

THE EXHUMATION OF CROSS SECTIONS OF THE CONTINENTAL CRUST: STRUCTURE, KINEMATICS AND RHEOLOGY

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ABSTRACT. The exhumation of coherent cross sections of the continental crust is a selective, usually multistage tectonic process that involves special kinematic and geothermal evolutions. Three general types of exhumation history are distinguished: (1) uplift at low to moderate geothermal gradients in a convergent intracratonic setting; (2) extensional uplift at high geothermal gradients followed by emplacement near an active plate margin; (3) uplift at high-temperature conditions in the core of a deeply eroded intraplate or plate-margin orogen. The degree of structural coherence among different crustal levels of a crustal section is generally greater in sections with type 1 and 2 histories than in sections with a type 3 history. Thermal disequilibrium during shearing led to asymmetrical strain and fabric distribution across large shear zones that accommodated uplift of the crustal sections. The subduction, underplating and thickening of continental crust beneath exposed crustal cross sections produced anomalous lithospheric structures that have isostatically balanced dense lower crustal rocks at or near the surface over long periods of time. Relict crustal features that predate the exhumation are best preserved in crustal sections where cooling induced a pronounced partitioning of stress and strain already during the early stages of uplift.

1. Introduction

Exposed cross sections of the continental crust are large terrains in which most levels of the crust have been brought to the surface as a coherent package. Rocks exposed at the surface vary systematically from unmetamorphosed to weakly metamorphosed upper crustal assemblages to highly metamorphosed, dense lower crustal units. The interpretation of these terrains as crustal cross sections is based on the observation that cross-sectional trends in their lithologic, metamorphic, structural and geophysical characteristics are consistent with petrologic models of the crust and with the imaged geophysical properties of unexposed continental crust (Fountain and Salisbury, 1981). It is the general continuity among the different crustal levels exposed at the surface that distinguishes these terrains from other suites containing lower crustal rocks and makes exposed crustal cross sections attractive for studying evolutionary crustal processes in the field. In order to determine whether such crustal cross sections are really representative samples of

unexposed continental crust, we must determine how entire sections of the crust are brought to the surface and assess the effect that uplift has on their original internal structure, metamorphism and composition.

Many popular models of crustal exhumation pertain either to (1) medium- to high-grade rocks in the cores of eroded orogens, or to (2) the special case of high-pressure metamorphic rocks within subduction melanges (see brief reviews in Fountain and Salisbury, 1981; Platt, 1986). However, recent progress in reconstructing the history of some crustal cross sections (reviews in this volume) reveals that cross sections of the continental crust are exposed in a much wider range of tectonic settings than are considered in these tectonic models. Certainly, applying uplift models for high-pressure metamorphic terrains to explain the exhumation of coherent crustal cross sections is unsatisfactory because the tectonothermal history of lower crustal granulites in crustal cross sections (next section) differs considerably from that of blueschist facies terrains (Ernst, 1988). These exhumation mechanisms result in extensive imbrication of rocks with contrasting metamorphic histories, rather than the preservation of structural and metamorphic continuity among different crustal levels.

This paper reviews the uplift history of several exposed crustal cross sections and relates the kinematics of their exhumation to the evolution of thermobarometric conditions and deformational style in different crustal levels. Experimental and theoretical rock-mechanical concepts are used to constrain the long-term, high-strain lithospheric rheology during uplift. Consideration of how stress and strain partition during uplift is crucial to understanding why lower crustal rocks survive the exhumation process and sections of continental crust maintain their structural integrity.

2. The Exhumation History of Some Exposed Crustal Cross Sections

Tables I and II summarize the uplift characteristics of cross sections where the tectonometamorphic evolution is reasonably well constrained. There are three main types of exhumation history: (1) cold intracratonic uplift along thrusts rooted in the lower crust; (2) hot extensional uplift of the section to depths of 10 to 15 km, followed by cold emplacement to shallow depths near a convergent plate boundary; (3) repeated uplift under high-temperature conditions in the core of an intracratonic or a plate boundary orogen. It is important to note that not all of the cross sections in Tables I and II are equally well exposed or well studied. Also, many more crustal cross sections probably exist in other areas than are currently reported in the literature or even recognized in nature. Nevertheless, some generalizations can be made about their evolution and exhumation.

The exposed crustal cross sections in Table I experienced type 1 or 2 uplift histories and have a high degree of structural integrity. Regardless of age and tectonic setting, these crustal cross sections all show (1) a granulite facies lower crust consisting of sill-like mafic intrusive rock and felsic rock of supracrustal origin, (2) a middle crust containing amphibolite to greenschist facies metasediment and intermediate to felsic intrusive rock, and sometimes (3) an unmetamorphosed to weakly metamorphosed upper crust. Generally, the ratio of mafic to felsic rock increases with paleodepth in the sections. Complex polyphase deformation associated with regional metamorphism clearly predates shearing and retrograde metamorphism during uplift (references, Table I). Therefore, the overall structure of the crustal cross sections in Table I was established at high geothermal gradients well before exhumation began. The structurally deepest parts of the cross sections yield peak geobarometric estimates of 8 to 12 kilobars and contain only volumetrically minor, laterally discontinuous outcrops of ultrabasic rock. This indicates that lower crustal rock presently exposed at the surface lay no deeper than 30 to 45 kilometers at the time of regional metamorphism. Either the original crustal thickness did not exceed this amount and

Table I Uplift Characteristics of Some Relatively Coherent Crustal Cross Sections

Crustal Section	Plate Tectonic Setting	Major Fault(s) active during Uplift	Fault Rocks & Metamorphic Conditions	Age of Uplift
Kapuskaing (6, 7, 8)	Intracratonic: Superior Province (Ontario, Canada)	Ivanhoe Lake fault zone: thrust at base of section	retrograde greenschist f. mylonite & cataclasite	2690 - 2450 Ma. 2200 - 1900 Ma.
Ivrea (4, 9, 11)	Plate Margin: European/Apulian plates (southern Alps, N. Italy, Europe)	Pogallo shear zone in the basal intermediate crust and other extensional mylonitic zones in the lower crust Fault at base of section truncated by Insubric line Insubric line: strike-slip & thrust fault at base of section	retrograde amphibolite f. to greenschist f. mylonite in footwall grading to cataclasite in hangingwall retrograde mylonite & cataclasite? cataclasite; greenschist f. mylonite at northern rim of section	270 - 160 Ma. transension & extension 75 - 60 Ma. Eo-Alpine stage 27 - 17 Ma. Insubric phase
Fiordland (1, 5)	Plate Margin: Pacific/Tasman plates (Southern Alps, South Island, New Zealand)	Doubtful Sound shear zone: extensional shear zone within the lower crust Alpine fault: strike-slip & thrust fault near base of section	retrograde amphibolite f. & greenschist f. mylonites in footwall grading to cataclasite in hangingwall cataclasite	100 - 80 Ma. ca. 40 - 15 Ma.
Pikwitonei (2, 3, 10)	Plate Margin: Superior/Churchill blocks (Manitoba, Canada)	extensional shear zones in the lower crust Thompson mobile belt: strike-slip & thrust fault at base of section	amphibolite f. mylonite? retrograde amphibolite f. to greenschist f. mylonite & cataclasite	2400 - 2300 Ma. 1900 - 1700 Ma.

