Shear stress in subducting continental margin from high-pressure, moderate-temperature metamorphism in the Ordenes Complex, Galicia, NW Spain

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Abstract

The Ordenes Complex, Galicia, NW Spain, preserves high-pressure, moderate-temperature metamorphism in continental margin rocks subducted during closure of the Rheic Ocean in the Variscan orogeny. The exposures extend across ≈90 km perpendicular to strike and include rocks that reached depths of 30 to 60 km. Estimates of *P–T* conditions of rocks found near the boundary between overriding and subducting plates range from 430 °C at 1.0 GPa to 520 °C at 1.65 GPa. Structural reconstructions including these data indicate an angle of subduction between 15 and 30°.

A mathematical solution and numerical models have been used to estimate shear heating experienced by this well-exposed paleo-subduction zone. Best fit of model to thermobarometric results occurs if shear stress in the upper reaches of the fault separating subducting and overriding slabs was ≈100 MPa (constant shear) or ≈10.0% of pressure (constant coefficient of friction) assuming a convergence rate of 6 cm year^{-1}. At greater depths negative feedback between temperature and shear stress caused the system to approach steady state with decreasing shear stress and with little increase in temperature. The decrease in shear stress at temperatures above 400 °C occurs as the rheological properties of the rock at higher temperature and (or) pressure allow more plastic behavior. This suggests that high-temperature metamorphism is unlikely to occur in subducting crust without special circumstances. A comparison of these results with estimates of shear stress inferred from seismicity and heat flow at active convergent boundaries in the Pacific indicates that shear stress is best described as a pressure-dependent variable not as a constant shear stress.

Keywords: Shear stress; Subduction; High-pressure metamorphism; Mathematical and numerical models; Variscan orogeny; Northwest Spain

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1. Introduction

Understanding processes that lead to high-pressure, low to moderate-temperature regional metamorphism is a goal of geologic research in areas affected by past and present plate convergence. The effects of underthrusting on the formation of blueschist and eclogite facies assemblages remains an important theme of this work, especially the efforts to better understand the nature of shear forces along the main underthrust and their impact on the thermal regime of the subducting crust (Graham and England, 1976; van der Beukel and Wortel, 1987; Molnar and England, 1990; Peacock, 1992, 1996). This problem has been approached in a variety of ways including measurement of heat flow to the surface above subducting crust (e.g. Tichelaar and Ruff, 1993; Springer, 1999; Von Herzen et al., 2001), using the depth of seismically active crust to estimate temperature gradients along the thrust (Tichelaar and Ruff, 1993; Peacock and Wang, 1999), and pressure–temperatures conditions inferred from blueschist and eclogitic rocks (Peacock, 1990, 1992, 1996). Each method relies on thermal models to estimate the contributions of various heat sources within the subduction zone. Typically heat that cannot be explained as derived from normal mantle flux or radiogenic sources is interpreted to result from shear heating (England and Richardson, 1977; England and Thompson, 1984; Peacock, 1990; Tichelaar and Ruff, 1993; Ernst and Peacock, 1996; Springer, 1999; Von Herzen et al., 2001).

The Basal Units of the Ordenes Complex, Galicia, NW Spain, are comprised of subducted continental margin rocks. This paleo-convergent margin is currently exposed across ≈90 km perpendicular to the strike of the orogen. Relict mineral assemblages in porphyroblasts have been used to determine pressure–temperature conditions during subduction. Estimates consistently increase from east to west, current coordinates, from 430 °C at 1.0 GPa to 520 °C at 1.65 GPa. In this paper we report results of models using the constraints provided by these data to better understand how shear stress impacted metamorphism of the Basal Unit during subduction.

2. Geologic setting

The Iberian Massif of northwest Spain is characterized by three allochthonous complexes thrust onto Upper Proterozoic and Paleozoic sequences and intruded by syn- to postkinematic Variscan granitoids (Fig. 1). The complexes consist of three groups of units, Upper, Ophiolitic and Basal, that were stacked during the closure of the Rheic Ocean at the beginning of the Variscan orogeny. The Upper Units represent pieces of a suspect terrane, probably an island arc (Andonaegui et al., 2002), and the Ophiolitic Units, below, are fragments of oceanic lithosphere of the Rheic Ocean or of basins marginal to it. The Basal Units represent the outermost edge of the northern passive margin of Gondwana, and underwent subduction below the accretionary prism formed previously by the stacking of Upper and Ophiolitic Units (Martínez Catalán et al., 1996, 1997, 2002). Their subduction marked the transition from oceanic closure to Variscan collision, and was followed by their exhumation along large thrusts, helped by the thinning of the overlying orogenic wedge by extensional detachments.

Ordenes is the largest of the allochthonous complexes in NW Spain and is exposed as a klippe, preserved in a late-Variscan synform. The four Basal Units outcrop along its southern and western margins (Figs. 1 and 2A). These Units consist of metasediments (schists and paragneisses), amphibolites and orthogneisses. The latter are extremely flattened metagranites of Ordovician age, and constitute, together with the amphibolites, excellent markers to identify large structures. From them it can be established that the Santiago, Lalín, and Forcarei Units were initially continuous (and remain nearly so) and became folded in a recumbent antiform, carried to the east by the Lalín-Forcarei thrust and cut by the late Pico Sacro extensional detachment (Fig. 2A).

