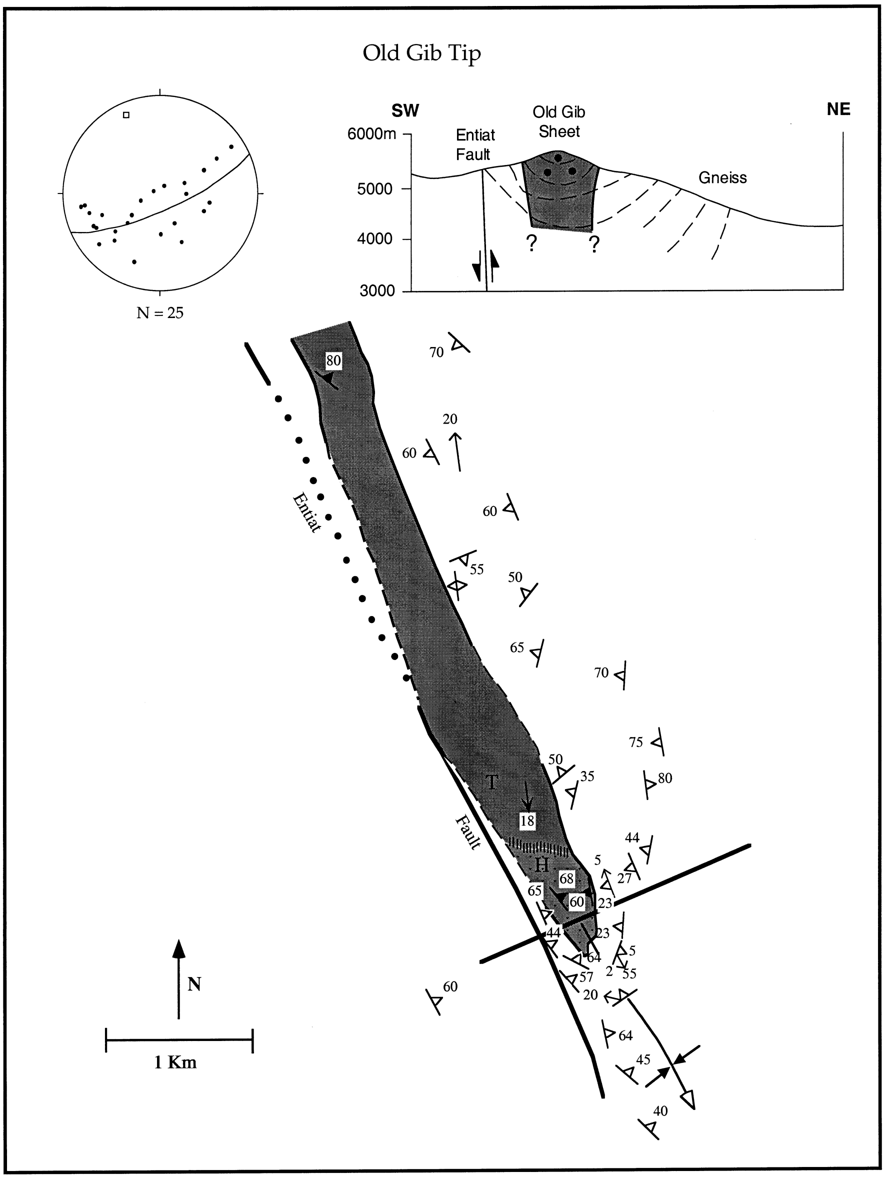

Fig. 6. Map and cross-section of Old Gib sheet tip (gray) showing magmatic foliation in the sheet (filled triangles) and foliation in the host rock (open triangles) which define an upright, syn-emplacement fold. In the sheet: H = hornblendite and T = tonalite. Short arrows = mineral lineations (in cross-section shown as black dots since subhorizontal), long arrows = fold axes. Stereonet shows best fit, great circle to poles of foliation. Location of cross-section shown on map with gray line.

Old Gib tip

Southwest of the main body of the Entiat pluton, an isolated sheet of largely biotite–hornblende tonalite is exposed just northeast of the Eocene Entiat fault near Old Gib Mountain (Fig. 6). Coarse-grained hornblendite and local fine-grained tonalite to diorite occur at the tip. This roughly parallel-sided sheet has a L/W ratio of ~25, a tip radius of ~100 m, and a D_T/W ratio of ~0.35. The sheet strikes 332° and dips ~90°.

Much of the tip region consists of coarse hornblendite with no obvious foliation or lineation. However, magmatic foliation and lineation, defined by igneous hornblende, plagioclase and mafic clots occur in the finer grained diorite and tonalite. The foliation defines a tight, upright, gently plunging synform which has a half-wavelength comparable to the tip width. This fold again mimics the orientation, style, and wavelength of the folds in the host rock (Fig. 6). The magmatic mineral lineation is subhorizontal and trends approximately parallel to the fold axis.

The host rock consists of quartz–plagioclase ± hornblende ± biotite gneisses of uncertain protolith and amphibolites that are probably part of the Napeequa unit. Local marble and ultramafic slices also occur. All rock types are strongly foliated and locally preserve a mineral lineation. The foliation and lithological contacts define a well exposed synform with an axial plane that strikes 336° and dips steeply. The fold axis plunges 14° towards 336° (Fig. 6 & Table 1). The sheet tip is located in the hinge of this fold with the sheet margins roughly parallel to the axial plane of the fold. During examination of the host rock, we did not find any new axial planar cleavage, faults extending from the tip nor along the sides of the sheet, nor evidence of an increased amount of fracturing near the tip. Furthermore, relatively flat-lying foliation and lithological contacts in the well exposed hinge of the fold occur immediately in front of the steeply dipping sheet tip, requiring that this sheet also intruded discordantly across host rock structures.

Pomas Pass tip

Two intermediate-sized Entiat sheets end in tips along the northeastern margin of the pluton near a region called Pomas Pass (Figs 4 & 7). We have briefly examined the more northeasterly tip (Jbq on Fig. 7). This sheet has a moderately nonplanar margin and a minimum L/W ratio of 14, the uncertainty again arising due to the southeastern end merging with other sheets. The tip radius is ~200 m and D_T/W ratio is ~0.50. The sheet strikes 333° and dips ~80°NE. Hornblende–biotite tonalite and quartz diorite are the most common rock types at the tip although there is widespread evidence of mingling which locally results in a chaotic mafic complex. Hornblendite, gabbro, and diorite are present in the mafic complex as well as in dikes in the host rock. These units preserve strongly developed banding and magmatic foliation, which are locally folded into symmetric to asymmetric upright, gently plunging folds. Foliation tends to dip moderately to the northeast (Fig. 7), except in the hinges of folds where foliation and layering dip gently.

