FAULT ROCKS AND STRUCTURE AS INDICATORS OF SHALLOW EARTHQUAKE SOURCE PROCESSES

by

RICHARD H. SIBSON

Department of Geology
Imperial College
London, SW7 2BP
U. K.

Abstract: Rock deformation textures and structures found in and around fault zones are a powerful potential source of information on the earthquake mechanism. In particular, deeply exhumed ancient fault zones and those with a large finite component of reverse dip-slip may provide information on the macroscopic fault mechanisms and associated processes of mineral deformation which occur at depth. One major task is to identify with which parts of the earthquake stress cycle particular features are related.

A methodology for building up a conceptual model of a major fault zone in quartzo-feldspathic crust is illustrated mainly by reference to field-based studies on the Outer Hebrides Thrust in Scotland, and the Alpine Fault In New Zealand. The shortcomings of the method, and some of the main unanswered questions are discussed.

(1) INTRODUCTION

Standard techniques for investigating the mechanism of earthquake faulting include the analysis of seismic radiation, geodetic observations around fault zones, consideration of theoretical faulting models, and laboratory experiments on the deformation and frictional sliding characteristics of rock specimens. There are shortcomings inherent to all these methods. For example, teleseismic observations yield information on the changing stress state in the source region but not the absolute stress values, the interpretation of geodetic observations and construction of theoretical faulting models necessarily involve gross simplification and idealisation of crustal structure and properties less the mathematics become over-complex, and there are severe scaling problems in relating short-term deformation experiments involving minute displacements in small homogeneous specimens to the natural environment. All this at a time when the essential heterogeneity of fault zones and the faulting process is becoming more and more apparent (e.g. Aki, 1978).

The structures and deformation textures of rocks found in fault zones form a tangible complementary source of information on faulting, as yet largely untapped. In particular, we should be able to increase our understanding of deep-level processes by studying the rock products of faulting associated with ancient, deeply exhumed fault zones, or those with a large finite component of reverse slip.

This paper discusses the application of this procedure to two major fault zones developed in quartzo-feldspathic crust, the Outer Hebrides Thrust in N.W. Scotland and the Alpine Fault in the South Island of New Zealand. By comparison and contrast of these structures, features of local and more general significance may be distinguished. Our methodology has two aims, one general and the other more specific.

(1.1) Conceptual Models of Fault Zones

Our general aim is to understand the change with depth in the crust of macroscopic faulting mechanisms (e.g. slip on discrete planes, shear across zones of finite width) and associated mineral deformation mechanisms, for both seismic and aseismic slip modes. The basic approach (Fig. 1) is to relate particular styles of faulting to mineral deformation mechanisms and metamorphic environments. Information of this kind may be gathered from a number of ancient fault zones exposed at different erosion levels, or from reverse-slip dislocations which have associated with them rock products generated over a range of depths, and integrated to build up a conceptual model of a major fault zone. To date, our attention has mainly been focused on fault zones developed in crystalline quartzo-feldspathic crust.

Fault models established in this way can to some extent be directly tested if current proposals for deep exploratory drilling in fault zones are put into effect.

(1.2) Structures and Microstructures in Relation to the Earthquake Cycle

Earthquake faulting is believed to be a cyclic process in terms of stress accumulation and release. Thus if information specific to the earthquake mechanism is to be obtained from within the general framework of our fault zone model, it is axiomatic that the various structural/microstructural features observed must be related to particular phases of the earthquake cycle.

In general we can envisage two basic types of stress/time regime in fault zones (Fig.2). More or less steady-state, aseismic displacements at all crustal levels should occur at constant or gently fluctuating levels of shear stress. In contrast, during the earthquake faulting cycle induced by some form of stick-slip process, shear stress may roughly follow a saw-tooth oscillation. There is no direct evidence, however, that the rate of stress accumulation stays constant while the fault is locked and in a more general approach, following Lensen (1971), the cycle for a major shallow earthquake can be divided into four phases.

These are; an a-phase of secular, mainly elastic strain accumulation; a 8-phase of precursory anelastic deformation including dilatancy, and possibly terminating in foreshock activity and accelerating fault creep; a y-phase involving actual seismic slip and rupture propagation; and a δ-phase including decelerating afterslip and aftershock activity. Clearly, structural features related to the β and y-phases in particular, may contribute significantly to our understanding of the shallow earthquake mechanism. For example, 8-phase structures may provide an explanation for observed precursory phenomena, while textures and structures developed during the γ-phase may yield valuable information on energy dissipation in fault zones during seismic slip, and thus on the efficiency of the faulting process (Sibson 1975, 1977b).

However, a major interpretative difficulty arises directly from the cyclical nature of earthquake faulting; structural features developed in one phase of one particular cycle are liable to be overprinted or obliterated by features developed either in later phases of the same cycle, or in succeeding cycles. The seriousness of this problem becomes apparent when one considers that a finite displacement of, say 10 km, in a major fault zone has probably developed by the accumulation of around 10^4 earthquake slip increments. Thus in some respects, features related to particular phases of the earthquake cycle may be easier to distinguish around faults with comparatively low finite displacement.

(1.3) Terminology

Throughout this paper the fault rock nomenclature generally follows that put forward by Sibson (1977a) (Fig.3), though some particularly descriptive terms (e.g. 'curly' schist-mylonite) applied to Alpine Fault material by Reed (1964) are retained because of widespread local usage.

(2) TWO MAJOR FAULT ZONES

In this section, variations in the style and the rock products of faulting are described for cross-strike traverses through two major fault zones. One structure, the Outer Hebrides Thrust, is an ancient feature now deeply exhumed, while the other, the Alpine Fault, remains active today. Both structures are developed mainly in crystalline quartzo-feldspathic crust and have large finite components of reverse dip-slip. Thus in passing from the downthrown to the upthrown sides of the fault zones, one can expect to find fault rocks generated over a range of depths. While the two structures have several basic features in common, there are also some major differences.