References: (1) Gibson et al., 1988; (2) Green et al., 1985; (3) Hubregtse, 1980; (4) Hurford, 1986; (5) Oliver & Coggon, 1979; (6) Percival & Card, 1985; (7) Percival & McGrath, 1986; (8) Percival et al., 1989; (9) Schmid et al., 1987; (10) Weber & Scoates, 1978; (11) Zingg et al., in press

TABLE II Uplift Characteristics of Some Crustal Sections in the Cores of Deeply Eroded Orogens

Crustal Section	Plate Tectonic Setting	Major Fault(s) active during Uplift	Fault Rocks & Metamorphic Conditions	Age of Uplift
Arunta (3, 6, 7)	Intracratonic: Arunta & Amadeus blocks (Central Australia)	Redbank fault: thrust at base of section & shear zones within nappes	amphibolite facies mylonites? retrograde amphibolite f. mylonite in hangingwall grading to greenschist f. mylonite & cataclasite in footwall	>1450-1100 Ma. 400 - 300 Ma. Alice Springs orogeny
Grenville Front (1, 2, 8)	Intracratonic: Grenville Province (Ontario, Canada)	Grenville Front tectonic zone: thrust at base of section & shear zones within section	granulite and amphibolite f. mylonite in hangingwall grading to greenschist f. mylonite & cataclasite in footwall	1400 - 1100 Ga. Grenville orogeny
Kasila (9, 10)	Plate Margin: W. African craton/Guayana shield (Sierra Leone, Africa)	Todi shear zone: thrust at base of section & shear zones within section	granulite and amphibolite f. mylonite in hangingwall grading to greenschist f. mylonite & cataclasite in footwall	2700 - 2200 Ma.
Central Gneiss Complex (4, 5)	Plate Margin: Alexander/Stikine terrains, (Coast Mountains, British Columbia, Canada)	thrusts & folds (reactivation of Todi s.z.) low-angle thrusts: shear zones within & at base of crustal imbricates	caataclasite granulite & amphibolite f. mylonite & synkinematic melt in hangingwall grading to greenschist f. mylonite & cataclasite in footwall	ca. 500 Ma. Rapid uplift: 98 to 90 Ma. 60 to 48 Ma.

References: (1) Davidson, 1988; (2) Green et al., 1988; (3) Goleby et al., (1989); (4) Hollister, 1982; (5) Hollister & Crawford, 1986; (6) Obee & White, 1985; (7) Teyssier, 1985; (8) Van Breemen & Davidson, 1988; (9) Williams, 1988, (10) Williams & Culver, 1988

detachment occurred at the base of the crust (unlikely in every case) or the crust was thicker and a major detachment horizon was located just below this range of depths within the crust.

Significant structural differences between crustal cross sections are related to their exhumation history. Cross sections that experienced a type 3 uplift history (Table II) were internally sheared and imbricated under high-grade metamorphic conditions during uplift and so are neither as complete nor as continuous as the cross sections with type 1 or 2 histories. Large thrust faults that bound all exposed crustal cross sections (Fountain and Salisbury, 1981) separate high-grade rock at the base of the cross section from lower grade rock outside of the cross section. In intracratonic uplifts, these lower grade rocks are lithological and metamorphic equivalents to the middle and upper crustal levels of the exposed cross section (e.g., Abitibi and Michipicotan belts separated by the Kapuskasing uplift; Percival and Card, 1985), whereas adjacent to crustal cross sections uplifted at or near plate boundaries these lower grade rocks sometimes have a tectonometamorphic history that contrasts with that of the exposed crustal section (e.g., the pre-Alpine metamorphosed Strona-Ceneri zone and the Alpine-metamorphosed Sesia zone separated by the Ivrea zone; Zingg et al., in press). In either case, the structural and metamorphic asymmetry across the boundary fault often coincides with paired gravity anomalies (Fig. 3 in Fountain and Salisbury, 1981). Gravity "highs" over the lower crustal base of the uplifted cross sections compliment gravity "lows" above the thickened crust. The highest positive Bouguer anomalies are measured above cross sections that underwent type 2 uplift (e.g., Ivrea geophysical body, Berckhemer, 1968; Fiordland section, Oliver and Coggan, 1979). Geophysical modelling of these large gravimetric and seismic anomalies suggests that some of the cross sections are underlain by large slices of mantle material at relatively shallow depths (ca. 10 to 20 km). In contrast, the cross section that experienced relatively cold intracratonic uplift (i.e., the Kapuskasing section) is underlain by thickened lower crust (Boland et al., 1988).

3. Kinematic and Rheologic Evolution during Uplift

3.1. COLD INTRACRATONIC UPLIFT

To date, the Kapuskasing section is the only documented example of cold intracratonic uplift that resulted in the exposure of lower crust at the surface. The deep crustal structure beneath the Kapuskasing section depicted in Figure 1a is based on the contoured velocity profiles of Boland et al. (1988) and on the extensive geologic work of John Percival and coworkers (references cited below). The uplifted crustal cross section is separated from underthrust lower crust by a SE-dipping wedge of lower crustal rocks (Fig. 1a). Large-scale imbrication and buckling of the lower crust isostatically balanced overthrusting of the Kapuskasing section, similar to the kinematic model of Percival et al. (1989) in which lower crustal material flows into a crustal root beneath the Kapuskasing section. Reflection seismic profiling indicates that the Ivanhoe Lake fault zone extends beneath the Kapuskasing section with a shallow (15 to 20°) dip to a detachment that originally lay at a depth of about 20 to 30 kilometers (Percival et al., 1989). Shortening above and below this intracrustal detachment is inferred to be equal (55 to 70 km; Percival et al., 1989) and occurred during the Early Proterozoic (Percival and Card, 1985; Percival and McGrath, 1986), possibly also during the Late Archean (Percival et al., 1989). The depth-temperature map in Figure 1b depicts the approximate geothermal conditions (thin lines) and P-T trajectories (arrows) of uplifted and underthrust lower crustal rocks in the Kapuskasing region. Isotopic mineral ages indicate that the Kapuskasing section cooled slowly from regional metamorphic conditions at about 2.7 Ga. (box in Fig. 1b) to below the 300 °C blocking temperature of the Rb-Sr biotite system