High-pressure metamorphism has been identified in the Basal Units. Pressure–temperature (P–T) conditions were established using microinclusions of the oldest-fine-grained foliation preserved in albite porphyroblasts (Arenas et al., 1995) in schists and in relict eclogite boudins (Arenas et al., 1997; Rubio Pascual et al., 2002). The metamorphic conditions reached during the high-pressure metamorphism in the Santiago Unit (location C) were 650 MPa and 90

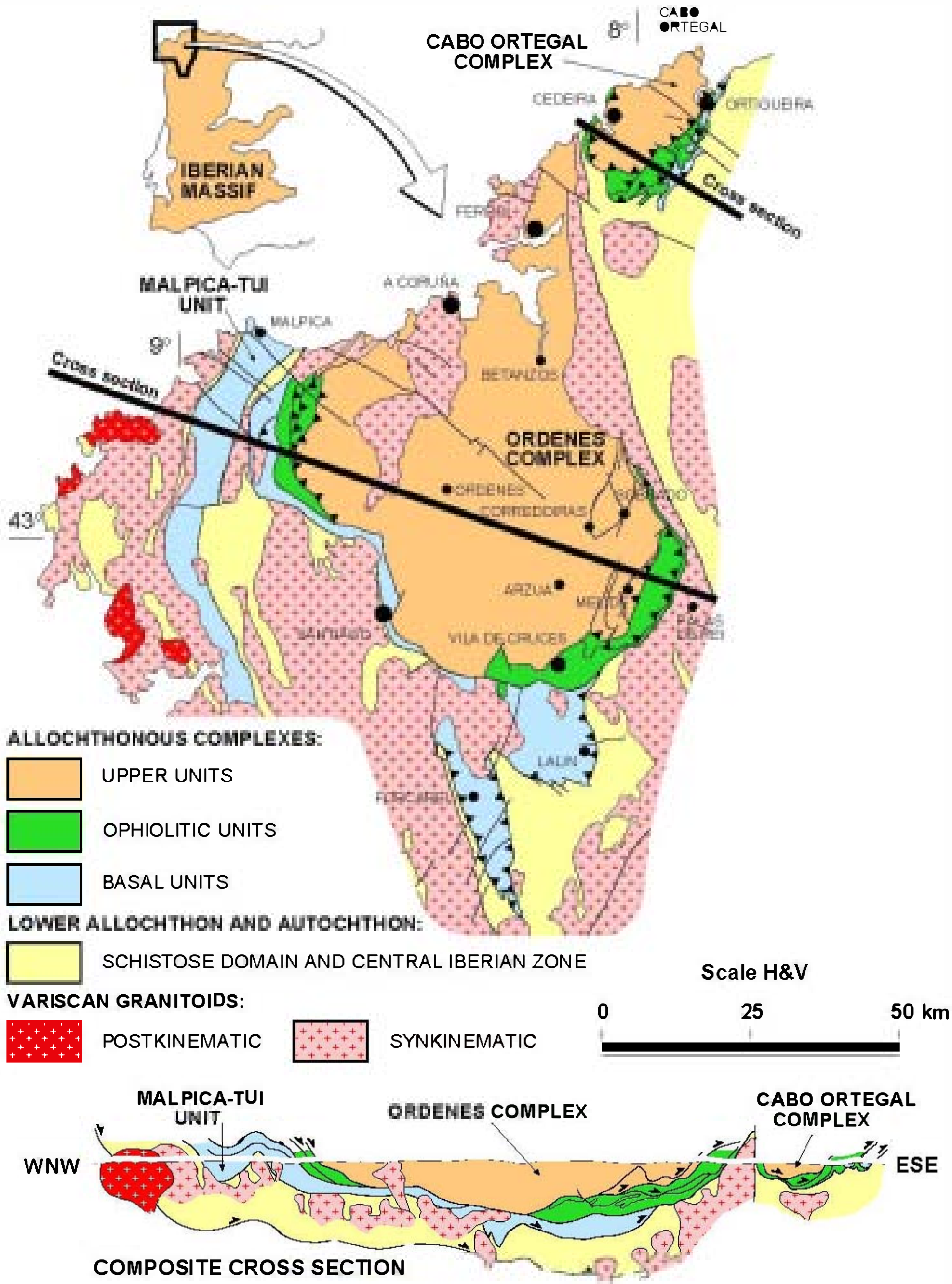

Fig. 1. Geological sketch-map and cross-section of NW Spain showing the three allochthonous complexes of Cabo Ortegal, Ordenes and Malpica–Tui.

°C higher than in the Forcarei Unit (location A), implying a sense of subduction toward the west (in present coordinates; see Fig. 2 and Table 1). Moreover, evidence of heat advection transferred from above in the upper part of the Santiago and Lalín Units (Martínez Catalán et al., 1996, 2002) and the