(2.1) The Outer Hebrides Thrust, N.W. Scotland

(2.1.1.) Background: With a sheet-dip of about 25° ESE, the Outer Hebrides Thrust is a complex dislocation zone which disrupts the crystalline Lewisian basement gneisses of the Outer Isles in N.W. Scotland (Fig.4). The thrust zone can be traced for some 190 km, bordering the eastern coastline of the archipelago, and must rank as a major tectonic element in the fabric of the British Isles. Parallelism with the Moine Thrust some 85 km to the east, coupled with radiometric ages obtained from the fault rocks, suggest a Late Caledonian (c.400 Ma) age for the thrust movements.

The host Lewisian Complex is largely a high-grade amphibolite facies assemblage, variably migmatised and consisting mainly of biotite-hornblende, quartzo-feldspathic banded gneisses with a varying amphibolite content (Coward et al. 1969). Large bodies of anorthosite, meta-gabbro and meta-tonalite occur locally, and pyroxene granulite assemblages are developed from place to place. Large Laxfordian (c. 1800 Ma) fold structures have imposed a general NW-SE structural grain, which is cut across by the thrust zone, (Coward et al. 1970).

Structural features indicate a general ESE-WNW direction of translation for the main thrust phase, but the lack of marker units makes it difficult to estimate the displacement across the thrust zone. However, indirect evidence suggests a total finite reverse slip of 10-15 km, giving rise to a vertical displacement of 4-6 km (Sibson. 1976).

(2.1.2) Cross-strike section: Here a characteristic section through the thrust zone in eastern North Uist is described, about midway along the island chain. From west to east there are progressive changes in the style of faulting and the associated fault rocks (Fig. 4 and Table I). Because the host gneisses have a fairly uniform bulk composition, these changes must reflect changes in deformation environment.

In the region west of the thrust zone proper, pseudotachylyte friction-melt has been generated from place to place by transient seismic slip on extremely brittle faults cutting the gneisses (Sibson, 1975). These localised failure zones increase in number towards the thrust base, where disrupted pseudotachylyte veining also occurs in intensely microfractured crush zones, ranging up to 30m in thickness and largely composed of cataclasite and ultracataclasite. Above the thrust base, the gneisses are variably deformed in a 'crush melange' comprising crush breccias, microbreccias, protocataclasite and locally protomylonite. The crush mélange is cut by ductile mylonitic shear belts up to 50m in thickness, which lie in a braided network more or less concordant with the thrust zone envelope. All of the rocks within the mélange are to some extent hydrated and metamorphically downgraded, but within the shear belts almost total retrogression to lower greenschist assemblages (e.g. quartz + albite + muscovite + chlorite + epidote ± actinolite) nas occurred and the rocks are phyllonites with strongly developed L-S tectonite fabrics. Intense local hydration is evinced both by the retrogression and by the widespread development of hydrothermal quartz-chlorite veining in the phyllonites.

Direct field evidence for the contemporaneity of brittle and ductile thrust components is lacking, but arguments can be made against one set of features preceding the other (Sibson, 1976). A deformation sequence has been deduced whereby a D_0 phase of thrust inception (with which the scattered high-stress pseudotachylyte faulting and the initiation of the crush melange are associated) precedes the main phase of thrusting, D_1. Minor structures within the phyllonites indicate that the D_1 phase was followed by a D_2 phase of down-dip lag sliding. The whole complex is further disrupted by D_3 faults, presumed to be comparatively high-level features because of their association with incoherent gouge and breccia.

(2.2) The Alpine Fault, New Zealand

(2.2.1) Background: The Alpine Fault in the South Island of New Zealand (Fig.5) is apparently the main strand in a dextral system of continental transform faults linking opposed Benioff Zones along the boundary between the Indo-Australian and Pacific plates. A total of 480 km of post-Jurassic dextral strike-slip is indicated by a displaced ophiolite belt and other related features (Wellman, 1956), though there is some dispute as to the timing of the movements. One school (Fleming, 1970; Grindley, 1963, 1974; Reed, 1964; Suggate, 1963) argues for two separate phases of strike-slip, with about 330km of dextral strike-slip taking place in the late Jurassic - Early Cretaceous Rangitata Orogeny, the remaining 150km occurring in the late Cainozoic Kaikoura Orogeny which continues today. Others, relating local geology to sea-floor spreading data for the S.W. Pacific, suggest that the transcurrent movements began in Late Eocene times (c. 38 Ma) at the earliest, continuing through the Cainozoic to the present day (Carter and Norris, 1976; Molnar et al. 1975; Walcott, 1978). All are agreed that in Late Miocene times there was a change from a transcurrent regime to one involving a strong component of reverse slip. This has led to the formation of the Southern Alps in the Late Cainozoic by uplift of as much as 20km across the fault zone (Suggate, 1963). The swing in orientation of the Interplate slip-vector leading to this oblique compression across the plate boundary reflects a change in the position of the Indo-Australian/Pacific pole of rotation, as revealed by the sea-floor data (Rynn and Scholz, 1978; Walcott, 1978).

For much of its length the fault zone forms a pronounced lineament about 1km in width, separating markedly different terrains. To the north-west, a cover series of faulted and folded Upper Cretaceous and Tertiary sediments, with extensive Quaternary gravels, overlies a basement of mainly Lower Palaeozoic sediments, variably metamorphosed and invaded by granitoid intrusions which range up to Early Cretaceous in age. The upthrown Southern Alps to the south-east are largely composed of deformed metasediments from the Carboniferous ? - Jurassic New Zealand Geosyncline, and for some 400km along strike the hanging wall rocks are fairly uniform quartzo-feldspathic schists of garnet-oligoclase grade.