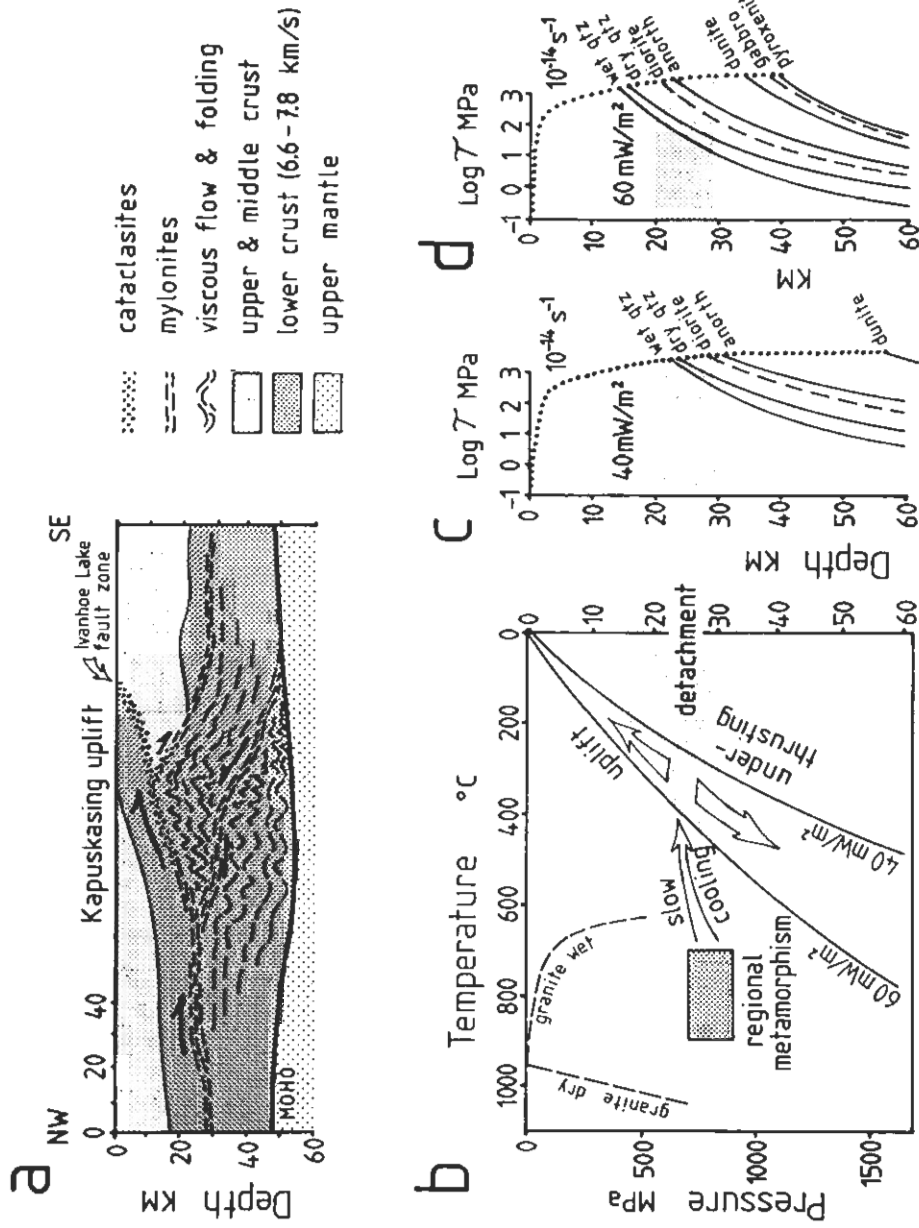


Figure 1. (a) Profile through the Kapuskasing uplift, Superior Province, Canada, modified from Percival et al. (1989); (b) Depth-temperature diagram showing conditions of regional metamorphism (Percival and Card, 1985) and predicted P-T paths of the Kapuskasing lower crust during uplift and underthrusting. Solidus curves for wet and dry granite from Johannes (1985), steady-state geotherms from Chapman (1986). (c) and (d) Log shear strength versus depth for steady-state geotherms in (b) at a regional shear strain-rate of 10^{-14} s^{-1} . Stippled area represents approximate depth of intracrustal detachment in (a) according to Percival et al. (1989). Dotted curves for frictional sliding (cataclasis) calculated with Byerlee's (1978) frictional coefficients. Solid curves are for viscous flow (dislocation creep) in wet and dry quartzite (Jaoul et al., 1984), anorthosite (Shelton and Tullis, 1981), and wet dunitite (Chopra and Paterson, 1981). Dashed curves are for diorite and gabbro (see text).

some 2.2 to 1.9 Ga. ago (Fig. 15 in Percival, 1988). The age and blocking temperature of biotite corresponds with the retrograde greenschist facies metamorphic grade of mylonites and cataclasites that overprint granulites in the Ivanhoe Lake fault zone at the base of the cross section (Percival and Card, 1985). Therefore, the average geothermal gradient had dropped from 30 to 40 °C/km during regional metamorphism to about 10 to 20 °C/km by the time of final uplift. Steady-state geotherms in Figure 1b corresponding to surface heat flows of 40 to 60 mW/m² (Chapman, 1986) approximate the average geothermal gradient during uplift inferred from isotopic ages and syntectonic metamorphic conditions in the Ivanhoe Lake fault rocks.

Lower crustal rocks that were initially adjacent on either side of the intracrustal detachment experienced divergent thermo-barometric and rheologic evolutions as intracratonic shortening progressed (Figs. 1b to 1d). The P-T paths of lower crustal rocks in Figure 1b are inferred to have followed the steady-state geotherms because two important observations indicate that thermal and isostatic equilibrium prevailed throughout the shortening history: (1) a foreland basin is conspicuously absent from the Kapuskasing section (Percival and Green, 1988); (2) there is no marked structural and metamorphic asymmetry within the Ivanhoe Lake fault rocks at the base of the crustal section. Together, these observations suggest that tectonic and erosional denudation kept pace with uplift and that uplift was sufficiently slow to allow thermal equilibration across the boundary fault during thrusting.

The implications of this P-T path for the evolution of rheologic properties during intracratonic thrusting becomes apparent in the shear strength versus depth diagrams in Figures 1c and 1d. Such diagrams (e.g., Brace and Kohlstedt, 1980) conveniently depict depth-dependent variations in rock strength associated with frictional sliding (i.e., cataclasis, dotted curves) and viscous solid-state creep (solid and dashed curves in Figs. 1c and 1d). Although the solid curves in Figures 1c and 1d are strictly valid for monomineralic aggregates, they adequately describe the viscous creep strength of polymineralic rocks at high strains provided that (1) they represent the weakest mineral in the rock, and (2) this mineral forms interconnected (i.e., contiguous) domains and makes up at least 20% of the rock (Handy, 1990). A minimum estimate of the creep strength of rocks with 80% or more of a relatively strong mineral (i.e., gabbros, diorites; dashed lines in Figs. 1c and 1d) is obtained with the nonlinear strength-composition relation for porous sintered aggregates (Tharp, 1983; see Handy, 1990). Due to the lack of reliable constraints on pore pressure, the effective pressure is simply assumed to equal the lithostatic pressure in Figures 1c and 1d, although it is recognized that pore pressure in hydrous shear zones can fluctuate extensively (Etheridge et al., 1984). Despite the problems of extrapolating laboratory flow laws to natural strain-rates (Paterson, 1987), the viscous strength relations predicted with the flow laws used to construct Figures 1c and 1d generally agree with the relative strength of lower crustal and upper mantle rocks inferred from strain measurements in the field (Handy and Zingg, in press) and so are a reliable qualitative guide to natural solid-state rheologies.

The intracrustal detachment at about 25 kilometers beneath the Kapuskasing section corresponds with two potential rheological discontinuities in Figures 1c and 1d: (1) the transition from frictional (i.e., brittle) to viscous deformation in quartz-bearing rock; (2) a change in the bulk composition of the lower crust from predominantly quartz- and feldspar-bearing paragneiss and diorite above the detachment (exposed in the Kapuskasing uplift) to quartz-free feldspathic, amphibolitic and gabbroic rocks below the detachment (in the thickened, underthrust part of the crust in Fig. 1a). The latter interpretation is consistent with the velocity structure of the Kapuskasing region (Fig. 4 in Boland et al., 1988) indicating that lower crustal rocks with P-wave velocities greater than 7.0 km/s predominate below 25 to 30 kilometers depth.