Here, the host rock is a strongly deformed Triassic Dumbell orthogneiss (Cater and Crowder, 1967). The orthogneiss is strongly layered at the meter scale and typically consists of medium- to coarse-grained quartz–plagioclase gneisses with variable amounts of hornblende and biotite. The orthogneiss preserves evidence of both magmatic and high temperature subsolidus foliation and lineation. Foliation is folded into meter-scale folds which are parasitic on folds with wavelengths of hundreds of meters. Folds are usually gentle to open, upright (average axial plane strikes 333° and dips ~80°NE), gently plunging (average plunges 2° towards 338°), and slightly asymmetric, verging to the southwest. Mineral lineation plunges 5° towards 343°, approximately parallel to fold axes.

The width of this sheet is larger than the typical wavelength of host rock folds and thus does not show the one-to-one relationship between folds and sheets that the two previous examples do. Again, however, the sheet margin is oriented parallel to the axial planes of these folds. Particularly in front of the sheet tip, host rock foliation commonly dips much more gently than the sheet margins emphasizing that the sheet is highly discordant to the host rock anisotropy. As with the other tips, we did not find any new penetrative axial planar cleavage, faults extending from the tip or along the sides of the sheet, or evidence of an increased amount of fracturing near the tip.

Hart Lake tip

The most northwesterly exposed tip marks the termination of the main part of the Entiat pluton (Figs 4 & 8). Host rock immediately northwest and north of the tip region is largely removed by emplacement of the 29 Ma Cloudy Pass pluton (Cater and Crowder, 1967). Our studies indicate that this shallow level, highly discordant pluton stoped out host rock, but did not otherwise alter the characteristics of the Entiat tip or nearby host rock. Because this sheet tip has several interesting characteristics different from previous sheets, we include it in this discussion. This sheet has irregular margins but on average strikes 328°, dips ~80° to the northeast, and has a minimum L/W ratio of 8. Because of its truncation by the Cloudy Pass pluton, the tip radius is uncertain, but is ~700 m. The D_T/W ratio is ~0.64. The tip consists of biotite–hornblende tonalite, with rare hornblendite inclusions, and shows some mingling with slightly more mafic phases. Compositional layering is well developed in places and is parallel to a generally moderately developed foliation that tends to strike parallel to the sheet and dips

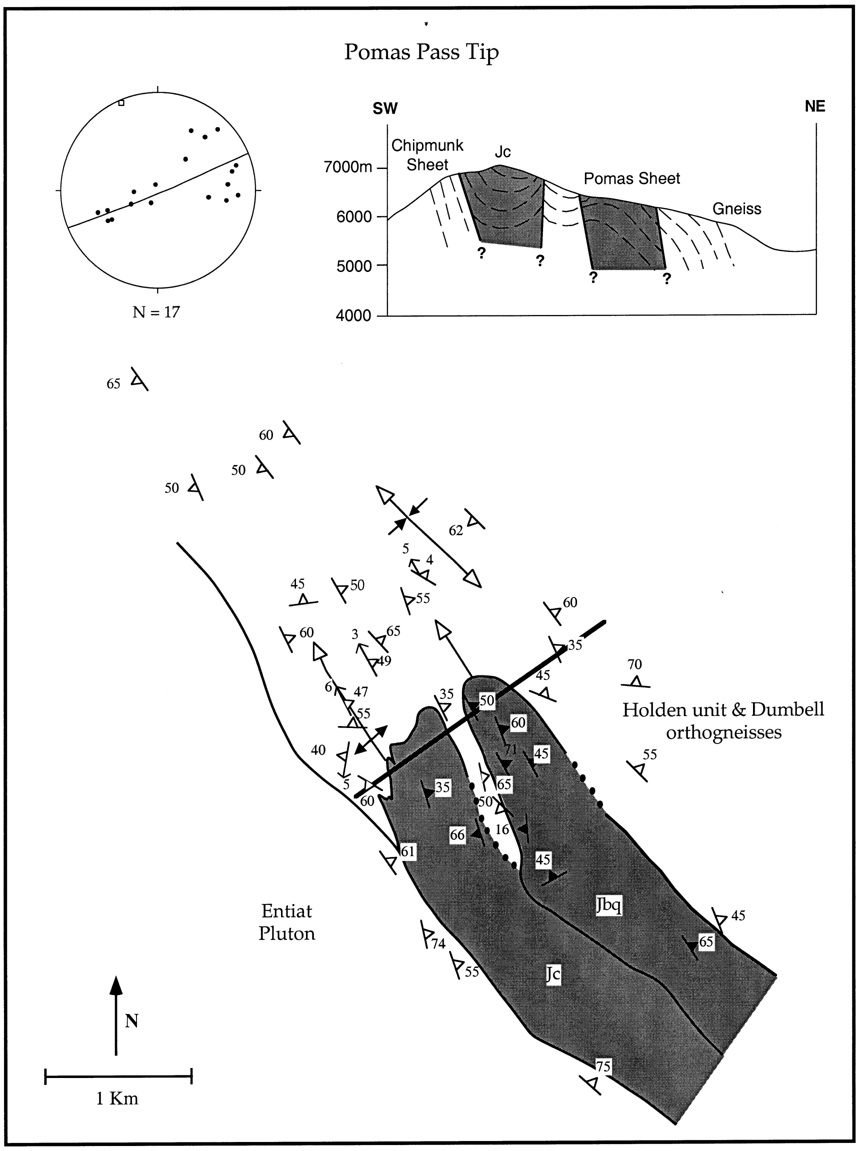

Fig. 7. Map and cross-section of Pomas Pass sheet tips (two shades of gray) showing magmatic foliation in the sheets (filled triangles) and foliation in the host rock (open triangles) which define upright, syn-emplacement folds. Short arrows on foliation = mineral lineations, long arrows = fold axes. Stereonet shows best fit, great circle to poles of foliation. Location of cross-section shown on map with gray line. 'Entiat pluton' = southeast part of Chipmunk sheet.

60–90°NE (Fig. 8). Foliation remains parallel to the sheet walls in the tip region, but unfortunately cannot be observed immediately at the tip. No folding of the foliation on the outcrop scale or on the scale of the sheet was observed. A hornblende-defined magmatic lineation is prominent, particularly near the tip termination, and some areas display constrictional shape fabrics. Unlike lineations in most of the Entiat pluton, the linear fabric has 60–90° plunges.

The host rock consists of the Dumbell orthogneiss similar to that near the Pomas Pass tip. This orthogneiss contains a well developed high-temperature foliation and moderately strong lineation, typically overprinted by chlorite-grade shearing. Foliation and layering are generally subparallel and immediately northeast of the tip region dip gently to moderately. However, both are folded by upright, relatively symmetric folds with gentle NW–SE plunges. Mineral lineation also has subhorizontal plunges, but shows some scatter in the foliation plane. Average dip in the limbs of the folds tends to steepen along the sheet margins (Fig. 8). These folds generally have much smaller wavelengths than the sheet width. Thus, although the sheet margins strike roughly parallel to the axial planes of these folds, the sheet intrudes across many folds in a manner more typical of the Pomas Pass tip than the other tips.