(2.2.2.) <u>Cross-strike section</u>: Rocks within the Alpine Fault Zone have previously been described by Reed (1964). On the basis of detailed petrographic studies, he recognised three broad textural groups: (a) incoherent fault pug, fault breccia and shattered rocks; (b) coherent cataclasite, mrtared and brecciated rocks; and (c) mylonite, augen mylonite, ultramylonite and blastomylonite. These three groups he assigned chronologically to periods of Alpine Fault displacement occurring respectively in the Quaternary and Late Tertiary phases of the Kaikoura Orogeny, and in the Late Jurassic to Early Cretaceous Rangitata Orogeny. The interpretation presented below differs fundamentally from that of Reed, because structural data indicate that all tectonite fabrics within the Alpine Fault Zone proper have been impressed during the Late Cainozoic phase of dextral-reverse-oblique slip (see 2.2.3, below).

Owing to the thick cover of rain-forest and the extensive Quaternary fanglomerates and glacial outwash gravels along the front of the Southern Alps, in situ rock exposure is generally far from continuous in stream sections cutting across the fault zone. However, by making many such traverses it has proved possible to piece together a composite 'hard-rock' section from the granitoid basement north-west of the fault to the Alpine Schists on the upthrown, south-eastern side (Fig.6). This procedure has been made mre simple by the near-constant lithology of the hanging wall (Fig.5). Note that in any particular traverse, the composite section may be truncated by the most recent fault break, with the south-eastern portion thrust over Quaternary gravels along a gouge zone. As a result, only the upper part of the section is exposed in many traverses. Other gouge zones locally disrupt the sequence and may bring about repetition of lithologies. Typically, the total width of the fault zone is about 1 to 1.5 km, but the relative proportions of the different fault rocks vary from one traverse to the next, particular textural types being completely absent in some sections. The proportions as shown in Fig. 6 are broadly characteristic of the central Alpine Fault region.

Passing into the fault zone from the downthrown side, we follow Reed (1964) in believing that the bulk of the quartzo-feldspathic cataclasites and augen mylonites are derived from the Tuhua Group granitoid assemblage. The origin of the green mylonites in the composite section remains a problem at this stage (Reed, 1964, categorises them as blastomylonites), but in the upper part of the sequence the fault rocks contain quartz-plagioclase-biotite-garnet assemblages and are clearly derived from the Alpine Schists. They include fine-grained, compact schist-mylonites interfingered to some extent with rather coarser 'curly' schist-mylonites. Progressing up through the schist-mylonites, recognisable enclaves of Alpine Schist appear, containing prolific 'fish-hook', minor fold hinges probably related to the Alpine F_2 structures of Rangitata age described by Grindley (1963) and Cooper (1974). In the higher strain regions, hinge-lines of these structures have been swung around from their general NE-SW trend, as seen in the alps, to be smeared out along the direction of finite elongation within the mylonites. The proportion of these Alpine Schist enclaves increases progressively through a hundred metres or so until the transition is complete.

Pseudotachylyte friction-melt was first described from a single locality in the Alpine Fault Zone by Wallace (1976) who referred to it as 'hyalomylonite', but the material has now been found in varying stages of devitrification at 17 localities spread over 270km along the fault trace (Fig 7). Where the lower, north-western portion of the fault zone is well exposed, the most intensive concentrations of pseudotachylyte can be seen to occur in association with cataclasite, but at one place or another it can be found cutting all other fault rock types apart from gouge. In mylonite series rocks, most of the pseudotachylyte has been generated by slip on south-easterly dipping fractures coincident with the mylonitic foliation, which has clearly acted as a preferential plane of brittle failure.

(2.2.3) Structural data in relation to the age of the fault rocks: Ductile shear zones develop by heterogeneous simple shear and may be expected to contain penetrative L-S shape fabrics. For high values of shear strain, mylonitic foliation should lie sub-parallel to the walls with a stretching lineation in the foliation indicating the transport direction across the shear zone (Ramsay and Graham, 1970). Pre-existing structures in the host rocks may be deformed as passive markers with linear features such as fold hinges, rotated and smeared out into near-parallelism with the transport direction (Escher 8 Watterson, 1974). Structural data from the Alpine Fault Zone suggest that all fault rock fabrics were impressed during the Late Cainozoic phase of oblique compression; mylonitic foliation indicates a dip of 40-50° SE for the fault zone at depth, while penetrative stretching lineations in the mylonites plunge in a direction sub-parallel to the present-day interplate slip-vector (Fig.8), and are consistent with dextral-reverse-oblique shear across the fault zone (Sibson et al. - in press).

Thus as the mylonitic fabrics are cut across by the pseudotachylytes and gouge zones, and to some extent are disrupted by the cataclasites, it appears that all of the fault rocks within the Alpine Fault Zone proper are Late Miocene or younger in age. This is in accord with isotope studies of the Alpine Schists adjacent to the fault (and of the schist-derived mylonites - C. J. Adams, pers. comm.), which suggest that major argon outgassing took place 4-5 Ma ago (Gabites 8 Adams - in press; Sheppard et al., 1975).

(3) INTERPRETATIVE MODELS

It is readily apparent from the cross-strike sections that there are major similarities and differences in the fundamental structure of the two fault zones. These structural aspects are now examined in relation to a general model for a fault zone developed in crystalline quartzo-feldspathic crust.