Final uplift of the cold crustal cross section occurred largely in the frictional regime (Fig. 1c), as shown by the ubiquity of cataclastic fault rocks in the Kapuskasing section and along the

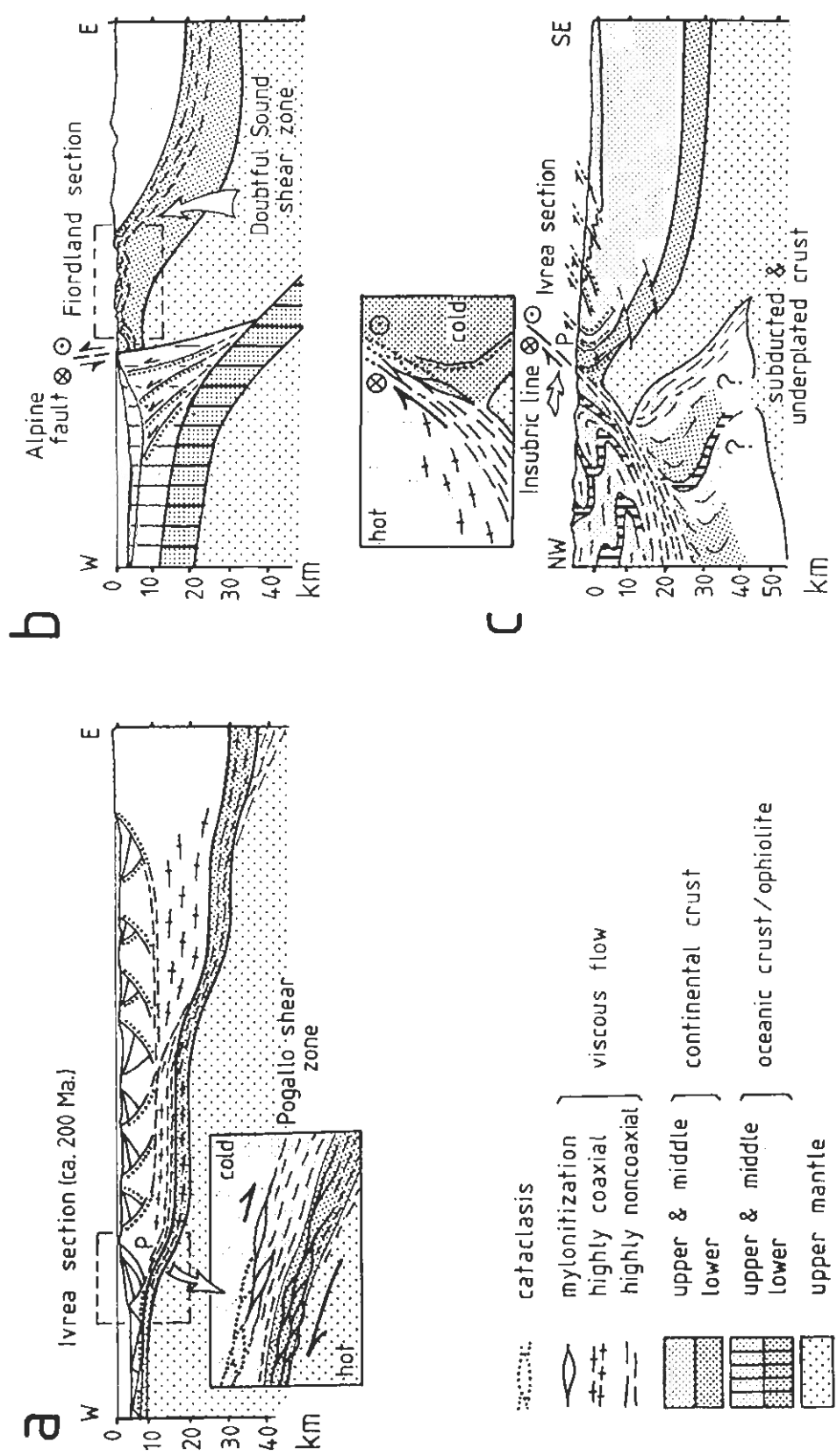


Figure 2. Exhumation involving hot uplift during extensional crustal delamination of a passive margin (a), followed by cold uplift near an ocean-continent active margin (b), and/or near a continent-continent active margin (c). (a) Reconstruction of the Early Mesozoic Apulian continental margin showing the approximate location of the Ivrea section in the dashed box (modified from Handy and Zingg, in press); (b) Present setting of the Fiordland section (dashed box), South Island, New Zealand (modified from Oliver and Coggon, 1979; Gibson et al., 1988); (c) Present setting of the Ivrea section south of the main part of the Alpine orogen (modified from Zingg et al., in press). Insets to (a) and (c) show distribution of fault rocks, respectively, in the Ivrea zone and at the Insubric line. P marks location of the Pogallo shear zone.

Ivanhoe Lake fault zone (Percival and Card, 1985; Percival and McGrath, 1986). Deformation within the underthrust lower crust is predicted to have become localized within discrete mylonitic shear zones after relatively minor amounts of strain. Shearing of the basic granulite facies rocks under retrograde metamorphic conditions (300 - 600 °C according to Fig. 1b) would have facilitated the syntectonic growth of fine grained and/or weaker minerals within the shear zones. The viscous strength of such fine grained reaction products is much less than that of the original basic host rocks plotted in Figures 1c and 1d (e.g., Brodie and Rutter, 1985), so that narrow shear zones probably accommodated most of the strain in the underthrust lower crust. In addition, the high viscosity contrasts associated with both localized grain size reduction in shear zones and the low homologous temperatures ($T_h < 0.4$) in the underthrust lower crustal rocks may have led to extensive buckling within the thickened lower crustal root underlying the Kapuskasing section (Fig. 1a). Temperatures in the underthrust rocks increased somewhat during shortening (Fig. 1b), but this increase was probably not sufficiently large to induce significant changes in flow regime and structure.

An important feature of this model is that the brittle deformed and uplifted crustal cross section preserves its general structural integrity whereas the warmer underthrust and wedged lower crust is predicted to lose its structural integrity through imbrication and viscous folding. This may explain the lack of strong reflectors in the crustal root underlying the Kapuskasing uplift (Fig. 2 in Percival et al., 1989).

3.2. EXTENSIONAL-CONVERGENT UPLIFT

The multistage uplift history and structural evolution of most crustal cross sections reflects a much more complicated interplay of kinematic, geothermal and rheologic factors than in the intracratonic Kapuskasing uplift. The Ivrea section (southern Alps of Europe) and the Fiordland section (Southern Alps, New Zealand) are two excellent examples of crustal cross sections that experienced early extensional uplift (Fig. 2a) followed by uplift either at an active ocean-continent margin (Fig. 2b) or near a continent-continent collisional boundary (Fig. 2c).

Extensional uplift of the crustal cross section to shallower depths (10-20 km) involved lithospheric shearing at very high transient geothermal gradients and broadly coincided with basin development and subsidence of the rifted upper crust (Fig. 2a). In the case of the Ivrea section, this extensional exhumation was probably polyphasic (Handy and Zingg, in press). The metamorphic core complexes of the American Cordillera represent possible precursors to the passive margin stage depicted in Figure 2a. Subsequent exposure of the attenuated cross section at or near a convergent plate boundary occurred after the crustal section had cooled to temperatures similar to or less than that of the crust with which it was eventually juxtaposed at the surface. As with the extensional stage of exhumation, convergent uplift of the crustal section may be polyphasic. The oblique convergent uplift of the Fiordland section shown schematically in Figure 2b is similar to the early (i.e., Eo-Alpine) episode of convergent uplift of the Ivrea section. At this time, the distal rim of the southern Alpine continental margin (Sesia zone) was partly subducted beneath the mantle and crust of the proximal part of the margin (i.e., the part containing the Ivrea section; Schmid et al., 1987), while the overriding part of the passive margin was uplifted. This early configuration may have resembled Figure 2b and preceded the final uplift configuration in Figure 2c. Thus, Figures 2a through 2c can be regarded as representing a complete evolutionary sequence of exhumation, beginning with a metamorphic core complex or a passive margin extensional stage, progressing through an ocean-continent collisional ("Fiordland") stage, and in some cases ending with a continent-continent collisional ("Ivrea") stage.