Marble Meadows tip

Southeast of where all the above sheets merge into the main body of the Entiat pluton, a separate more

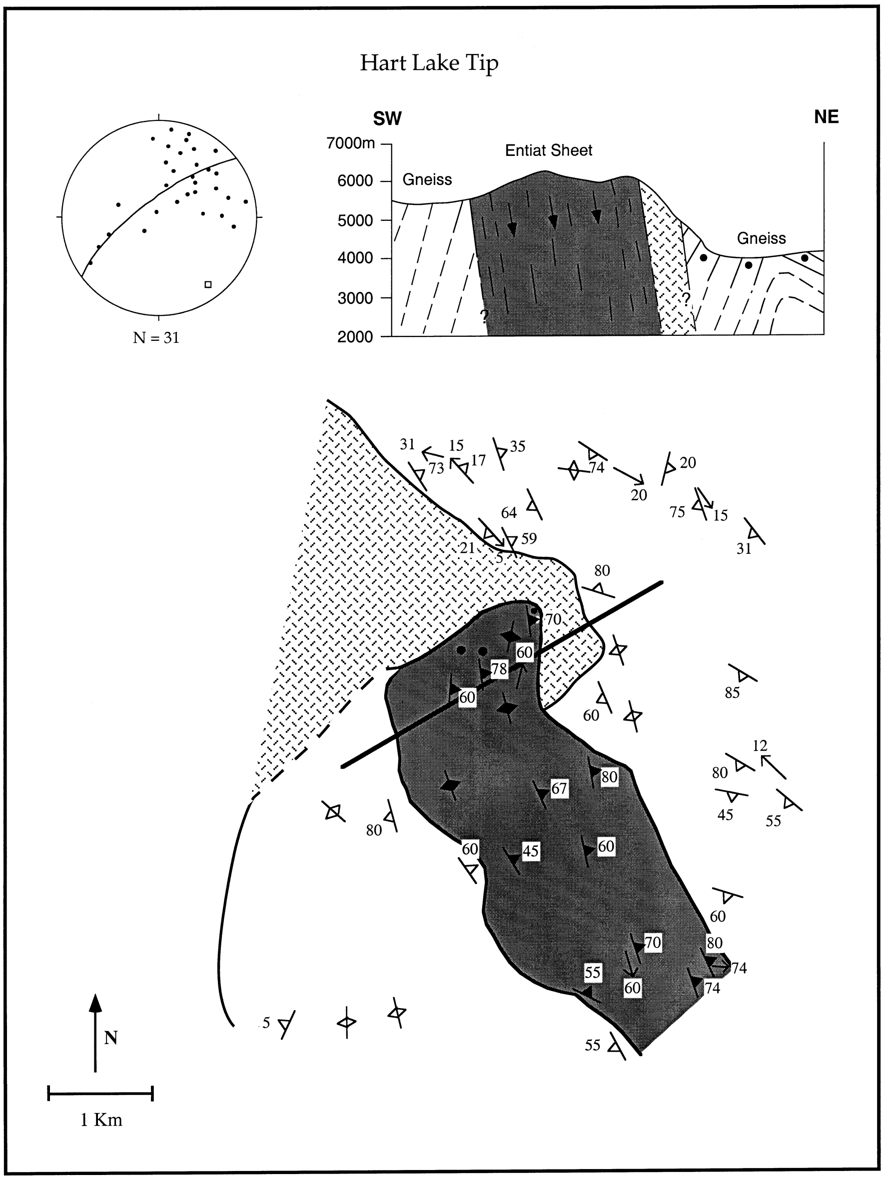

Fig. 8. Map and cross-section of Hart Lake sheet (shaded gray) showing magmatic foliation in the sheet (filled triangles) and foliation in the host rock (open triangles) which define upright, syn-emplacement folds. Cross hatched pattern = Tertiary Cloudy Pass pluton. Short arrows (subhorizontal plunges) or black dots (steep plunges) = mineral lineations (in cross-section shown as black dots where subhorizontal or arrows where steep plunges), long arrows = fold axes. Only largest wavelength fold shown on the cross-section. Many smaller wavelength folds exist. Stereonet shows best fit, great circle to poles of foliation. Location of cross-section shown on the map with gray line. Note that this broader tip cuts across host rock folds and does not preserve folded magmatic foliation in the tip, although small-scale magmatic folds are preserved in this sheet well away from the tip region.

leucocratic body intrudes the pluton (Figs 4 & 9). This sheet has a L/W ratio of ~6.2, the smallest of any of the sheets examined. It strikes 321° and dips ~85°, although in detail margins are curviplanar (this is accentuated in Fig. 9 due to topographic effects along the northeast margin). Both tips of this sheet are exposed, but we have only examined the tip at the northwest end at Marble Meadows. This tip has a radius of ~850 m and a D_T/W ratio of ~0.66, and is also the largest of all the sheets examined. Much of the sheet consists of heterogeneous light colored gneiss ranging from fine-grained alaskite with local pegmatite to medium-grained tonalite (Tabor *et al.*, 1987a, b; Dawes, 1994). Magmatic foliation is moderately to weakly developed and typically margin parallel. Of particular interest is that it parallels the sheet margin at the tip, rather than forming a fold pattern such as those in the sheets with greater L/W ratios.

Host rock consists of the main body of the Entiat pluton, here mostly medium- to coarse-grained hornblende–biotite tonalite. Magmatic foliation and a less common mineral lineation are defined by igneous plagioclase and hornblende. Locally the foliation is folded into upright folds, typically with half-meter wave-