(3.1) General Fault Model

In both sections there is a change from truly cataclastic deformation to quasi-plastic mylonisation, accompanied by an increase in the metamorphic grade of the fault rocks, passing from the downthrown to the upthrown side of the fault zone. On the presumption that this lateral passage largely reflects the original distribution of fault rocks with depth, a simple essentially 2-layer fault model may be established by removing the effects of reverse shear (Sibson, 1977a). Largely as a consequence of the changing response of quartz to deformation with depth, a zone of frictional, mainly discontinuous deformation gives way beneath the greenschist transition to a zone where continuous shearing occurs by quasi-plastic processes in mylonite belts. The cut-out of seismic activity at a depth of IO-15km along the San

Andreas Fault (Eaton et al., 1970), corresponding to the expected transition depth to a greenschist environment for normal thermal gradients, accords well with this model. Provided the ratio of pore-fluid pressure to lithostatic load remains fixed, the shear resistance of fault zones may be expected to increase approximately linearly with depth through the frictional reglme, attaining a peak value just above the greenschist transition, beneath which the dominance of thermally activated deformation mechanisms causes a progressive weakening of the fault zone with increasing temperature and depth. Because of higher frictional constraints against movement, the peak shear resistance is likely to be greatest for thrusts and least for normal faults, with wrench faults having intermediate values (Sibson, 1974).

(3.1.1.) Mesh structure or single dominant shear zone?: The anastomosing mesh of both brittle and ductile shear belts which makes up the Outer Hebrides Thrust zone encloses large lozenge-shaped regions of comparatively undeformed material. In marked contrast, the Alpine Fault Zone consists of a single dominant lineament in which brittle and ductile deformation is closely juxtaposed. This is not to say, however, that deformation is completely homogeneous In the Alpine Fault Zone, because apart from the obvious discontinuities, textural variations indicate considerable small-scale strain heterogeneity in the mylonites.

Similar variations in the degree of fault zone localisation at high crustal levels are revealed by contrasting patterns of earthquake rupturing. For example, aftershock studies of the 1966 Parkfield earthquake on the San Andreas Fault show that rupture occurred on two linked segments spaced I km apart and extending to depths of 15 km (Eaton et al., 1970). On the other hand, flights of post-glacial river terraces progressively offset along several major, well-defined New Zealand faults (e.g. Lensen, 1968; Lensen & Vella, 1971) Indicate that in some regions at least, successive earthquake ruptures are restricted to essentially the same fault plane for periods of at least 20,000 years.

The application of "barrier theory" to rupture propagation (Das 8 Akl, 1967; Aki, 1978) has emphasised the need for an understanding of fault zone heterogeneity, why it occurs and what factors govern its scale. Does mesh structure arise solely from the geometrical need to accomodate large-scale asperities or fault zone curvature, as Flinn (1977) has suggested for the Walls Boundary Fault in Shetland? Are shear displacements restricted to one component of the mesh at all crustal levels for extended periods, the remainder becoming redundant, or is activity more or less evenly distributed between the mesh components? Do very large displacements tend to 'homogenIse' the structure of fault zones, especially in the more ductile regions?

Strain concentration along boundaries between regions of marked competence contrast in a well known characteristic of ductile deformation In high-grade basement terrains (e.g. Coward, 1976). This might account for the differing localisation of the two fault zones under consideration. Thus whereas the Alpine Fault Zone separates markedly different lithological terrains and is well localised, the Outer Hebrides Thrust has developed within a gneissose assemblage which, while being fairly uniform on a broad scale, contains many comparatively small-scale Inhomogeneities such as metabasite bodies, around which the mesh structure may develop.

(3.1.2) Fault zone width versus depth: In the general fault zone model (Fig.9), the width of the fault zone is shown as increasing with depth. An important consequence is that shear strain-rates in the ductile regime would decrease with depth. The inference of increasing width with depth is based on only a few ancient fault systems exposed at different erosion levels along their strike. They include the Outer Hebrides Thrust Zone (Sibson, 1976), the Nordre Stromfjord shear belt in West Greenland (Bak et al., 1975), and the Najd fault system in Saudi Arabia (Moore - in press). Much further work is needed to substantiate this concept.

(3.1.3) Cataclastic crush zones and distributed cataclasis: Within the Outer Hebrides Thrust Zone. well defined crush zones containing cataclasite - ultracataclasite and some pseudotachylyte range up to several tens of metres in thickness. Field evidence suggests that significant shear displacements, in places of the order of kilometres (Francis 8 Sibson, 1973), have occurred across these structures. Whether such cataclastic crush zones have developed largely by seismic or aseismic shearing processes remains uncertain though the former seems more likely in view of the pervasive brittle mlcrofracturing, the quartzo-feldspathic nature of the host rocks which favours stick-slip (Jackson 8 Dunn, 1974), and the presence of some pseudotachylyte friction-melt. However, distributed cataclasis also occurs throughout the crush mélange, though the finite shear strain developed across this material is probably slight.

A similar situation occurs in Shetland along the Walls Boundary Fault, the possible northern continuation of the transcurrent Great Glen Fault (Flinn, 1961). The actual fault is traceable for over 40 km as a well defined planar crush zone, often less than 0.5 m in width and containing intensely comminuted cataclasite - ultracataclasite. across which contrasting lithologies are sharply juxtaposed (Flinn, 1977). Flanking country rocks, including sandstones, metamorphics and a granitoid complex are variably cataclased to crush breccias, microbreccias and protocataclasite, and have undergone minor subsidiary faulting for distances of a kilometre or more away from the fault. Mineralogy suggests that much of the cataclasis took place under zeolite or lower conditions of metamorphism. Flinn considers that almost all the strike-slip displacement of more than 30 km has occurred by slip along the main fault; very little can be associated wlth the broad belt of cataclasis.

One possible interpretation, assuming the fault to have been seismically active, is that while comminution in the cataclastic crush zones occurred during actual slip episodes (the γ and possibly the δ phases of the earthquake cycle), the distributed cataclasis is a consequence of progressive microcracking caused by long-continued stress cycling of the region adjacent to the fault (the α and, especially. the 8 phases of the earthquake cycle).