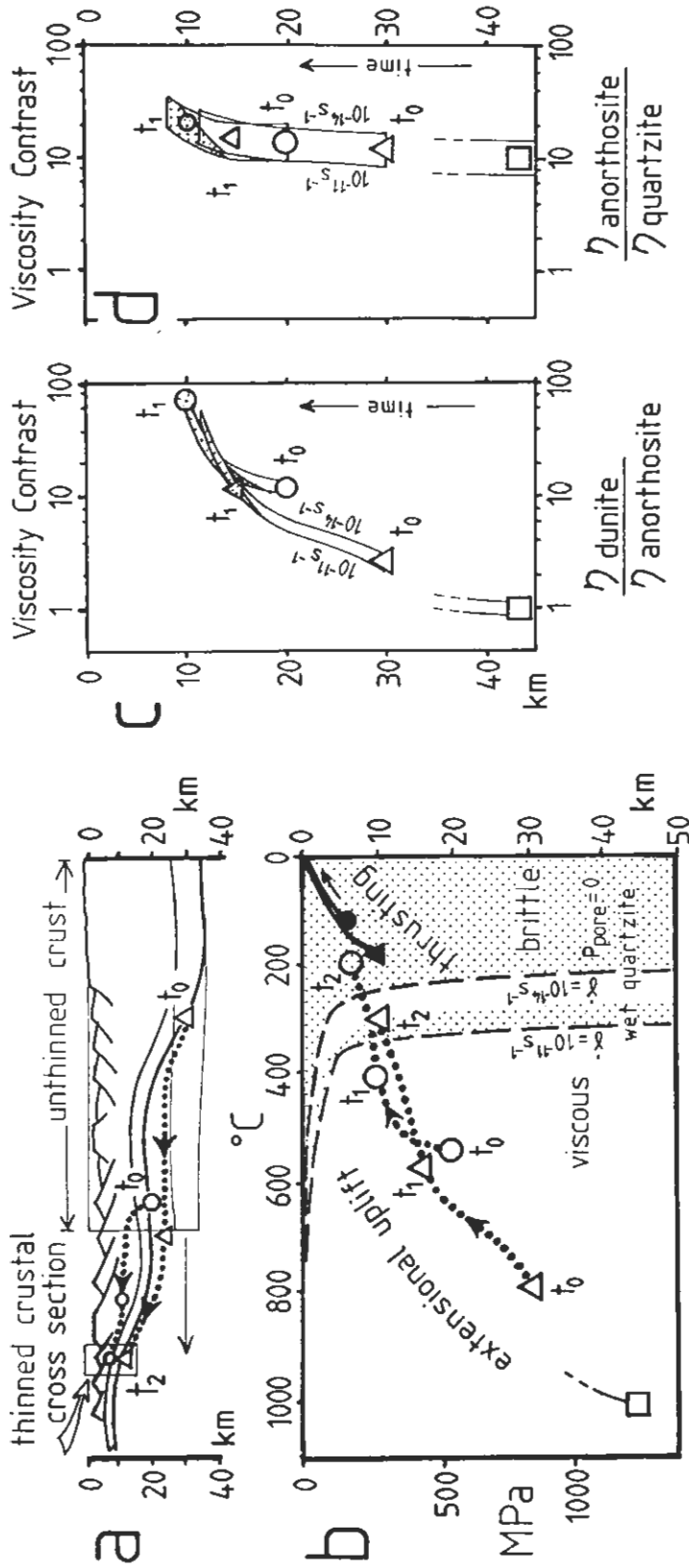


Figure 3. (a) Passive margin showing hypothetical particle paths during extensional uplift of lower and intermediate crustal levels (respectively, triangles and circles) between times t_0 at the onset of extension and the final position of the thinned crustal cross section at t_2 . A major noncoaxial extensional shear zone separates the two crustal levels, as shown in Fig. 2a. (b) Depth-temperature diagram with hypothetical P-T paths for crustal levels shown in (a) during extension (dotted lines) and cold thrusting (solid lines). Dashed curves for transition from frictional (stippled region) to viscous flow in hydrous quartzite at shear strain-rates of 10^{-11} and 10^{-14} s^{-1} (taken from Handy, 1989). (c) and (d) Viscosity contrast versus depth diagrams, respectively, for dunite-anorthosite and anorthosite-quartzite during extensional uplift of the two crustal levels symbolized in (a) and (b). Squares in (b), (c), and (d) indicate conditions in hypothetically deeper crust (see text). Creep parameters as in Fig. 1 above.

The faults at the base of the exposed crustal cross sections dip either toward the section (Fig. 2b; e.g., the Alpine fault west of the Fiordland Section; Oliver and Coggon, 1979) or away from the section (Fig. 2c; e.g., the Insubric line northwest of the Ivrea section; Schmid et al., 1987). The opposite fault dip directions probably reflect differences in the isostatic behaviour of continental and oceanic crust during different stages of plate collision. During early ocean-continent collision, oceanic crust subducts without substantial thickening along a master fault dipping beneath the crustal cross section. In contrast, continental crust is more buoyant and thickens during collision. If convergence and thickening rates are high compared to the rates of tectonic and erosional denudation, then continental crust eventually overthickens and escapes upwards and outwards. During the late stages of the Alpine orogeny, this escape was primarily in the direction of subduction and resulted in both south-vergent refolding ("backfolding") of the north-vergent Alpine basement nappes and rapid uplift concentrated along the Insubric line (Fig. 2c). The Insubric line formed at the northern rim of the Ivrea geophysical body because the cold, creep-resistant rocks of the previously uplifted upper mantle and lower crust acted like a rigid indenter within the thick pile of warm, relatively weak intermediate crust forming the core of the Alpine orogen. The Insubric line is therefore a comparatively young orogenic structure and is interpreted to flatten at depth within the crust, truncating the older SE-dipping master fault that was active during Eo-Alpine subduction (Fig. 2c).

The model of lithospheric attenuation and exhumation in Figures 2a and 3a involves stepwise crustal delamination along large zones of noncoaxial flow (like the Pogallo shear zone in the Ivrea section) that accommodate inhomogeneous pure shear extension of the entire lithosphere (e.g., Lister et al., 1986). The predicted pattern of shearing at the base of the extending crust is shown with the dashed (mylonitic) and dotted (cataclastic) zones in Figures 2a and 3a. Large extensional shear zones nucleate at major pre-existing compositional boundaries and then coalesce, stepping upwards towards the more thinned, distal part of the passive margin. The master fault is highly noncoaxial and dips towards the continent where it crosses compositional boundaries and drags the underlying lithosphere upwards and outwards. In the absence of well-constrained pressure - temperature - time (P-T-t) paths in deeply eroded extensional terrains, the extensional P-T paths for intermediate and lower crustal levels in Figure 3b (dotted curves) have been constructed conceptually for the lithospheric extensional model described above and depicted in Figure 3a. The corresponding particle paths for these crustal levels between times t_0 and t_2 within the evolving passive margin are indicated in Figure 3a. The predicted sinuosity of the extensional P-T paths reflects the passage of attenuating crustal material through highly noncoaxial, low-angle shear zones at crustal "necks" and through areas of more coaxial shear and passive uplift (Fig. 3a). The juxtaposition of disparate crustal levels along low-angle normal faults is associated with higher rates of thermal equilibration per amount of uplift than is homogeneous coaxial (pure) shear (Ruppel et al., 1988). Therefore, periods of highly noncoaxial shear in Figure 3a are depicted with the less arcuate and less steep parts of the P-T paths in Figure 3b than are periods of highly coaxial shear.