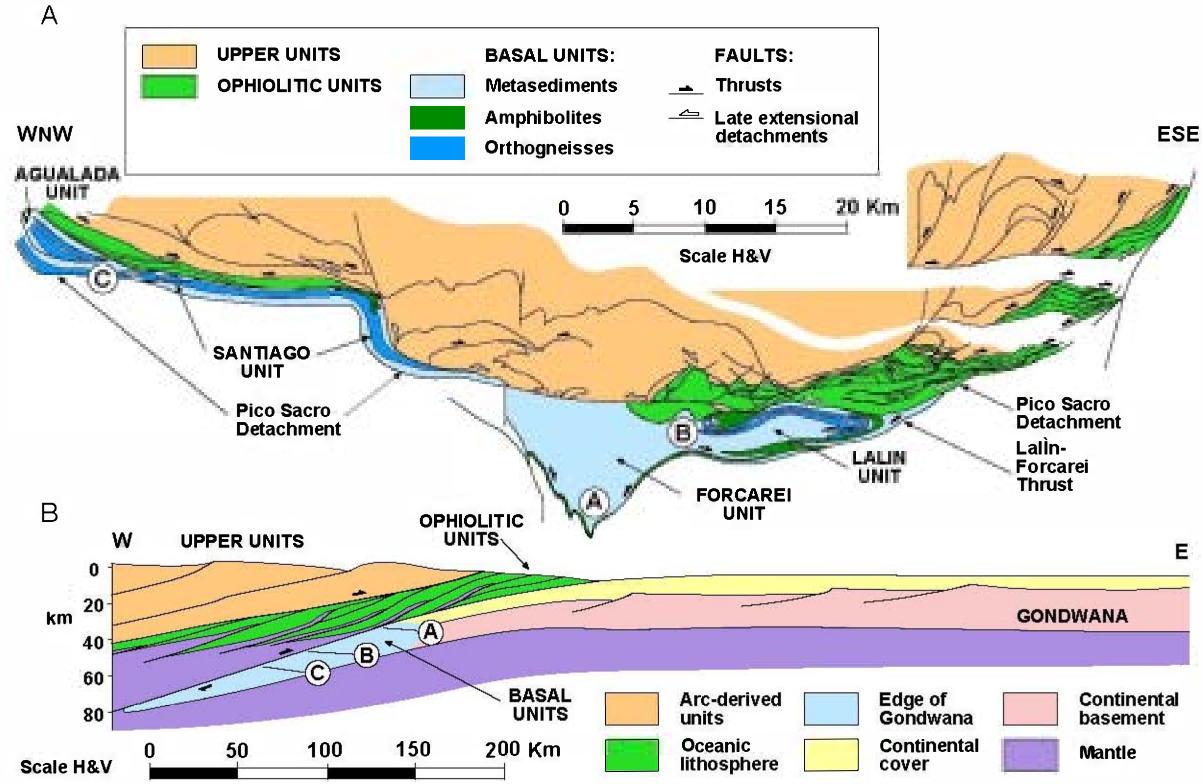

Fig. 2. A: Schematic geological composite section across the Ordenes Complex depicting the relationships among the three groups of allochthonous units and the internal structure of the Basal Units (after Martinez Catalán et al., 2002). A, B and C show locations where *P–T* estimations were made by Arenas et al. (1995) and Martinez Catalán et al. (1996). B: Tectonic model showing the subduction of the outermost edge of northern Gondwana, represented by the Basal Units, below an accretionary prism formed by the Upper and Ophiolitic Units in stages prior to the Variscan collision. The approximate position of sample locations in the upper part of the subducting slab (A, B and C) allows estimate of angle of subduction.

preservation of subophiolitic mantle on top of the latter suggest that the Basal Units occupied the uppermost part of the subducting slab.

These data permit an estimate of the paleosubduction dip. Taking a mean density of 3000 kg m^{-3} for the wedge above the over-riding slab (Fig. 2B), the difference in pressure inferred from metamorphic assemblages corresponds to a depth increment of 22 km. The outcrops used to estimate *P–T* data A and C (Fig. 3) are presently separated by 89 km as measured following folded orthogneiss and amphibolite bodies (Fig. 2A). However, this distance must be reduced because the Santiago Unit was cut and displaced at least 5 km to the WNW by the Pico Sacro detachment. Furthermore, the major recumbent anticline folding the Basal Units caused stretching in both the normal (54 km) and reverse (30 km) limbs of the recumbent fold. Strain was concentrated in the reverse limb, where the orthogneisses and amphibolites were flattened to layers just a few meters thick by shearing along the Lalín-Forcarei thrust (Fig. 2A). Although the stretching cannot be calculated in the absence of reliable strain markers, we may constrain the original distance between A and C by estimating the extension between zero (no extension) and one (100% extension, possibly exceeded in the reverse limb but unreasonable for the larger normal limb). Removing 5 km of movement along the Pico Sacro detachment in the direction of the cross section, and between 0 and 42 km for the stretching in both limbs of the recumbent fold, we estimate the distance from A and C to have been between 84 and 42 km before folding and extensional faulting. Combining these distances along dip with the depth

Table 1

Summary of evidence for initial high-P metamorphism in the Basal Units of the Ordenes Complex