(3.1.4) The gouge-cataclasite transition: In both fault zones, the sections are disrupted by late dislocations containing incohesive gouge and breccia which are presumed to have developed under comparatively near-surface conditions. The distinction between gouge and cataclasite is based rather loosely on the absence or presence of 'primary cohesion' (Higgins, 1971). It is not clear to me that primary cohesion can always be distinguished from secondary cementation, especially in very fine grained rocks. I have assumed that cohesion develops by grain impingement and cementation in quartz-bearing rocks once they are at a sufficient temperature for quartz to be re-distributed by hydrothermal solutions. Available evidence suggests that this may generally occur at temperatures of around 100°C (Kerrick et al., 1977).

It seems unlikely, therefore, that truly incohesive gouge can exist to depths of around 10 km as Wu et al (1975) have suggested for the San Andreas Fault. One suspects that hydrothermally altered, but essentially cohesive cataclasite could account for the gravity low coincident with the fault zone equally as well as the clay-rich gouge postulated by Wang et al. (1978).

(3.1.5) Evidence for high fluid pressures: One major difference between the two fault zones is the evidence for intense hydration and high fluid pressures in the phyllonite belts of the Outer Hebrides Thrust as compared to the mylonites of the Alpine Fault Zone. The phyllonites have formed by the mylonisation and syntectonic retrogression of a high-grade gneiss assemblage, the latter process involving exothermic hydration reactions. Tension gash shear zones, comprising en-echelon arrays of quartz-chlorite veins, occur from place to place within the phyllonites, while disrupted and deformed veins can be found sheared-out into the mylonitic foliation. If, as seems likely, the veins originally developed by hydraulic fracturing under shear (Secor, 1969), one may infer that fluid pressures within the phyilonite belts locally exceeded the least principal compressive stress so that mylonisation was proceeding under zero effective confining pressure.

In contrast, hydrothermal veining within the Alpine Fault Zone is scarce and retrogression in the mylonites, especially those derived from the biotite-garnet assemblages of the Alpine Schists, is slight. This may be attributed at least in part to a general lack of circulating hydrothermal fluids.

(3.1.6) Distribution and development of pseudotachylyte: Again, there are both similarities and differences in the distribution of pseudotachylyte friction-melt, the product of seismic slip on discrete planes (Sibson, 1975), in the two fault zones. The most intensive concentrations occur in association with cataclasite-ultracataclasite at the base of both zones. In the Outer Hebrides distributed pseudotachylyte faulting, perhaps associated with the inception of the thrust zone, occurs structurally below the base of the thrust zone but does not affect the phyllonitic mylonites above the thrust base.

However, above the base of the Alpine Fault Zone, all mylonlte series rocks have been affected by pseudotachylyte faulting, though not usually to any great extent. In terms of the progressive evolution of the fault zone (see below) it appears that as a result of reverse shear, originally ductile mylonites were raised to levels where seismic slip along discrete planar fractures became the dominant failure process (cf. Grocott, 1977).

A possible explanation for this contrasting behaviour lies in the disparity between the fluid pressure regimes postulated for the two sets of mylonites. In the Outer Hebrides phyllonites, the presence of an aqueous intergranular fluid would have inhibited friction-melting during seismic slip (Sibson, 1977b).

(3.2) Mineral Deformation Mechanisms

In building up models of fault zones we are now at a stage where by means of optical and electron microscopy, textures and microstructures can tentatively be linked to crystal deformation mechanisms. It is apparent that the mechanical response of shallow to intermediate quartzo-feldspathic crust is largely governed by the varying behaviour of quartz, though the presence of other minerals may be critical in allowing certain mechanisms to operate (Elliott, 1973; White, 1976). For example, the presence of phyllosilicates may greatly enhance deformation involving grain boundary diffusion (Rutter and Mainprice, 1978).

We first consider the distribution with depth in the crust of fault rocks produced by steady-state aseismic shearing, in relation to probable metamorphic environments and the dominant deformation modes for quartz (Fig. 10). The change of metamorphic facies with depth for a constant geothermal gradient of 30°C/km is modified from Turner (1968) in the light of more up to date information on metamorphic temperatures (Kerrick et al., 1977; Miyashiro, 1973; Winkler, 1974). Note that these boundaries are all based on the assumption that $P_{load} = P_{H_2O}$. Probable fields in which particular quartz deformation mechanisms dominate are shown schematically on a plot of mean grain size versus temperature. Such plots are based on the assumption of a constant shear strain-rate throughout the zone. Typical strain-rates might be in the range 10^{-10} to $10^{-12}s^{-1}$, corresponding to displacement at 1 cm/yr across shear belts respectively 3 and 300 m in width (Sibson, 1977a). With time it should be possible to determine the boundary parameters for a particular strain-rate fairly accurately; at present our controls are few and the boundaries shown are little more than guesses. Elliot (1976) has suggested that pressure solution sliding (involving water-assisted grain-boundary diffusion) on faults becomes important at temperatures in excess of 250°C. Kerrich et al. (1977) have determined the transition between flow mechanisms involving grain boundary diffusion and dislocation creep as a function of temperature and grain size, but for quartzo-feldspathic rocks that have probably been deformed at much slower strain-rates (favouring diffusion processes) than those expected in fault zones. Several workers (Elliott, 1973; McClay, 1977; Rutter, 1976; White, 1976) have used other forms of deformation map for quartz to show that decreasing grain size at constant temperature and strain-rate can induce a change from flow by crystal plastic processes (mainly dislocation creep) to flow involving grain boundary diffusion, possibly water-assisted (i.e. pressure solution). White (1976) has suggested that in quartz-rich mylonites this transition, accompanied by marked strain-softening, occurs as grain size is reduced from 100 to 10 μm.