The overall concave upward arcuation of the P-T paths in Figure 3b reflects the fact that the rate of uplift is generally high with respect to the rate of thermal equilibration. Direct evidence in the field for tectonically induced thermal disequilibrium during uplift comes from the pronounced structural and metamorphic asymmetry of large shear zones in many exposed crustal cross sections (Table I). Cataclasite and low-grade mylonite at the cool contact between differentially uplifted crustal levels grade into high-grade mylonitic rock within the warmer uplifted footwall. Extensional shear zones that have exhumed hot lower crust are vertically zoned (inset to Fig. 2a; the Pogallo shear zone, Handy, 1987; the Doubtful Sound detachment, Gibson et al., 1988) whereas some steeply dipping, late-stage faults bounding several uplifted crustal cross sections are

subhorizontally zoned and reflect high lateral heat flux during rapid uplift and shearing (e.g., the Insubric line; inset to Fig. 2c).

An important consequence of inhomogeneous thinning is that different levels of the crust pass through highly noncoaxial shear zones at different times as the lithosphere attenuates. Uplift rates for different crustal levels vary, so pieces of crust that are vertically juxtaposed at time t_2 in the distal parts of the passive margin were not part of a vertically continuous crustal cross section at the onset of extension at time t_0 (particle paths in Fig. 3a). The point emphasized here is that any extension involving a component of noncoaxial flow produces cross sections of attenuated crust whose constituent levels occupied laterally different positions prior to extension.

The long-term rheological characteristics of rocks in the extending lithosphere can be approximated by tracking the temperature and strain-rate dependent changes in the relative strength of rocks (Figs. 3c and 3d) between times t_0 and t_2 on the P-T paths in Figure 3b. The curves in Figures 3c and 3d yield maximum estimates of strength contrast because the assumption of a constant regional strain-rate for all lithologies leads to an underestimation of strain-rate in the weak rocks and an overestimation of strain-rate in the strong rocks (see Handy, 1990). In addition, stress concentration at lithological contacts can cause a localized reduction in grain size and therefore also a strain-dependent decrease in the viscous strength of the relatively stronger rock with respect to that of the weaker rock (Rutter and Brodie, 1988). Nevertheless, Figures 3c and 3d demonstrate two important points: (1) decreasing temperature during uplift is associated with increased viscous strength contrast among the principle rock-forming minerals in the lithosphere; (2) strain localization at litho-rheological boundaries depends on the pre-extensional compositional and metamorphic characteristics of the lithosphere. For example, at an initial depth of 45 kilometers, an upper mantle rock comprising at least 20% olivine is predicted to have about the same viscous strength as a lower crustal rock comprising at least 20% feldspar (square symbol in Figs. 3b-3d). With decreasing temperature, olivine rock becomes progressively stronger with respect to feldspathic rock (Fig. 3c). This suggests that at high geothermal gradients in initially thick crust (> 45 km according to the flow laws used in Fig. 3c) the MOHO only becomes a significant rheological discontinuity during the latter part of extensional uplift, whereas in thinner crust (< 45 km) the MOHO is already a potential rheological discontinuity at the onset of extension. Under the same conditions, the predicted strength contrast between quartzitic and feldspathic rocks is not as temperature dependent as that between olivine and feldspathic rocks. The considerable magnitude of this strength contrast ($>10:1$ in Fig. 3d) is such that the intracrustal boundary between quartz-bearing and quartz-free feldspathic compositional domains is predicted to remain an important rheological discontinuity throughout the extensional evolution.

Viscous instability leading to strain localization develops from inhomogeneities in the rock that either predate deformation (inherited structure and composition) or are deformationally induced (syntectonic reactions, thermal perturbations). Once strain has localized, the increased stress concentration associated with decreasing temperature during uplift also leads to grain size reduction within shear zones and so results in decreased viscosity of the shear zones with respect to the viscosity of the less sheared or unsheared country rock (Handy, 1989). Therefore, strain is progressively localized during uplift, because a localized decrease in viscosity reduces the volume of deforming rock needed to accommodate a given regional strain-rate. For a given strain-rate and grain size, strain localizes at higher temperatures (and therefore, at greater depths) in feldspar, amphibole, and olivine rocks than in quartz-bearing rock. This is observed in the Ivrea section (inset to Fig. 2a), where crustal-scale extension is accommodated within high-temperature mylonitic zones in the lower crust (e.g., Brodie and Rutter, 1987) and within a kilometer-wide zone of predominantly noncoaxial shear at the base of the quartz-rich intermediate crust (Handy, 1987). Anastomosing cataclastic and mylonitic shear zones grade downwards into broader zones of

penetrative viscous shear at the originally deeper end of the crustal cross section (Handy and Zingg, in press).

Stress and strain partition not only on the lithospheric scale, but also on the outcrop and even on the microscopic scale. An important consequence of stress and strain partitioning during uplift is that large portions of the crust between shear zones remain undeformed or only weakly strained and therefore preserve the pre-extensional features of the rock. What makes extension (as opposed to collision) such an important exhumation mechanism is that intracrustal detachment along low-angle normal faults does not significantly change the depth-sequence of structural levels during uplift. While extensional detachments accommodate large lateral displacements of crust during thinning, a significant proportion of the vertical displacement results from passive uplift of all crustal levels as buoyancy forces act on the warm thinning continental lithosphere (Bott, 1981). However, it is important to note that extension on its own is not a viable mechanism for exposing large tracts of the deepest crust and upper mantle at the surface for long periods of time. Firstly, magmatic underplating related to decompression melting or incipient seafloor spreading can pre-empt large-scale uplift of the MOHO to the surface at very high extensional strains ($\beta > 3$ to 4; Le Pichon and Sibuet, 1981). Secondly, the tensional regional stress field and low crustal viscosities associated with lithospheric extension at high geothermal gradients are not suitable for supporting large volumes of dense, isostatically unstable material at the surface for long.

Final uplift and exposure of attenuated crustal cross sections is driven by compressional forces from a nearby convergent plate boundary (Figs. 2b and 2c). If the subducted crust is continental rather than oceanic, then the buoyancy of this crust exerts an additional upward force on the overlying column of rock, including the crustal cross section. In Figure 2c, the dipping sliver of mantle material responsible for the large positive Ivrea gravity anomaly was trapped between the overlying and subducted continental crust of the passive margin. Note that no large-scale obduction of mantle into the crust was necessary to generate this structure (compare with Laubscher and Bernoulli, 1982). At the surface, thrusts and folds associated with the latest stages of collision have the same vergence as subduction at the convergent boundary (Fig. 2c). Final emplacement of the crustal cross sections under low-temperature conditions is characterized by large overall crustal strengths. Shearing in most parts of the crustal section is extremely localized and involves cataclasis and brittle folding (solid P-T paths in the stippled frictional regime in Fig. 3b), as observed in the Ivrea zone (Zingg et al., in press).

4. Other Mechanisms for Exposing Sections of the Continental Crust

The crustal sections listed in Table II all come from the crystalline cores of deeply eroded orogens. The tectonic setting and history of these sections vary considerably, and in most cases, the details of their uplift history remain ambiguous pending further investigation. What distinguishes all of them from the crustal cross sections discussed so far is the high degree of internal deformation that accompanied exhumation. For this reason, none of these deep crustal exposures is coherent or complete in the sense defined in the Introduction.