Location and high-P metamorphism	High-P mineral assemblages	*P–T* values	Methods used for *P–T* calculations
C: Santiago Unit. Eclogite facies. Metapelites: garnet zone. Metabasites: eclogites and amphibolites in the hornblende–garnet zone.	Schists: garnet (Grs=15–19 mol%)–phengite (Cel=27–33 mol%)–chlorite (XMg=0.45–0.52)–albite (An<3 mol%)–quartz–clinozoisite (Pst=23–28 mol%)–rutile–ilmenite. Amphibolites: garnet (Grs=26–32 mol%)–hornblende–phengite (Cel=13–34 mol%)–chlorite (XMg=0.50–0.65)–albite (An<3 mol%)–clinozoisite–quartz–rutile–ilmenite. Eclogites: garnet (Grs=22–32 mol%)–omphacite (Jd=38–43 mol%)–zoisite–paragonite (Pg=82 mol%)–phengite (Cel=22–25 mol%)–quartz–rutile.	520 °C 1.65 GPa	*P–T* estimations based on Arenas et al. (1995), Martinez Catalán et al. (1996) and Rubio Pascual et al. (2002). Metapelites: -Garnet–phengite thermometry (Green and Hellman, 1982; Hynes and Forest, 1988; Krogh and Råheim, 1978). -Garnet-chlorite thermometry (Ghent et al., 1987). -GRIPS barometry (Bohlen and Liotta, 1986). -Phengite barometry (Massone and Schreyer, 1987).
B: Lalín Unit. (Upper part of the unit). Blueschist–eclogite facies transition. Metapelites: garnet zone. Metabasites: glaucophane–hornblende–garnet zone.	Schists: garnet–phengite–chlorite–albite–quartz–clinozoisite–rutile–ilmenite. Amphibolites: garnet–hornblende–chlorite–albite–clinozoisite–quartz–rutile–ilmenite. Eclogites not present.	470 °C 1.35 GPa	–Selected equilibria in grids. Metabasites: -Garnet–clinopyroxene thermometry (Krogh, 1988). -Garnet–hornblende thermometry (Graham and Powell, 1984).
A: Forcarei Unit. (Lowest part of the unit). Blueschist facies. Metapelites: chlorite–garnet zone. Metabasites: lawsonite–glaucophane zone	Micaschists: phengite–chlorite–±garnet–albite–quartz–rutile–ilmenite. Greenschists: lawsonite (pseudomorphed by epidote–clinozoisite)–actinolite–chlorite–albite–quartz–rutile–ilmenite.	430 °C 1.0 GPa	-Jadeite–albite–quartz barometry (Holland, 1980, 1983). -GRIPS barometry (Bohlen and Liotta, 1986). -Garnet–hornblende barometry (Kohn and Spear, 1990). -Selected equilibria in grids.

increment, the dip of subduction ranges between 15 and 32°.

Other geological data of interest to the models are the age of the oceanic crust subducted before the continental margin of Gondwana and the rate of convergence during the closure of the Rheic Ocean.

Oceanic crust of the youngest ophiolitic unit in the Ordenes Complex has been dated at 395 Ma (U–Pb in zircons; Díaz García et al., 1999; Pin et al., 2002). This and similar units were underthrust between 390 and 380 Ma (Peucat et al., 1990; Dallmeyer et al., 1991, 1997), as deduced from $^{40}Ar/^{39}Ar$ ages of the amphibolite-facies foliation. The subduction of the continental margin represented by the Basal Units is constrained by $^{40}Ar/^{39}Ar$ geochronology of phengites in eclogites, dated around 370–365 Ma (Rodríguez et al., 2003). Consequently, oceanic crust generation of the youngest ophiolitic unit preceded continental subduction by 20–25 m.y. In addition, any thermal effect of that young oceanic lithosphere would have been masked by the subsequent subduction of older ophiolites that formed adjacent to the Gondwana margin at the beginning of the Early Ordovician (Arenas et al., 2004; Sánchez Martínez et al., 2004), and the thermal regime of the subduction zone probably equilibrated under conditions controlled by subduction of older lithosphere with limited heat transfer from the asthenosphere.

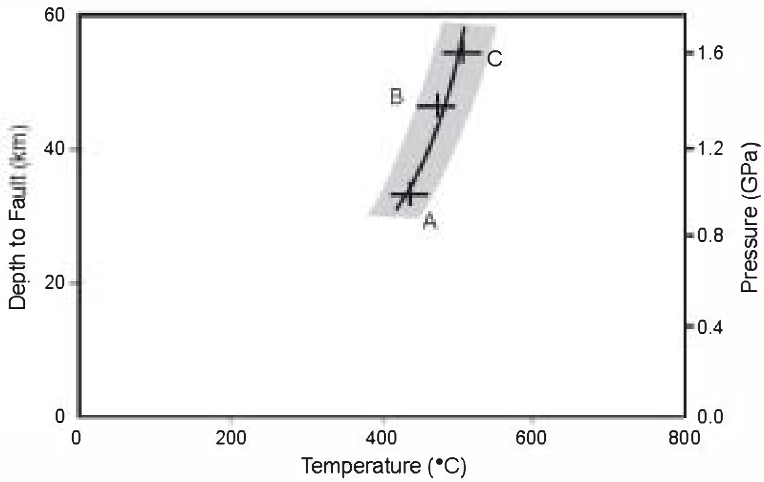

Fig. 3. Pressure–temperature conditions near the top of the Basal Units of the Ordenes Complex, according to Arenas et al. (1995, 1997) and Martínez Catalán et al. (1996). See Fig. 2 for sample location and Table 1 for data used in thermobarometry. Error bars are approximate.

To assess the rate of convergence, we use the continental reconstructions based on paleomagnetism. According to Scotese and McKerrow (1990) and Scotese (2002), Laurentia and Baltica joined each other by the Middle Silurian (425 Ma) and their southern margins were separated from northern Gondwana by an arc of 20–25°, equivalent to 2200–2800 km. Assuming that the subduction of the Basal Units represents the closure of the Rheic Ocean, and that it started ≈385 Ma, the convergence rate can be estimated between 5 and 7 cm year^{-1}.

3. Estimating shear stress

Molnar and England (1990) presented a mathematical solution for estimating temperatures along the surface of subducted lithosphere for systems that have reached thermal equilibrium:

$$T = \{(Q_b + \tau V)z/k\}/S \tag{1}$$

where S is a term that approximates the effect of advection on the system. The value of S is given by

$$S = 1 + \left\{b[(Vz\sin\delta)/\kappa]^{0.5}\right\} \tag{2}$$

Table 2 lists symbols, units and representative values used in Eqs. (1) and (2).