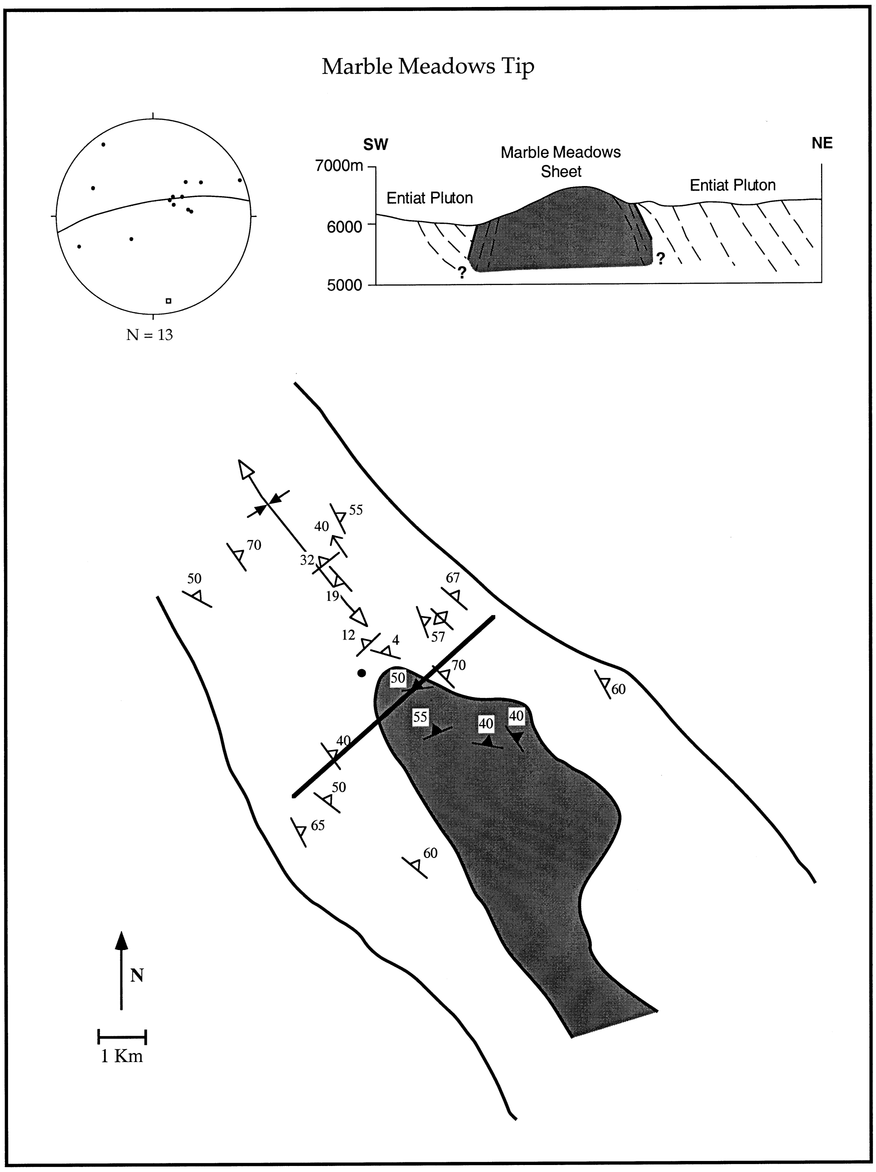

Fig. 9. Map and cross-section of Marble Meadows sheet tip (shaded gray) showing magmatic foliation in the sheet (filled triangles) and foliation in the host rock (open triangles) which define upright, syn-emplacement folds. Irregular margin along the northeast edge of sheet is in part due to topography. Margin parallel magmatic foliations also occur along the southwest margin of sheet, but are not well exposed at sheet tip. Host rock = main body (largely tonalite) of the Entiat pluton. Black dot = steeply plunging mineral lineation, long arrows = fold axes. Stereonet shows best fit, great circle to poles of foliation. Location of the cross-section shown with gray line on map.

lengths. A regional-scale fold also occurs immediately northwest of the tip (Fig. 9). This fold is an open, strongly asymmetric *S*-fold verging to the southwest. It differs from other folds near tips in that it has moderately NE-dipping limbs and flat-lying foliation in the hinge. No evidence of subsolidus deformation was observed in the pluton in this folded region. Magmatic mineral lineation in the region away from the tip usually has gentle to moderate plunges: immediately at the tip, magmatic lineation changes abruptly to steep plunges.

This end of the Marble Meadows sheet intrudes along the flat-lying limb of the *S*-fold with sheet margins approximately parallel to the axial planes of the small-scale magmatic folds but steeper than the moderately NE-dipping magmatic foliation in the limbs of the *S*-fold.

DISCUSSION

Summary of sheet tip characteristics

These intrusions are all roughly parallel-sided bodies with relatively blunt tips and L/W ratios between what is typically associated with dikes ($L/W > 1000$) and with diapirs ($L/W < 3$). They largely consist of tonalite (or derivatives in Marble Meadows), commonly with evidence of mingling between tonalite and slightly more mafic phases in the tip region. The only obvious petrological differences near tips are a decrease in grain size from the main mass of the Entiat pluton and, particularly in the Old Gib tip, an increase in hornblendite. Beyond these general features, the characteristics of the sheets summarized in Table 1 indicate that a well-defined transition is represented by these bodies. Decreasing L/W ratios correlate well with increasing tip radii and increasing D_T/W ratios indicating a transition from thin sheets with narrow, blunt tips towards more elliptical bodies. Furthermore, sheets with large L/W ratios show a more intimate relationship to host rock folds and have magmatic foliation and lineation patterns continuous with host rock patterns. Sheets with low L/W ratios intrude across multiple folds and have magmatic foliation and lineation patterns that differ from host rock patterns.

In all cases, foliation patterns in host rock at, and immediately in front of, sheet tips define upright, gently plunging folds with no pervasive axial planar cleavages. Some folds are synformal and some antiformal. Some have half wavelengths approximating the width of the sheets, whereas others have much smaller wavelengths. Sheet margins are consistently parallel to the axial planes of these folds, cut across host rock anisotropy in the fold hinges, and are not folded, at least at the scale of our mapping. In no case did we find faults near tips or along sheet margins, nor did we find obvious zones of increased fracturing or ductile shearing comparable to process zones seen near dike tips at shallow crustal levels.

Similarities of host rock and magma fold geometries, changes near the tips, and the timing constraints listed earlier in this paper indicate that for each sheet the duration of folding and sheet emplacement significantly overlapped. In particular, the presence of the folds of magmatic foliation in the sheets requires that sheets of melt existed parallel to axial planes during the folding process. Our regional studies indicate that these folds and the subhorizontal mineral lineation formed during arc-wide NE–SW contraction, vertical thickening, and arc-parallel, subhorizontal extension (Paterson and Miller, 1995). Therefore, the regional strain field ($S_1 \geq S_2 \geq S_3$; where S = principal stretch) during development of these sheets has S_3 ~horizontal and perpendicular to the arc, S_2 or S_1 arc-parallel and approximately horizontal, and S_1 or S_2 approximately vertical (uncertainty in the orientation of S_1 and S_2 arises because the magnitude of thickening vs arc parallel extension is unknown). This implies a regional stress field in which σ_1 is approximately horizontal and at high angles to the sheet margins, and σ_2 and σ_3 approximately parallel to the sheets.

Are these really sheets with tips?

The steep dips of sheets listed in Table 1 were determined from three-point calculations where topographic relief occurred near sheet margins. These calculated dips agree well with dips of sheet margins measured in the field. Enough topographic relief occurs along all of the sheets (except the Marble Meadows) to convince us that vertical dimensions of these sheets are greater than their widths. This conclusion is supported as well by a regional examination of the contacts of the Entiat pluton, which typically dip steeply northeast over a vertical distance of > 1.5 km. However, without knowing the displacement directions of magma during ascent and growth of a sheet, it is impossible to determine whether the end of each sheet is a tip or a lateral edge. For several reasons we suggest that this is an unimportant distinction. First, studies of the growth of and flow within sheet-like bodies conclude that growth/flow occurs in all directions within the plane of the sheet. Thus, even lateral edges represent propagating tips (e.g. Lister and Kerr, 1991). Second, several of the sheets show changes near the tips consistent with the tips being former propagating margins. For example, the hornblendite in the Old Gib tip and hornblendites near the Pomas and Chipmunk Creek tips are consistent with these being H_2O-rich zones near tips, such as those proposed for propagating dike tips. The unusual down-dip lineations at the Marble Meadows and Hart Lake tips also support the conclusion that these may be the growing edges of sheets. In any case, we note that the three-dimensional geometrical relationships between sheets and host rock structure do not change whether the ends of these bodies are viewed as tips or lateral edges.