4.1. HIGH-GRADE THRUSTING AND UPLIFT

Uplift under high-grade metamorphic conditions involved penetrative shearing and considerable imbrication in all structural levels of the deeply eroded Precambrian crustal sections listed in Table II. This internal shearing is temporally and kinematically related to high-grade thrusting at the base of the crustal sections which puts the higher grade lower crust above lower grade rocks of the

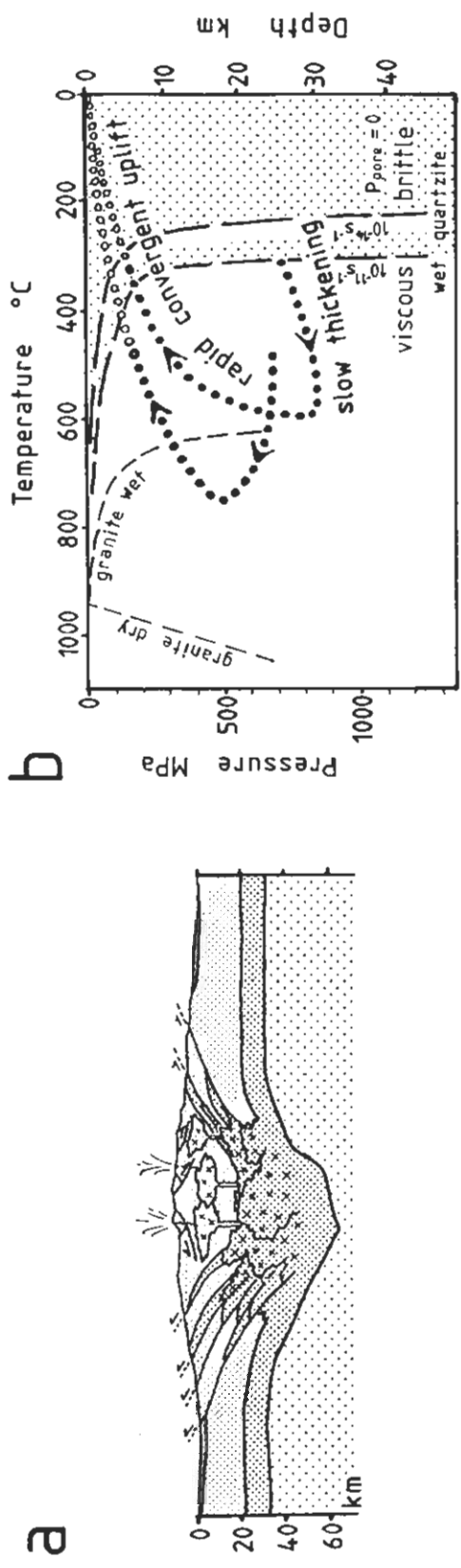


Figure 4. (a) Schematic profile through an oblique collisional magmatic arc based on Hollister and Crawford (1986); (b) Depth versus temperature diagram with P-T paths for two rapidly uplifted deep crustal segments in the Central Gneiss Complex, Coast Mountains, British Columbia (Hollister and Crawford, 1986). Open dots are extrapolated parts of P-T paths. Frictional to viscous transition (thick dashed lines) and granite solidus curves (thin dashed lines) as in Figs. 1b and 3b.

intermediate to upper crust. Thermal disequilibrium during thrusting is reflected in the asymmetrical zonation of fault rocks at the base of the cross section, with high-grade mylonite in the hot hanging wall grading to low-grade mylonite and cataclasite in the colder footwall (Table II). The sense of asymmetry of this zonation is therefore opposite to the asymmetry of fault rock zonation described above for extensional detachments. The observation of two or more generations of tectonites in some of these large shear zones (e.g., Redbank fault zone, Arunta Section; Obee and White, 1985) suggests that uplift of the crustal cross section involved several episodes of thrusting.

Reflection seismic profiling reveals that the high-grade thrusts bounding some of these crustal cross sections can be traced to the base of the crust (e.g., Grenville Front tectonic zone, Green et al., 1988) and are sometimes interpreted to cross, or even to displace the MOHO (e.g., Arunta Section, Goleby et al., 1989; for contrasting interpretation, see Teyssier, 1985). This raises the possibility that imbrication and uplift of the lower crust to the surface are directly connected to deep-seated shear zones within the mantle. An immediate implication of discrete faults that offset the MOHO is that the viscosity contrast between lower crustal and upper mantle rocks is much smaller than predicted from the extrapolation of laboratory creep laws for feldspar- and olivine aggregates to temperatures along typical orogenic geotherms (Carter and Tsenm, 1987).

4.2. MELT-ENHANCED THRUSTING AND UPLIFT:

Crustal cross sections from Cordilleran terrains of North America show substantial magmatism and crustal reconstitution associated with severe internal imbrication during oblique convergence and uplift. Hollister et al. (1989) have reconstructed their evolution from fragments of various crustal levels exposed in the Coast Mountains (western Canada; Table II) and the Idaho and Sierra Nevada batholiths (western United States). Figure 4a is a schematic profile through such an oblique convergent magmatic orogen, based on the work of Hollister and coworkers. Crustal shortening, metamorphism, magmatism, and uplift are broadly coeval and the interaction of these processes can be described in terms of a negative feedback cycle. The P-T paths of two deep crustal imbricates during two such cycles in the Central Gneiss Complex near Prince Rupert, British Columbia (Hollister and Crawford, 1986) are shown in Figure 4b. Collision leads to crustal shortening and thickening, burial metamorphism, and both mantle and crustal anatexis. These melts lubricate high- and low-angle faults that accommodate, respectively, rapid vertical and outward escape of deeper orogenic levels. Granitic intrusions in mid- to upper crustal levels may be associated with extension of the overthickened orogen (Hollister et al., 1989). Upward magma transport in such batholithic intrusions is also linked to localized viscous (mylonitic) downflow of the country rocks adjacent to the batholith (Saleeby, 1988). Decreasing temperatures during uplift freeze the magma, stop or slow the crustal uplift, and re-initiate a period of slower crustal thickening. Exhumation in these orogenic settings occurs during short periods or "surges" of very rapid uplift (1 to 2 mm/yr; Hollister, 1982) along melt-lubricated shear zones in the deep crust that grade upwards to mylonites at higher crustal levels (Hollister and Crawford, 1986). At the surface, sheets of higher grade rocks are emplaced over lower grade rocks (Fig. 4a).

This melt-enhanced deformational style contrasts with the solid-state creep deformation effecting exhumation of most other crustal cross sections. Partial melting is usually inferred to have a drastic weakening effect on the bulk strength of a rock (Arzi, 1978), but the magnitude of this strength drop as well as the long-term strength of partially melted rock depend strongly on the viscosity contrast between solid rock and melt, and on the amount and distribution of this melt within the deforming aggregate. If hot magma is injected into a significantly cooler rock, then the initial deformation may be brittle, while the magma intrudes rapidly along cracks and exerts a pore pressure that reduces the effective strength of the rock to levels favoring further fracture and