The term b is a numerical constant that serves as a correction for changes in the amount of heat added to a particular increment of the fault as compared to the average heat added to thrust by shear stress. In the case of constant shear stress, b is equal to 1.0 because the average stress and stress at any point along the thrust are equal. However, if shear heating increases linearly with depth (constant coefficient of friction) then the heat added at depth z will be greater than the average heat added to the fault surface above depth z. Molnar and England (1990)

Table 2
Equations and parameters used to estimate shear stress

Eq. (1) (Molnar and England, 1990)	$T = \{(Q_b + \tau V)z/k\}/S$	
Eq. (2) (Molnar and England, 1990)	$S = 1 + \left\{b[(Vz\sin\delta)/\kappa]^{0.5}\right\}$	
Q_b	Heat to base of lithosphere (W m^{-2})	0.05[a]
τ	Shear stress (MPa)	
V	Velocity of subduction (m s^{-1})	
z	Depth from surface perpendicular to fault plane (m)	
k	Thermal conductivity (W m^{-1} K^{-1})	2.5[a]
δ	Dip of fault plane	
κ	Thermal diffusivity (m^{2}s^{-1})	8.3×10^{-07}[a]
γ	Coefficient of friction	
	Density (kg m^{-3})	3000[a]

[a] Values used in most numerical and mathematical solutions.

and Peacock (1992) set b=1.33 to correct for this effect and reported that results obtained from this value are consistent with results obtained from finite-difference numerical experiments that model the process.

In this investigation, we first used the mathematical solution to estimate shear heating along the thrust necessary to create the thermal conditions inferred from the metamorphic assemblages of the Basal Units of the Ordenes Complex. However, as is shown below, temperatures experienced by the Basal Units do not conform to either a constant shear force or a constant coefficient of friction. Instead, the geotherm along the thrust surface flattens above a threshold temperature and (or) pressure implying that shear stress decreases above that threshold. This result is consistent with the expected rheological properties of crustal rock.

Because the numerical solution assumes either a constant shear stress or constant coefficient of friction, it is not appropriate to use the approach to infer relationships between shear stress and temperature if shear stress is temperature dependent. For this reason, a two-dimensional finite-difference model was employed to include feedback between shear stress and temperature along the thrust. Shear stress above the threshold of 400 °C was calculated using the equation

$$\tau_z = \tau^*\mathrm{Exp}(-(\mathrm{temp}_{z-1} - 400)/75) \qquad (3)$$

where τ_z is the shear stress at depth z and temp_{z-1} is the temperature of the thrust at depth $z-1$ (after Peacock, 1996). Fig. 4 presents a schematic representation of the numerical model.

4. Model results and discussion

Figs. 5, 6 and 7 summarize estimates of the strength of the shear stress acting on the subducting slab necessary to produce metamorphic conditions that affected the Basal Units. Also shown are results from experiments used to determine the effect of changing the value of specific variables on temperature estimates resulting from a given shear stress. Using best estimates of initial conditions (δ=20°, V=6 cm year^{-1}, and ρ_b=0.05) shear stress acting at the thrust was ≈100 MPa or 10.0% of pressure between 0 and 30 km. If forces are averaged to depths of 60 km, shear stress was less, ≈85 MPa or 3.5% of pressure. This decrease in average shear stress with depth is best explained if shear forces decreased when a threshold temperature and (or) pressure was exceeded. Models that place that threshold at 400 °C provide a good fit with the observed conditions (Fig. 7).

5. Robustness of results

A number of issues might affect the confidence that one can place in these estimates of shear stress. First, the mathematical solution used applies only to systems that have attained thermal equilibrium. However, if rates of shear heating decreased rapidly above the threshold temperature, then equilibrium would be reached relatively quickly in rocks at depth. A numerical model has been used to test this hypothesis and shows that within 5 m.y. (V=6 cm year^{-1}, δ=20°), the thrust surface has experienced

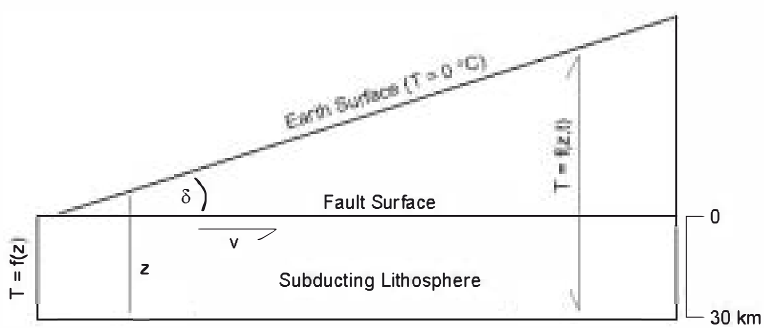

Fig. 4. Visual representation of finite-difference model used to estimate shear stress and the effect of negative feedback between shear stress and temperature above the threshold temperature of 400 °C.