The above statements are valid if these bodies are viewed as dikes, but potentially less meaningful if viewed as diapirs. For example, Buddington (1959) and Paterson *et al.* (1996) have shown that the mechanisms by which diapirs ascend may be different at their tops and sides. However, even if these are diapirs, Hand and Dirks (1992) noted that magma bodies with their margins at high angles to σ_1 will flatten and thus grow in the σ_2–σ_3 plane, implying that the Entiat sheet tips would represent former propagating edges. Thus, although we recognize that the ascent of the Cascade sheets may involve additional processes not preserved at the examined tips, we believe that these tips represent former migrating margins and that their characteristics have important implications for how the sheets grew/ascended.

Do the sheets provide information about emplacement or ascent?

To a large degree, making a distinction between magma ascent and emplacement is misleading, since both require simultaneous movement of melt and displacement of host rock. However, all igneous bodies may change their characteristics through time and thus may not faithfully record magma and host rock processes that operated during an earlier part of their history. The degree to which they do retain information about ascent will depend on whether the igneous body runs out of thermal energy and freezes faster than the magma–host rock system can change its characteristics during final emplacement. This is more likely in sheet-like bodies, such as dikes or those examined in this study, because of their faster cooling rates. Furthermore, the operation of *all* of the ascent mechanisms listed in Fig. 2 have been defended by others through comparison of theoretical and experimental predictions to igneous bodies preserved in the rock record. We therefore feel justified in arguing that these sheets record information about magma ascent at the presently exposed crustal levels (20–24 km), but recognize that the type(s) of mechanisms possibly changed with time or with depth. Most importantly, we note that the relationships described above and conclusions reached below are just as intriguing whether viewed as relating to ascent or emplacement.

Likely ascent mechanisms

Of the four ascent mechanisms presented in Fig. 2, faulting is the easiest to rule out for the Cascade sheets. None of these sheets have faults at their tips or along their margins. Furthermore, the lack of any visible offsets of the widespread host rock markers, from small-scale layering to large-scale lithological contacts, rules out any cryptic faults potentially missed at sheet tips. This in itself is an interesting conclusion because of the close connection often assumed between sheeted plutons and faulting (Hutton, 1988; D'Lemos *et al.*, 1992; Paterson and Fowler, 1993). The Entiat pluton provides an excellent example where an elongate, internally sheeted body did not form along an active fault.

The Entiat sheets share one common characteristic with migmatites in that magma movement is intimately related to regional deformation. However, the scale, consistent orientation, and centralized location of sheets, and thus focussed flow represented by these sheets, distinguish them from the migmatite end-member represented in Fig. 2.

At first glance the sheet-like shape of these bodies makes a dike ascent mechanism attractive. However, most of their other characteristics argue against this. L/W ratios are much smaller than typical dikes. In detail, margins are curviplanar to a degree not characteristic of dikes. Mingling of different magmas within the sheets, evidence of ductile flow in host rock along margins, and magmatic foliation patterns are also not typical of dikes, nor is the lack of process zones at their tips. Furthermore, preserved sheet thicknesses would require unrealistically large flow velocities if they represented dikes. The strongest deterrent to concluding that they are dikes is their emplacement at high angles to σ_1 rather than to σ_3 as is predicted by dike models (Lister and Kerr, 1991). Lucas and St-Onge (1995) noted that if deviatoric stresses are small, such as may occur in the mid and lower crust, dikes can intrude parallel to host rock anisotropy rather than perpendicular to σ_3. But the Cascade sheets intrude across anisotropy, precluding this explanation. It also seems unlikely that sheet orientations are caused by a local switch in σ_1 and σ_3 due to folding or a regional switch due to tectonic processes. The former is ruled out by the examples of sheets that intrude across many folds. The latter explanation would require some process by which hundreds of flips in stress occurred over a time-scale of several million years, a hypothesis for which there is no known mechanism, no supporting data in the Cascades, and which is difficult to reconcile with our evidence of NE–SW contraction before, during, and after emplacement of the Entiat pluton.

We suggest that the following scenario may explain the formation of the sheets (Fig. 1). While the Cascades arc was undergoing arc-perpendicular contraction, an elongate body of magma with its long dimension oriented parallel to the arc began rising diapirically through a heterogeneous lower crust. We imagine that sheet-like, arc-parallel diapiric ridges of magma leave the main body of magma and grow vertically and laterally. In the case of the Entiat pluton, the above mechanisms must allow these ridge-like instabilities to grow vertically for hundreds of meters or more away from the main chamber with ridge widths ranging up to 0.5 km. At some point sheets begin to coalesce and/or the main magma chamber rises to a higher level engulfing the earlier formed sheets. Thinner sheets with large L/W ratios must have ceased ascent long enough ago to have formed magmatic fabrics subparallel to host rock fabrics and then be folded before final freezing (note that actual crystallization rates of these sheets are probably not rapid because of the high ambient temperatures of $>550°C$). Tips of the larger, more elliptical bodies were not as directly affected by this folding. Thus these sheet tips either crystallized while still ascending, and so preserving the margin parallel foliations and/or steep lineations (e.g. Hart Lake and Marble Meadow tips), or the tip margins acted as strain discontinuities during regional deformation causing magmatic foliation and lineation to form parallel to margins. The blunter tips and more elliptical shapes of the latter sheets may reflect three processes: (1) the tendency for flow instabilities to form finger-like shapes (see discussion below); (2) an attempt to reduce mechanical drag and loss of thermal energy (Marsh, 1982); and (3) at least for the Marble Meadows sheet, ascent through an existing magma chamber in which 'host rock' properties were more isotropic and stresses were approximately hydrostatic. In this scenario, the present surface exposure of the Entiat pluton preserves an excellent example of the transition from well defined individual sheets at higher elevations between ~1 and 2000 m (the northwest end), to coalesced sheets and a continuous chamber at elevations below ~1000 m (the southeast end).

If the above scenario is correct, explanations are needed for why sheet-like pulses of magma leave the main chamber and why these pulses maintain their sheet-like shapes. Flow instabilities in diapirically rising masses, including isothermal Rayleigh–Taylor instabilities (e.g. Ramberg, 1981; Olsen and Singer, 1985), instabilities due to initial variability in source regions (Bruce and Huppert, 1989), Saffman–Taylor instabilities due to variability in hydraulic resistance, and instabilities driven by thermally controlled fluid rheology (e.g. Whitehead and Helfrich, 1991) have been confirmed in fluid dynamics experiments. For example, Pollard *et al.* (1975), referring to experiments by Saffman and Taylor (1958), discuss how magma fingers can form due to instabilities along an interface between two viscous fluids where movement of the fluids is perpendicular to the interface (Fig. 10). They concluded that if magma viscosity is less than host rock viscosity such an interface is unstable and local perturbations will develop. These perturbations will grow into fingers if they experience less hydraulic resistance to growth than points along the interface lagging behind (see also Whitehead and Helfrich, 1991). Subsequent studies by Talbot *et al.* (1991) and Whitehead and Helfrich (1991) among others have confirmed this behavior.