During seismic faulting, cataclasis is likely to be the main deformation mechanism at all crustal levels because of the high slip rates and power dissipation. Under dry conditions friction-melting with the production of pseudotachylyte may occur at depths in excess of perhaps a kilometre, provided slip is localised within a thin zone (Sibson, 1977b). However, textures developed in these transient events are susceptible to obliteration by the longer term processes of deformation in fault zones, especially in deep mylonite belts developing by aseismic sheaf.

(3.3) Evolution of Reverse-Slip Fault Zones

Reverse shearing across a general fault model such as that in Fig. 9 will lead to the juxtaposition at the surface of fault rocks generated over a range of depths. However the details of the process can be quite complicated. As an illustration we consider the evolution of a fault zone consisting of a single dominant shear belt at all crustal levels (Fig. 11). For simplicity the active gouge - cataclasite and cataclasite-mylonite transitions are considered to be sharply defined interfaces remaining at a fixed depth with respect to the ground surface. Clearly the final distribution and range of fault rocks exposed at the surface depends both on the extent of reverse shear and the erosion level at which the fault zone is exposed. Also, original fault rock textures formed at depth may be disturbed during their passage to the surface. Thus mylonites are likely to suffer some cataclastic deformation during upwards transport and as a result of pressure relief, originally cohesive fault rocks may tend to disaggregate and undergo near-surface hydrothermal alteration.

In reality fault zone evolution will be complicated by a number of factors. For a start, more elaborate models would have to consider the effects of reverse shear across a multi-strand fault zone. Then there will be severe problems of strain compatability in the vicinity of the change-over from discontinuous to continuous deformation, giving rise to a complicated transition zone. It should also be borne in mind that larger earthquake fault ruptures may be expected to propagate down through mylonitic shear belts causing occasional transient loss of continuity; though the textural evidence for this will rarely be preserved (Sibson, 1977a). The transition levels will to some extent be governed by the effects of differential uplift and erosion on isotherms.

A further possibility is that shear heating within the fault zone may cause upwards migration of the cataclasite-mylonite transition with time. Such may well have been the case with the Alpine Fault where patterns of argon outgassing in the Alpine Schist are strongly suggestive of significant shear heating during the Late Cainozoic phase of reverse slip (Gabites & Adams, in press; Scholz et ai., in press; Sibson et al., in press). Shear heating could also help to account for the lack of retrogression in the Alpine Fault mylonites. Against this it has to be pointed out that micro-earthquake activity in the vicinity of the Alpine Fault, while failing to define the fault zone, still extends to depths of around 15 km as on the San Andreas Fault (Scholz et al., 1973).

(4) CONCLUSIONS

Despite difficulties arising mainly from textural overprinting, fault rocks and structures can be used to build up conceptual models of crustal fault zones. Though in need of much refinement, these models show great potential for providing a better understanding of fault zone processes relevant to the shallow earthquake mechanism, especially when combined with other sources of information.

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FIGURE CAPTIONS

Fig. I - Methodology for evolving conceptual models of fault zones.

Fig. 2 - Stress/Time relationships in and around fault zones.

Fig. 3 - Textural classification of fault rocks (after Sibson, 1977a)

Fig. 4 - Map and schematic cross section of the Outer Hebrides Thrust zone in eastern North Uist (after Sibson, 1977a).

Fig. 5 - The Aipine Fault Zone: sampling localities (after Sibson et al. - in press).

Fig. 6 - Schematic composite section through the Alpine Fault Zone (after Sibson et al. - in press).

Fig. 7 - Established pseudotachylyte localities along the Alpine Fault Zone (after Sibson et al. - in press).

Fig. 8 - Lineation plunge directions and earthquake slip-vectors (after Walcott, 1978) in the Alpine Fault System. (Note - all lineations are penetrative stretching lineations in ductile mylonites, apart from that associated with the White Creek Fault which is a fault surface striation)(after Sibson et al. - in press). Data quality: A A C - well clustered measurements; B & D - poorly clustered measurements; E - I measurement only.

Fig. 9 - General model for a fault zone in quartzo-feldspathic crust.

Fig. 10 - Fault rocks, metamorphic environment and dominant quartz deformation mechanism for steady aseismic shear across a fault zone in quartzo-feldspathic crust.

Fig. II - Evolutionary model for a I-strand reverse fault zone.

Table i - Fault rocks and style of faulting in the Outer Hebrides Thrust Zone (after Sibson, 1977a).