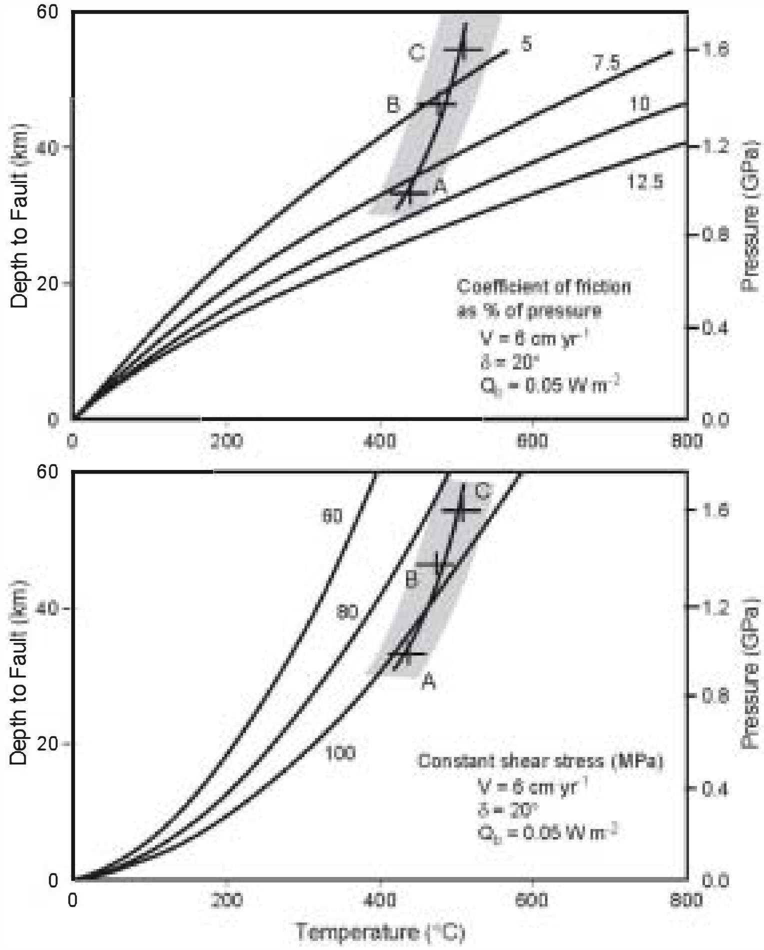

Fig. 5. Estimated temperature along a thrust derived from Eq. (1) (Molnar and England, 1990) assuming constant coefficient of friction or constant shear stress. A comparison of results with other estimates of shear heating by other methods indicates that a constant coefficient of friction better approximates shear stress until depths of subducted rocks exceed 30 km. At greater depths negative feedback reduces shear stress and acts to stabilize temperature. Points A, B, and C refer to the sample locations identified in Fig. 2. Shaded area includes estimate of uncertainty.

93% of the cooling required to reach equilibrium (Fig. 8). It follows that the deeper portions of the underthrust would have approached equilibrium within the 15 to 20 m.y. that occurred between subduction of the youngest ophiolitic unit and the subduction of the Gondwanan margin rocks that form the Basal Units.

A large change in the angle of subduction will result in significant differences in the temperature of the underthrust for a given amount of shear stress. The lower the angle, the higher the temperature at depth z because the rock must travel farther and experience more shearing to reach that depth. The effect of the angle of subduction on shear heating for a given shear stress is presented in Fig. 6B As discussed below, estimates of shear stress reported here are higher than estimates reported elsewhere, it seems unlikely, therefore, that the dip of subduction was significantly greater than 20°.

A third consideration is the age of the oceanic crust that was subducted prior to involvement of the continental margin. Young oceanic crust with higher heat flow to the surface and higher geotherm has been shown to affect the seismic properties of subducting crust (Peacock and Wang, 1999). However, as seen in