We note that although Pollard *et al.*'s (1975) discussion focussed on explaining the growth of shallow sills, the above experiments are more applicable to the diapiric rise of sheets during viscous flow of host rock in middle and lower crust than to fracturing and diking in the upper crust. Where Pollard *et al.* (1975) had a difficult time explaining why their initial instabilities grew into magma fingers, our observations indicate that there are several reasons why instabilities continued to grow in the Cascades: (1) thermal instabilities (Whitehead and Helfrich, 1991) or the existence of pre-heated magma pathways (Talbot, 1977; Marsh, 1982) will focus additional melt migration (supported by mingling in tips) towards propagating edges and thus more quickly heat the surrounding host rock around the growing protrusions; (2) lower plastic yield strengths or strain-rate softening of host rock near sheet tips may focus further growth/ascent in these softened zones (Kirby, 1985; Weinberg and Podladchikov, 1994); and (3) rising pulses along fold hinges may be favored because of favorable stress gradients in these regions (e.g. Biot, 1961; Price and Cosgrove, 1990).

Fluid dynamics modeling generally shows that flow instabilities result in finger-like, rather than sheet-like protrusions (e.g. Talbot *et al.*, 1991; Whitehead and Helfrich, 1991) except in three cases: (1) where pre-existing rigid boundaries exist (Talbot, 1977; Rönnlund, 1987; Talbot *et al.*, 1991); (2) faults, facies changes, topography, or other lateral changes in material properties exist (Talbot, 1977; Rönnlund, 1987; Talbot *et al.*, 1991); or (3) if regional deformation is occurring during rise of materials (Koyi, 1988; Talbot *et al.*, 1991). All three of these cases are applicable to the Entiat sheets and, therefore, the Entiat sheets probably represent diapiric instabilities in which ascent is not controlled by growth along extensional fractures. Thus their orientations need not be controlled by σ_3 and instead may reflect a variety of processes (Rubin, 1993). Below we discuss several controls on the shapes and orientations of such magma bodies formed in contractional orogens.

1. Pluton shapes and orientations may be a residual effect of the melt generation site. In arcs in which units with similar bulk compositions are commonly continuous for long distances along the arc, con-

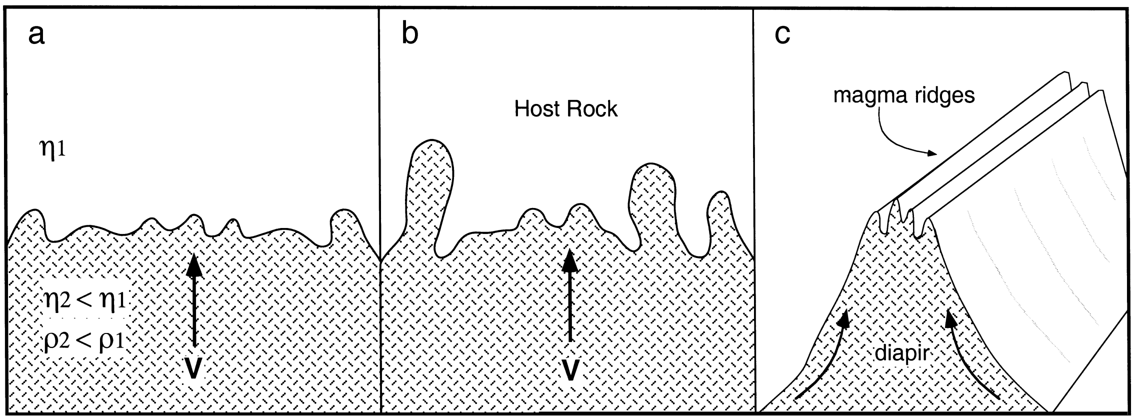

Fig. 10. Cartoon showing proposed model of initiation and diapiric rise of magma 'ridges' (a) and (b) reproduced from fluid dynamics experiment of Saffman and Taylor (1958) showing the growth of fingers along an unstable interface between two fluids with different viscosity. V = displacement direction in fluids. (c) Magma 'ridges' forming perpendicular to S_1 and rising off a diapir of unknown shape, but generally elongate parallel to the arc.

ditions for magma generation will be attained at approximately the same time in linear belts parallel to the arc. If so, arc-parallel ridges of melt may begin to ascend at approximately the same time as has been suggested for some salt walls (Jackson and Talbot, 1994). The migration of the position of arc-parallel belts of plutons of similar age, in some cases hundreds of kilometers in length, supports this contention (e.g. Bateman, 1992). Miller *et al.* (1989) note that such migration of linear magmatic belts did occur in the Cascades.

2. As flow instabilities begin to form protrusions above elongate diapirs, the existence of rigid boundaries and other crustal heterogeneities, along with actively contracting host rock, will preserve existing sheet-like shapes or force new protrusions into sheet-like shapes (Talbot *et al.*, 1991). We emphasize that early formed sheets will form vertical walls that potentially focus later rise of magma.
3. Plastic yield strength anisotropy of the host rock may play an important role in controlling pluton shapes and orientations. A heterogeneous crust will have some units with compositionally or thermally controlled lower yield strengths (e.g. Kirby, 1985). Magma bodies rising by any mechanism involving flow of host rock (e.g. Weinberg and Podladchikov, 1994) will change shape in response to their ability to rise faster and farther in these 'weak' units in preference to adjacent stronger units. If these weak lithological units strike parallel to the arc, then rising magma bodies will tend to become elongate parallel to the arc. The heterogeneous crust in the Cascades and overall trend of lithological units parallel to the arc are consistent with this hypothesis at the batholithic scale. But the common flat-lying geometries of crustal layering near the Entiat pluton make it unlikely that this is an important effect for the earliest sheets in the Entiat pluton.
4. Hand and Dirks (1992) note that a stress gradient must form in host rock along the margins of magma bodies because these low viscosity bodies cannot support regional deviatoric stresses. If the host rock is capable of flowing, this stress gradient will drive a change in shape of the magma body towards sheet-like shapes with the sheet oriented parallel to the σ_2–σ_3 plane, that is into the shape and orientation represented by the Cascade sheets. This process has been successfully modeled by Grujic and Mancktelow (1998).

The actual means by which the propagating magma ridges displace host rock include tightening of already active folds, ductile flow along sheet margins, and stoping (supported by stoped blocks within sheets). The ductile deformation includes both lateral and downward flow since host rock markers are sometimes intensely folded and sometimes deflected downwards near sheet margins.