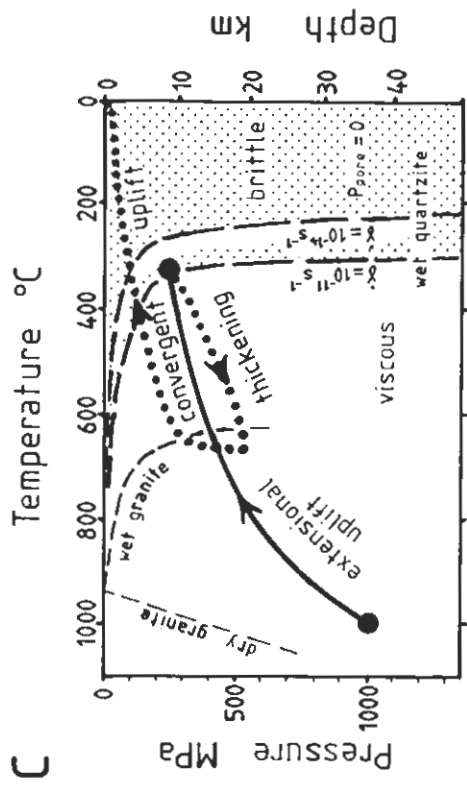
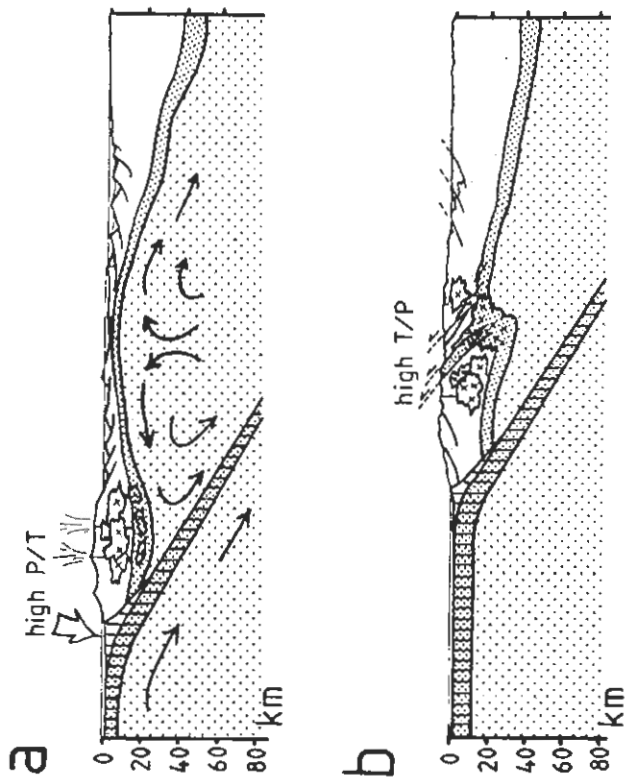


Figure 5. Extensional uplift of hot lower crust in a back-arc spreading environment (a) followed by imbrication and partial magmatic reconstitution during crustal thickening in (b). Hypothetical P-T evolution of the lower crust during extensional uplift (solid line) and shortening, thickening and uplift (dotted lines) appears in the depth-temperature diagram in (c). Brittle (i.e. frictional) to viscous transition in hydrous quartzite and solidus curves for granite as in previous figures.

dilation. After high strains or during in-situ synkinematic anatexis, however, the solid rock either comprises a viscously deforming framework containing lenticular layers of melt, or sufficient melt is present (> 20%; van der Molen and Paterson, 1979) to accommodate bulk strain within contiguous layers. These melt layers generally define a foliation parallel to the plane of shear and wrap about deformed lozenges of host rock (or restite; Fig. 1b, Hollister and Crawford, 1986). The mechanical role of the melt is not just primary (i.e., that of a weak, strain-accommodating phase), it is also secondary inasmuch as the incipient melt forms thin films along grain boundaries that act like diffusive corridors and so enhances viscous grain boundary sliding in the host rock (Cooper and Kohlstedt, 1987). Thus, it is the combination of these effects that ultimately leads to a drop in flow strength coupled with an increase in strain-rate during syntectonic partial melting.

Generally, the volume proportion of sheared rock at a given strain-rate is inversely and nonlinearly proportional to the viscosity contrast between sheared (or melted) and unsheared rock. Shear zones are widest at deformational conditions above the solidus, where the viscosity contrasts both between rock and melt and among the constituent minerals in the rock are low. As partially melted middle and lower crustal rocks pass through the solidus on their way to the surface (Fig. 4b), shearing affects much of whatever original structure managed to survive magmatic reconstitution at depth. Therefore, rapid melt-enhanced uplift tends to preserve less of the pre-exhumational structure and metamorphism than do the other uplift mechanisms discussed above.

Other tectonic scenarios can be envisioned which exhume incomplete, dismembered sections of the continental crust in orogenic settings. In Figure 5a, for example, an early stage of subduction and crustal accretion is associated with granulite facies metamorphism and magmatism at the base of the continental crust. Localized instabilities in mantle convection lead to back-arc extension and roll-back of the oceanic subduction zone. Granulites at the bottom of the attenuating crust in the back-arc region are extensionally uplifted to midcrustal levels (Fig. 5a, solid P-T path in Fig. 5c). Exposure of the granulites at the surface results from thrusting and uplift while the thinned crust is still hot (dotted P-T path in Fig. 5c) or after this crust has cooled. Hot thrusting is associated with considerable internal imbrication and shearing of the uplifting crust under high-grade conditions (Fig. 5b), whereas cold thrusting and uplift might resemble the final emplacement stage depicted in Figure 2b. A similar exhumation mechanism (e.g., Fig. 2a in Fountain and Salisbury, 1981) may explain the pairing of metamorphic belts in many circum-Pacific oblique convergent margins, where the successive collision of island arcs resulted in the emplacement of rocks with high-T/P assemblages onto subducted and rapidly exhumed rocks with older high-P/T mineral assemblages (e.g., Hidaka crustal cross section, Hokkaido, North Japan; Komatsu et al., 1983).

5. The Exhumation of Crustal Cross Sections: A Selective Process

What part of the Earth's crust do we actually see in exposed crustal cross sections? A fundamental point to emerge in this paper is that crustal exhumation reveals certain types and features of the deep crust better than others. Moreover, different exhumation mechanisms preserve different stages of crustal evolution. This has implications for the question of how representative exposed crustal cross sections are of the unexposed continental crust. Intracratonic crustal sections emplaced as relatively cold wedges (the Kapuskasing section) best represent that part of the craton which lies above a deep intracrustal detachment. The unexposed crust below the detachment probably has different compositional and metamorphic characteristics. Crustal cross sections that were attenuated prior to final convergent emplacement are obviously good examples of thinned continental crust. However, the crustal levels that presently constitute these thinned crustal sections did not necessarily occupy the same vertical section of crust prior to the extensional stage of exhumation.

Finally, cross sections of crust that were uplifted in orogens at high geothermal gradients are rarely complete and are often internally imbricated and/or magmatically reconstituted under high-grade metamorphic conditions. Such exposures yield abundant information about crustal reconstitution and deformation in the core of an orogen, but preserve relatively little information about the crustal evolution prior to uplift.

What sets true crustal cross sections apart from most deeply eroded orogenic exposures is the high degree of structural and metamorphic coherence maintained during uplift and emplacement to the surface. Compressional and isostatic forces associated with the anomalous lithospheric structure beneath exposed crustal cross sections balance dense lower crust near the surface over long periods of time. It is not fortuitous that the bottom of all exposed crustal cross sections examined so far originally occupied depths no greater than about 45 kilometers. In some cases, this depth represents the thickness of the crust prior to extensional uplift, whereas in other cases it was the depth of basal intracrustal detachment and thrusting at the onset of convergent uplift. Original deep crustal features in crustal cross sections survived exhumation because decreasing temperatures during uplift led to increased viscous strength contrasts among different compositional domains and hence to progressive stress and strain localization within narrow shear zones.

In conclusion, the exhumation of coherent cross sections of the continental crust is a very selective, usually multistage process involving special kinematic and geothermal evolutions. The sampling of the deep crust exposed at the surface is therefore only as complete as the environments conducive to its uplift are varied.

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