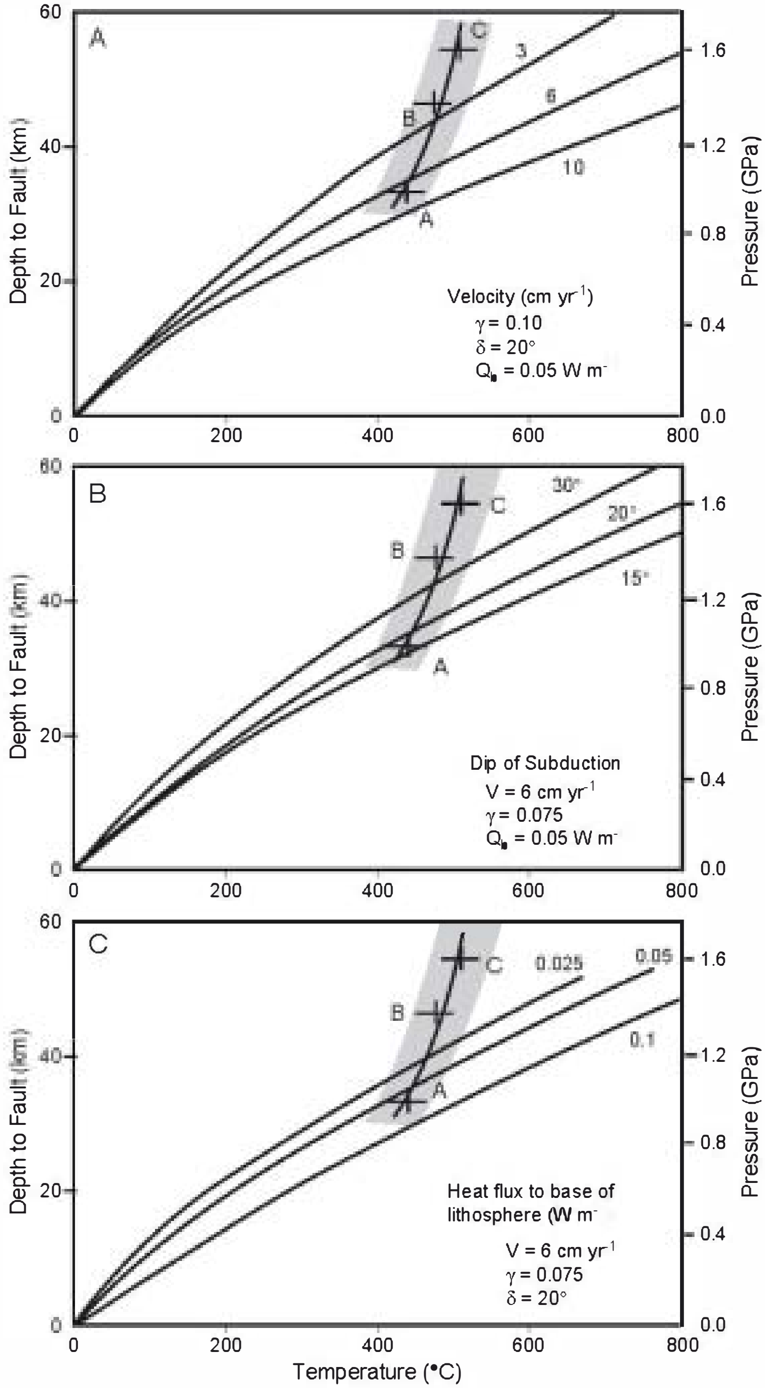

Fig. 6. Comparison of model temperatures along underthrust with temperature gradient inferred from metamorphic assemblages found near the upper surface of the Basal Units. Points A, B, and C refer to the sample locations identified in Fig. 2. Shaded area includes estimate of uncertainty. Figures show : A: Effect of velocity on heating resulting from shear stress, B: Effect of the dip of subduction on heating of the thrust surface, and C: Effect of heat flux to the base of the lithosphere (Q_b) on thermal character of the thrust surface. Model parameters are included in each figure.

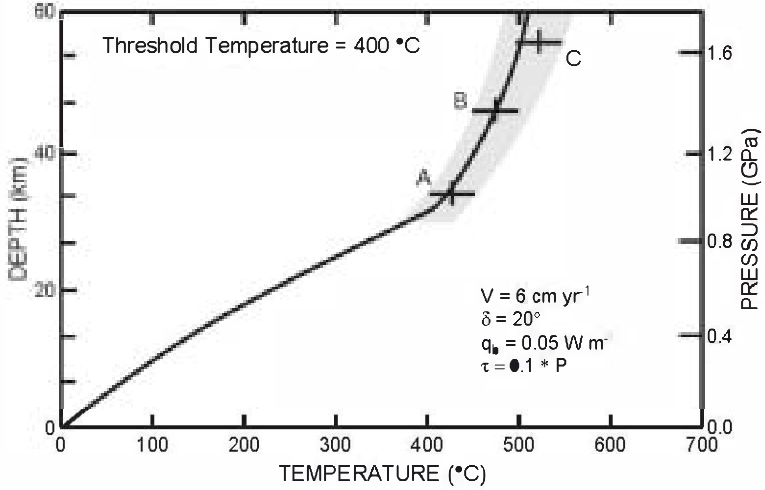

Fig. 7. Best-fit model based on a constant coefficient of friction at temperatures less than the brittle-plastic transition set at 400 °C and exponentially decreasing shear stress if temperature exceeds the transition. Results obtained from two-dimensional finite-difference model.

Fig. 6C, the effect of increasing basal heat flow on predicted temperatures at depth is relatively small. A doubling of heat flow from 0.05 W m^{-2} to 0.10 W m^{-2} is equivalent to an increase of the coefficient of friction from 8% to 10%. The impact would be even smaller at the depths of interest (30–60 km) because negative feedback between temperature and shear stress acts to limit temperature above ≈400 °C.

Perhaps the most important variable affecting the estimate of shear stress experienced by the Basal Units is the velocity of convergence. The velocity of convergence has been estimated to be between 5 and 7 cm year^{-1} based on the plate reconstructions of Scotese and McKerrow (1990) and Scotese (2002) and a velocity of 6 cm year^{-1} has been used in most models. Slower convergence would reduce heating derived from a given amount of shear. The coefficient of friction would have to increase by >30% to reach a temperature ≈400 °C at a depth of 33 km if the convergent rate were 3, not 6 cm year^{-1}. Rapid convergence of 10 cm year^{-1} would reduce the necessary coefficient of friction by ≈20% (Table 3).

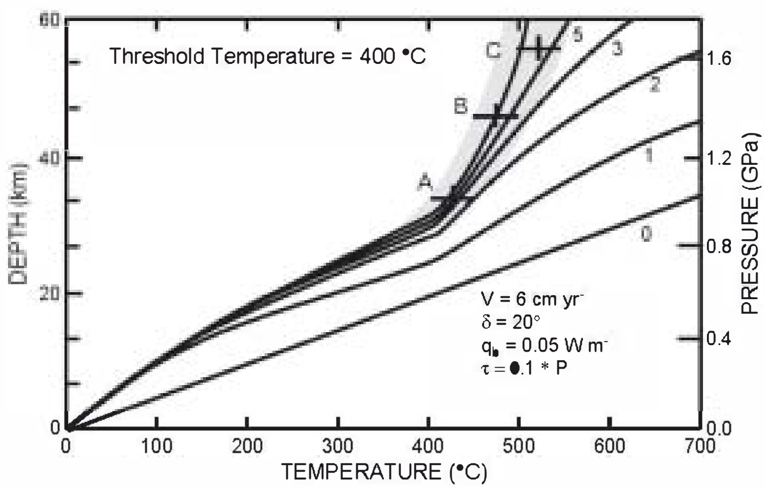

Fig. 8. Development of thermal regime along overthrust as subduction zone approaches thermal equilibrium. Thermal gradient along thrust is shown at 0, 1, 2, 3, and 5 m.y. after initiation of subduction. Equilibrium is established after 10 to 15 m.y.