The above scenario predicts that the sheeted region of the Entiat pluton represents the upper portion of the pluton instead of the root or feeder zone as is sometimes proposed (Hopson and Dellinger, 1987). The latter interpretation does not agree with the higher elevations of these sheets, their abrupt termination in fold hinges, their relatively blunt tips, and the lack of evidence of host rock melting at sheet terminations. Our proposed ascent scenario also predicts that the blunter, more elliptically shaped, bodies would typically be slightly younger than the high L/W sheets.

CONCLUSIONS

Although diking, diapirism, ascent along faults, and ascent during heterogeneous deformation are championed as 'the ascent mechanism' for magmatism in the crust, we believe that these mechanisms form end-members in a continuum. In Fig. 2, we have emphasized the natural transitions between these end-members in both host rock and magma behavior. We therefore believe that one goal of future magma ascent studies should be to examine the relative importance of and interplay among these mechanisms. Looking at the tips of propagating margins of igneous bodies is one useful way of doing this. For example, the characteristics of mid-crustal sheets and sheet tips for the Entiat pluton indicate that these bodies have geometries falling between those typical of dikes and elliptical diapirs, are not associated with faults or fracture zones extending from their tips, and instead are oriented parallel to the axial planes of open, upright, syn-emplacement folds. These observations rule out fault and simple elastic dike models and instead favor diapiric rise of magma sheets during complex, visco-elastic behavior of host rock. We suggest that the final L/W ratio and internal characteristics of these bodies reflect an interplay between regional deformation of a rheologically complex crust and ascent of sheet-shaped diapirs, and therefore have transitional characteristics between migmatites and 'visco-elastic' diapirs (right edge of Fig. 2).

This study of sheets in the Entiat pluton has several other important implications. The most convincing is that elongate, internally sheeted plutons can form during arc-perpendicular contraction without forming in local zones of extension (e.g. Hutton, 1988) or during regional arc-perpendicular extension. The consistent orientation of sheets parallel to the axial planes of active folds and perpendicular to the regional contraction direction also implies that sheet-like bodies may form perpendicular to σ_1 rather than σ_3 as is typically assumed for dikes.

If our ascent model is correct, it implies that at least one mechanism of magma ascent at mid-crustal levels in arcs is the diapiric rise of large sheets during active deformation. Thermal considerations (this mechanism requires warm host rocks and slow cooling of sheets) clearly imply that such a mechanism is unlikely at shallower crustal levels. Sheeted bodies presently exposed in the Cascades core were emplaced at depths greater than 15–20 km. Plutons emplaced at shallower levels tend to be more elliptical and much less sheeted.

Finally, our ascent model implies that, at least in the mid and lower crust, sheeted zones may sometimes represent the advancing front of a magma chamber, rather than the trailing feeder zone. This conclusion could be readily tested by detailed geochemical studies along the sheets.

Acknowledgements—This research was supported by NSF Grants EAR-9218741 awarded to Paterson and EAR-9219536 to Miller. We thank Hugh Hurlow for discussions about Chelan block geology and geochronological data, Ralph Dawes for discussions of Entiat petrology, and Chris Hill for help with drafting figures, and Calvin Miller, Ned Brown, and Sandy Cruden for thoughtful reviews of the manuscript.

REFERENCES

Baer, G. and Beyth, M. (1990) A mechanism of dyke segmentation in fractured host rock. In *Mafic Dykes and Emplacement Mechanisms*, eds A. J. Parker, P. C. Rickwood and D. H. Tucker, pp. 3–12. Balkema, Rotterdam.

Bateman, P. C. (1992) Plutonism in the Central Part of the Sierra Nevada Batholith, California. *United States Geological Survey Professional Paper* **1483**.

Berger, A., Rosenberg, C. and Schmidt, S. M. (1996) Ascent, emplacement, and exhumation of the Bergell pluton within the southern steep belt of the central Alps. *Schweizerische Mineralogisch Petrographische Mitteilungen* **76,** 357–382.

Biot, M. A. (1961) Theory of folding of stratified viscoelastic media and its implication in tectonics and orogenesis. *Geological Society of America Bulletin* **72,** 1595–1620.

Brown, M., Rushmer, T. and Sawyer, E. W. (1995) Segregation of melts from crustal protoliths: mechanisms and consequences. *Journal of Geophysical Research* **100,** 15,551–15,564.

Bruce, P. M. and Huppert, H. E. (1989) Thermal control of basaltic fissure eruptions. *Nature* **342,** 665–667.

Buddington, A. F. (1959) Granite emplacement with special reference to North America. *Geological Society of America Bulletin* **70,** 671–747.

Cadman, A., Tarney, J. and Park, R. G. (1990) Intrusion and crystallization features in Proterozoic dyke swarms. In *Mafic Dykes and Emplacement Mechanisms*, eds A. J. Parker, P. C. Rickwood and D. H. Tucker, pp. 13–24. Balkema, Rotterdam.

Cater, F. W. and Crowder, D. F. (1967) *Intrusive Rocks of the Holden and Lucerne Quadrangles, Washington—The Relation of Depth Zones, Composition, Textures, and Emplacement of Plutons.* United States Geological Survey Professional Paper **1220**.

Cater, F. W. and Crowder, D. F. (1967) *Geologic Map of the Holden Quadrangle, Snohomish and Chelan Counties, Washington.* United States Geological Survey Map, **GQ-646,** scale 1:62,500.

Cater, F. W. and Wright, T. L. (1967) *Geologic map of the Lucerne quadrangle, Chelan County, Washington.* United States Geological Survey Map, **GQ-647,** scale 1:62,500.

Clemens, J. D. and Mawer, C. K. (1992) Granitic magma transport by fracture propagation. *Tectonophysics* **204,** 339–360.

Collins, W. J. and Sawyer, E. W. (1996) Pervasive granitoid magma transfer through the lower and middle crust during non-coaxial compressional deformation. *Journal Metamorphic Geology* **14,** 565–579.

D'Lemos, R. S., Brown, M. and Strachan, R. A. (1992) The relationship between granite and shear zones: magma generation, ascent, and emplacement in a transpressional orogen. *Journal Geological Society London* **149,** 487–490.

Dawes, R. L. (1994) Mid-crustal plutons of the north Cascades: Windows into deep crustal magmatic processes. Unpublished Ph.D dissertation, University of Washington, Seattle.

Grujic, D. and Mancktelow, N. S. (1998) Melt-bearing shear zones: analogue experiments and comparison with examples from southern Madagascar. *Journal of Structural Geology* **20,** 673–680.

Hand, M. and Dirks, P. H. G. M. (1992) The influence of deformation of the formation of axial–planar leucosomes and the segregation of small melt bodies within the migmatitic Napperby Gneiss, Central Australia. *Journal of Structural Geology* **14,** 591–604.