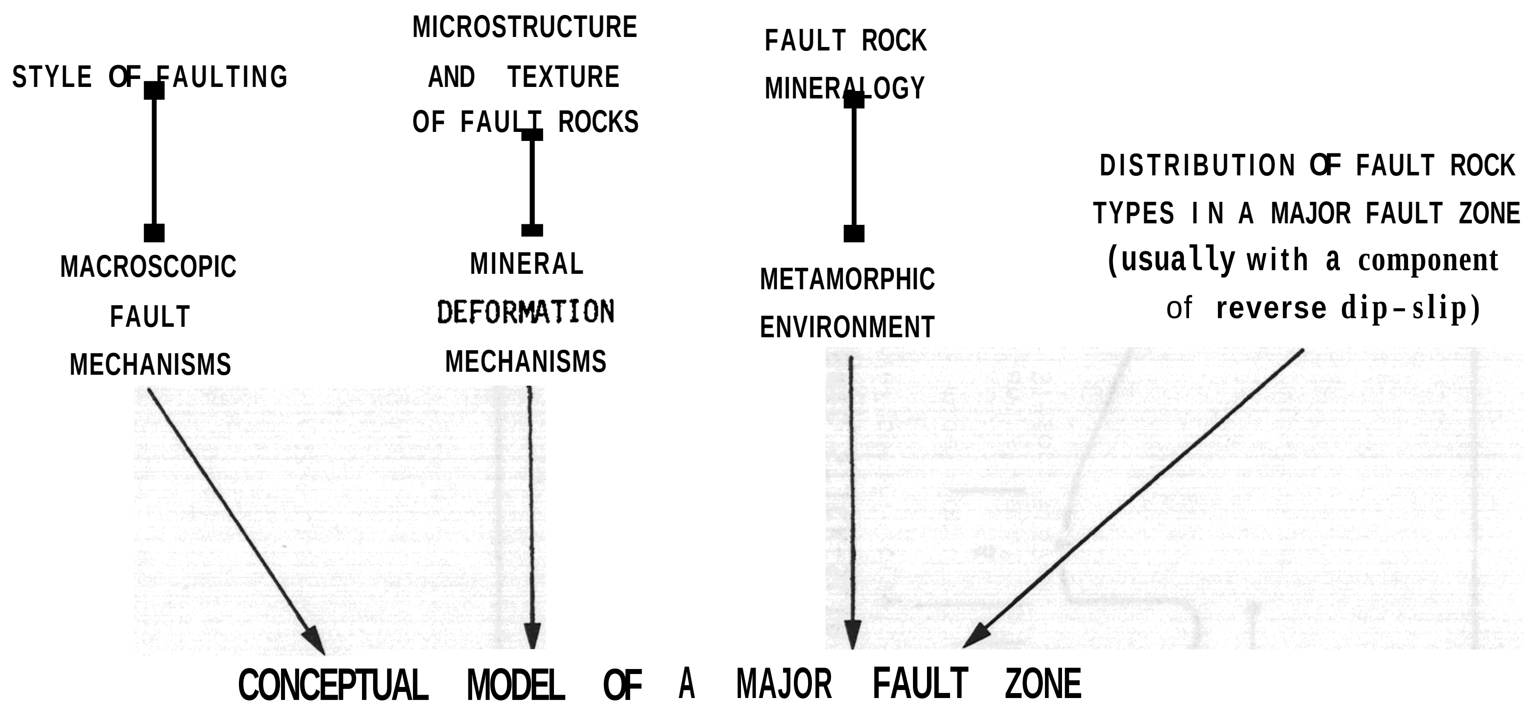

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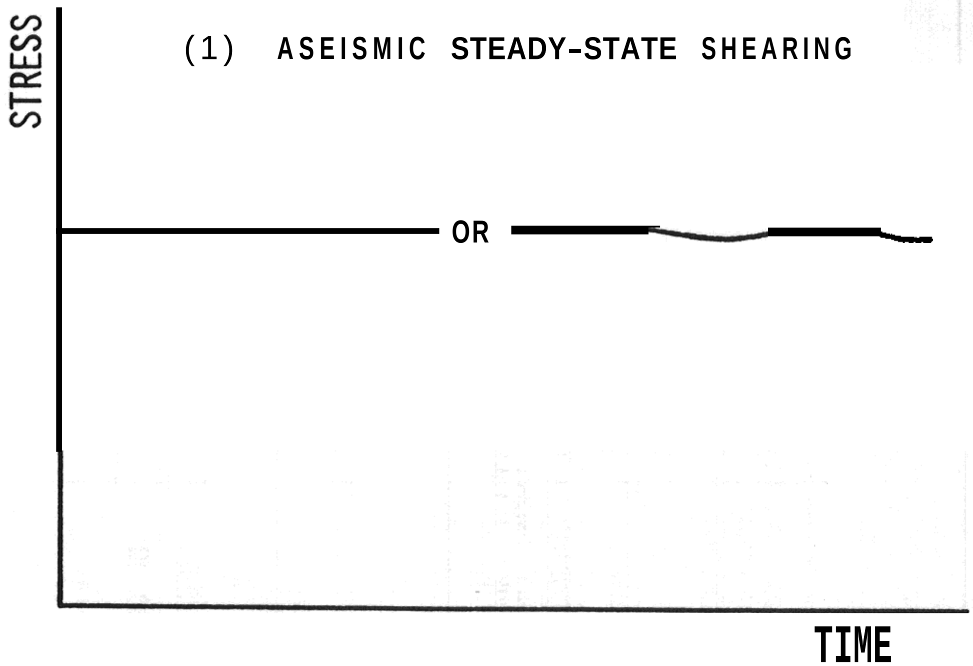
Fig.2 - **Stress/Time** relationships in and around fault zones.