Table 3
Representative results from numerical solution

Coefficient of friction	Dip	Velocity (cm year^{-1})	Temperature (°C at 33 km)	Temperature (°C at 55 km)
Temperature inferred from mineral assemblages and compositions			430	520
0.05	15	6	278	**532**
		10	319	637
	20	6	246	470
		10	282	561
	30	6	208	396
		10	238	470
0.075	15	6	368	733
		10	**440**	**903**
	20	6	327	647
		10	389	795
	30	6	276	**545**
		10	327	667
0.1	15	6	458	934
		10	560	1169
	20	6	**407**	825
		10	494	1030
	30	6	344	694
		10	**416**	864
0.125	15	6	548	1134
		10	680	1435
	20	6	487	1002
		10	601	1264
	30	6	**412**	844
		10	506	1061

Bold indicates results consistent with metamorphic conditions.

Finally, the models that provide the best fit to the observed data (Fig. 7) have shear stress decrease exponentially at temperatures above a threshold ≈400 °C. This will have the effect of limiting the impact of the uncertainty associated with various parameters to the region of the thrust fault experiencing temperatures below the threshold. Because shear stress decreases rapidly above the threshold, the effect of changing the angle of subduction, the velocity of convergence or the heat delivered to the base of the lithosphere become less important to determining the temperature during high-pressure metamorphism. In fact, the results suggest that high-temperature, high-pressure metamorphism requires special circumstances to occur within subducting crust. Possible causes might be intrusion of high-temperature magmas, the subduction of a spreading ridge, or the preservation of metamorphism that occurred during the initial stages of subduction before thermal equilibration.

6. Comparisons with results from other methods

Estimates of shear heating in subduction zones made using heat flow measurements above ten circum-Pacific subduction zones are 14 MPa (constant shear stress) or 5.9% of pressure (constant coefficient of friction) (Tichelaar and Ruff, 1993). A study of heat flow in the Central Andes yielded similar results (Springer, 1999); however, a recent study of heat flow in the Kermadec forearc yields significantly higher estimates of shear stress (Von Herzen et al., 2001). The latter study did not consider the contribution of radiogenic heating in the hanging wall to surface heat flow and so may overestimate shear forces. The significant difference between estimates of constant shear stress and better agreement between estimates of the coefficient of friction reported by Tichelaar and Ruff (1993) and in this study suggest that these differences may be caused by differences in methodology. Most likely the low estimates of constant shear stress derive from Tichelaar and Ruff's interest in the upper reaches of the subduction zone where the fault is seismically active. At a depth of 10–15 km, temperatures of the fault surface inferred from Tichelaar and Ruff's estimates of shear stress as constant shear stress or as a constant coefficient of friction are similar. With increasing depth, the difference in temperature derived from the two models increases, with temperatures predicted by a constant coefficient of friction becoming significantly higher than those based on constant shear stress.

Tichelaar and Ruff (1993), Ruff and Tichelaar (1996) also used their estimate of shear stress to calculate temperature at the transition of the fault from seismically active to quiescent behavior. Using a constant coefficient of friction, they estimate that the transition occurs at ≈400 °C if subduction occurs beneath thickened crust. This value is again consistent with best-fit models of shear stress acting on the Basal Units.

The better agreement between estimates of the coefficient of friction obtained by Tichelaar and Ruff (1993) and by this study and the significant difference between results obtained from constant shear stress indicate that shear stress is best modeled as a pressure-dependent variable. Although estimates of the coefficient of friction derived from the different studies are similar, the estimates reported here that are based on metamorphism of the Basal Units are significantly higher, 10% to 5.9%. The source of this difference is uncertain. It may be caused by differences in the subduction zones themselves. For example, the rheological properties of continental margin rocks of the Basal Units may be sufficiently different from basaltic oceanic crust to change shear heating in the subduction zone.

The presence of blueschists and eclogites among exhumed rocks has also been used to constrain thermal conditions within subduction zones (Peacock, 1992, 1996). However, because most exposures of these rocks are of limited extent and bounded by later faults associated with their uplift and return to the surface, their position within the subducted crust is not well constrained. If they originally were in the interior of the subducted lithosphere, away from the fault surface, then shear heating would have less impact on them. Should these rocks be used to estimate temperatures along the thrust and by extension the amount of shear heating required to produce those temperatures, they would yield underestimates of both. Furthermore, evidence derived from the Basal Units implies that shear stress decreases above a threshold temperature. It follows that single exposures of moderate-temperature, high-pressure metamorphic rocks can only be used to estimate the average shear stress that affected the rocks during underthrusting. Estimating actual shear stress at any particular depth requires a more complete image of the subduction zone, similar to that obtained from the multiple exposures of the Basal Units of the Ordenes Complex.

7. Conclusion

Exposures in the Ordenes Complex of continental margin rocks that were subducted to depths ranging between 30 and 60 km provide a view of these portions of a paleo-subduction zone. Temperatures and depths derived from study of the high-pressure, moderate-temperature metamorphism of these rocks and the dip of subduction estimated from structural reconstructions constrain estimates of shear forces acting along the underthrust. Best estimates are that shear stress was ≈10% of pressure at temperatures below a threshold of 400 °C. Once temperature exceeded the threshold, the coefficient of friction decayed exponentially.

Additionally, the decrease in shear stress above the threshold implies a negative feedback between shear stress and temperature. Negative feedback tends to stabilize system behavior and so causes temperature and shear stress to approach constant values. One implication is that moderate-temperature metamorphism of high-pressure and ultra-high-pressure rocks can only be used to estimate average shear stress in the subduction zone above those rocks. The results also imply that high-temperature, high-pressure metamorphism should be rare in subducting crust because it requires an additional heat source, distinct from the normal processes of subduction.

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