Hopson, C. A. and Dellinger, D. A. (1987) Evolution of four-dimensional compositional zoning, illustrated by the diapiric Duyncan Hill pluton, north Cascades, Washington. *Geological Society of America Abstracts with Programs* **19,** 707.

Hurlow, H. (1992) Studies of the Pasayten fault, Okanogan Range batholith, and southeastern Cascades crystalline core, Washington. Ph.D thesis, University of Washington, Seattle.

Hutton, D. H. W. (1988) Granite emplacement mechanisms and tectonic controls: inferences from deformation studies. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **79,** 245–255.

Jackson, M. P. A. and Talbot, C. (1994) Advances in salt tectonics. In *Continental Deformation*, ed. P. L. Hancock, pp. 159–179. Pergamon Press, Oxford.

Kirby, S. H. (1985) Rock mechanics observations pertinent to the rheology of the continental lithosphere and to the localization of strain along shear zones. *Tectonophysics* **119,** 1–27.

Koons, P. O. (1987) Some thermal and mechanical consequences of rapid uplift: an example from the southern Alps, New Zealand. *Earth Planetary Science Letters* **86,** 307–319.

Koyi, H. (1988) Experimental modeling of role of gravity and lateral shortening in Zagros mountain belt. *Bulletin of the American Association of Petroleum Geologists* **72,** 1381–1394.

Lister, J. R. and Kerr, R. C. (1991) Fluid–mechanical models of crack propagation and their application to magma transport in dikes. *Journal of Geophysical Research* **96,** 10,049–10,077.

Lucas, S. B. and St-Onge, M. (1995) Syn-tectonic magmatism and the development of compositional layering, Ungava Orogen (northern Quebec, Canada). *Journal of Structural Geology* **17,** 475–491.

Marsh, B. D. (1982) On the mechanics of igneous diapirism, stoping, and zone melting. *American Journal of Science* **282,** 808–855.

Mattinson, J. M. (1972) Ages of zircons from the Northern Cascades Mountains, Washington. *Journal of Geology* **40,** 604–633.

McKenzie, D. P. (1984) The generation and compaction of partially molten rock. *Journal of Petrology* **25,** 713–765.

Miller, R. B., Bowring, S. A. and Hoppe, W. J. (1989) Paleocene plutonism and its tectonic implications, North Cascades, Washington. *Geology* **17,** 846–849.

Olsen, P. and Singer, H. (1985) Creeping plumes. *Journal of Fluid Mechanics* **158,** 511–531.

Paterson, S. R. and Fowler, K., Jr (1993) Extensional pluton emplacement models: Do they work for the large plutonic complexes? *Geology* **21,** 781–784.

Paterson, S. R. and Miller, R. B. (1995) Mid-crustal structures in the Cascades core, Washington: Cretaceous contraction or dextral shear? *Geological Society of America Abstracts with Programs* **27,** 71.

Paterson, S. R. and Tobisch, O. T. (1992) Rates of processes in magmatic arcs: implications for the timing and nature of pluton emplacement and wall rock deformation. *Journal of Structural Geology* **14,** 291–300.

Paterson, S. R., Fowler, T. K., Jr and Miller, R. B. (1996) Pluton emplacement in arcs: a crustal-scale exchange process. *Transactions of the Royal Society Edinburgh, Earth Sciences* **87,** 115–124.

Petford, N. (1996) Dykes and diapirs? *Transactions of the Royal Society Edinburgh, Earth Sciences* **87,** 105–114.

Pollard, D. D., Muller, O. H. and Dockstader, D. R. (1975) The form and growth of fingered sheet intrusions. *Geological Society of America Bulletin* **86,** 351–363.

Price, N. J. and Cosgrove, J. W. (1990) *Analysis of Geological Structures.* Cambridge University Press.

Ramberg, H. (1981) *Gravity, Deformation and the Earth's Crust: In Theory, Experiments and Geological Application.* Academic Press, New York.

Rönnlund P. (1987) *Diapiric Walls, Initial Edge Effects and Lateral Boundaries*. UUDNMP Research Report **45**.

Rubin, A. M. (1993) Dikes vs diapirs in viscoelastic rock. *Earth Planetary Science Letters* **119,** 641–659.

Ruppel, C. and Hodges, K. V. (1994) Pressure–temperature–time paths from two-dimensional thermal models: prograde, retrograde, and inverted metamorphism. *Tectonics* **13,** 17–44.

Saffman, P. G. and Taylor, G. (1958) The penetration of fluid into a porous medium or Hele–Shaw cell containing a more viscous liquid. *Royal Society of London Proceedings Series A* **245,** 312–329.

Sawyko, L. (1994) The geology and petrology of the Swakane Biotite Gneiss, North Cascades, Washington. M.S. thesis, University of Washington, Seattle.

Schmeling, H., Cruden, A. R. and Marquart, G. (1988) Finite deformation in and around a fluid sphere moving through a viscous medium: implications for diapiric ascent. *Tectonophysics* **149,** 17–34.

Shaw, H. R. (1980) The fracture mechanism of magma transport from the mantle to the surface. In *Physics of Magmatic Processes*, ed. R. B. Hargraves, pp. 201–264. Princeton University Press, Princeton.

Stevenson, D. J. and Scott, D. R. (1987) Melt migration in deformable media. In *Structure and Dynamics of Partially Solidified Systems*, ed. D. E. Loper, pp. 403–415. Martinus Nijhoff Publishers, Boston.

Tabor, R. W., Zartman, R. E. and Frizzell, V. A. (1987a) Possible tectonostratigraphic terranes in the North Cascades, Crystalline Core, Washington. In *Selected Papers on the Geology of Washington*, ed. J. E. Schuster, pp. 107–127. Washington Division of Mines and Geology Bulletin **77**.

Tabor, R. W., Frizzell, V. A., Whetten, J. T., Waitt, R. B., Swanson, D. A., Byerly, G. R., Booth, D. B., Hetherington, M. J. and Zartman, R. E. (1987b) *Geologic Map of the Chelan 30-minute by 60-minute Quadrangle, Washington.* United States Geological Survey Map **I-1661**.

Talbot, C. J. (1977) Inclined and asymmetric upward-moving gravity structures. *Tectonophysics* **42,** 159–181.

Talbot, C. J., Rönnlund, P., Schmeling, H., Koyi, H. and Jackson, M. P. A. (1991) Diapiric spoke patterns. *Tectonophysics* **188,** 187–201.

Weinberg, R. F. and Podladchikov, Y. (1994) Diapiric ascent of magmas through power-law crust and mantle. *Journal of Geophysical Research* **99,** 9543–9560.

Weinberg, R. F. (1996) Ascent mechanism of felsic magmas: news and views. *Transactions of the Royal Society Edinburgh: Earth Sciences* **87,** 93–104.

Whitehead, J. A. and Helfrich, K. R. (1991) Instability of flow with temperature-dependent viscosity: a model of magma dynamics. *Journal of Geophysical Research* **96,** 4145–4155.