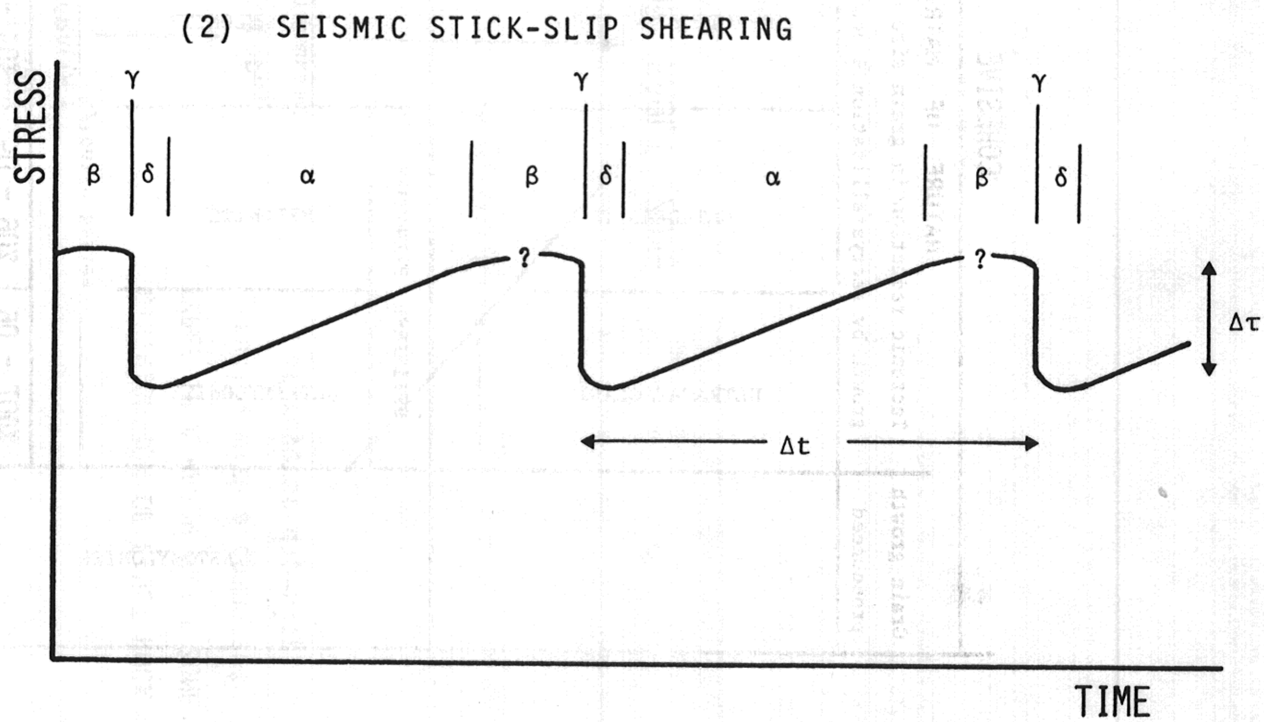


Fig.3 - Textural classification of fault rocks (after Sibson, 1977a).

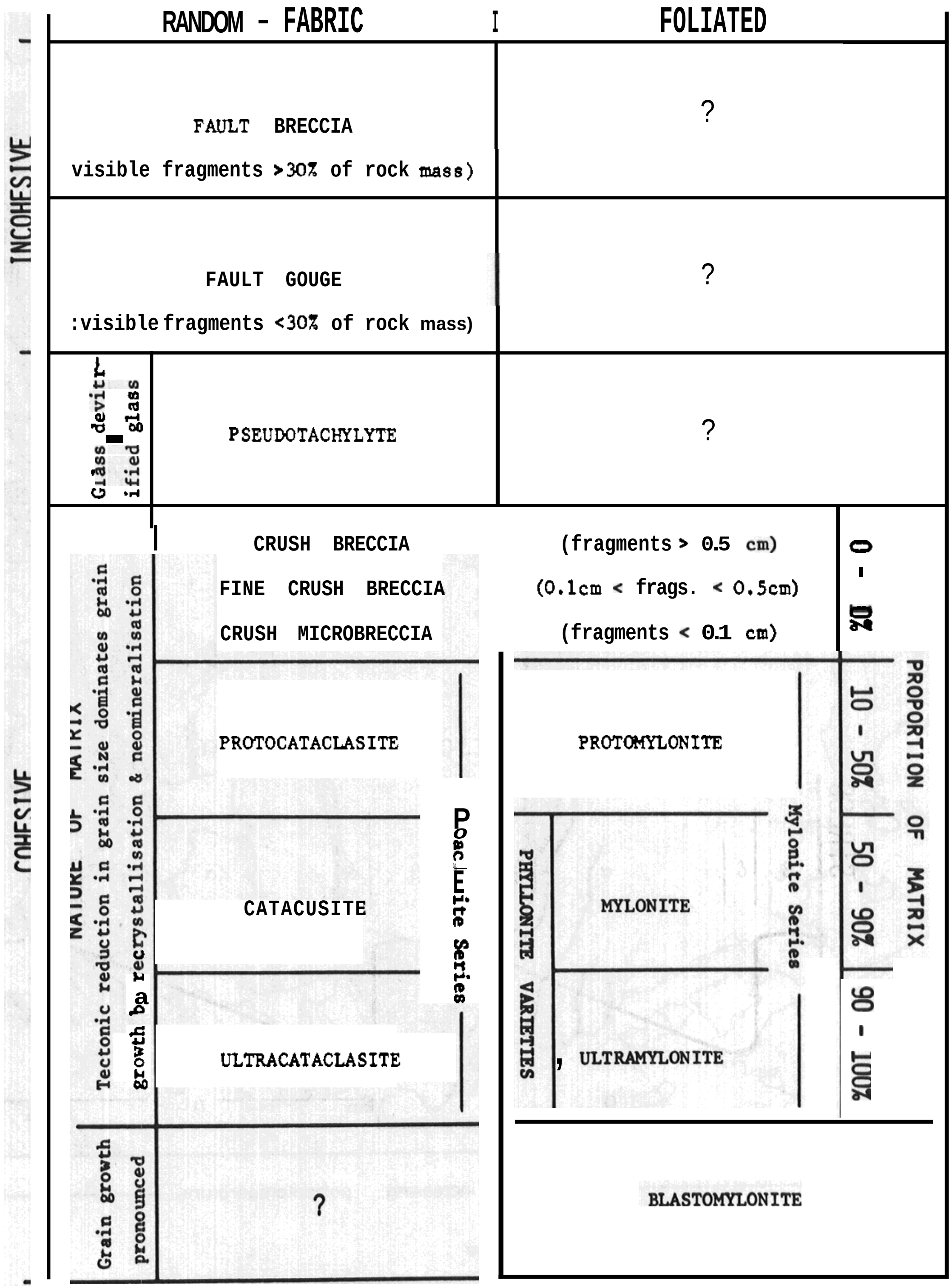


Fig.4 - Map and schematic cross-section of the Outer Hebrides Thrust zone in eastern North Uist (after Sibson, 1977a).

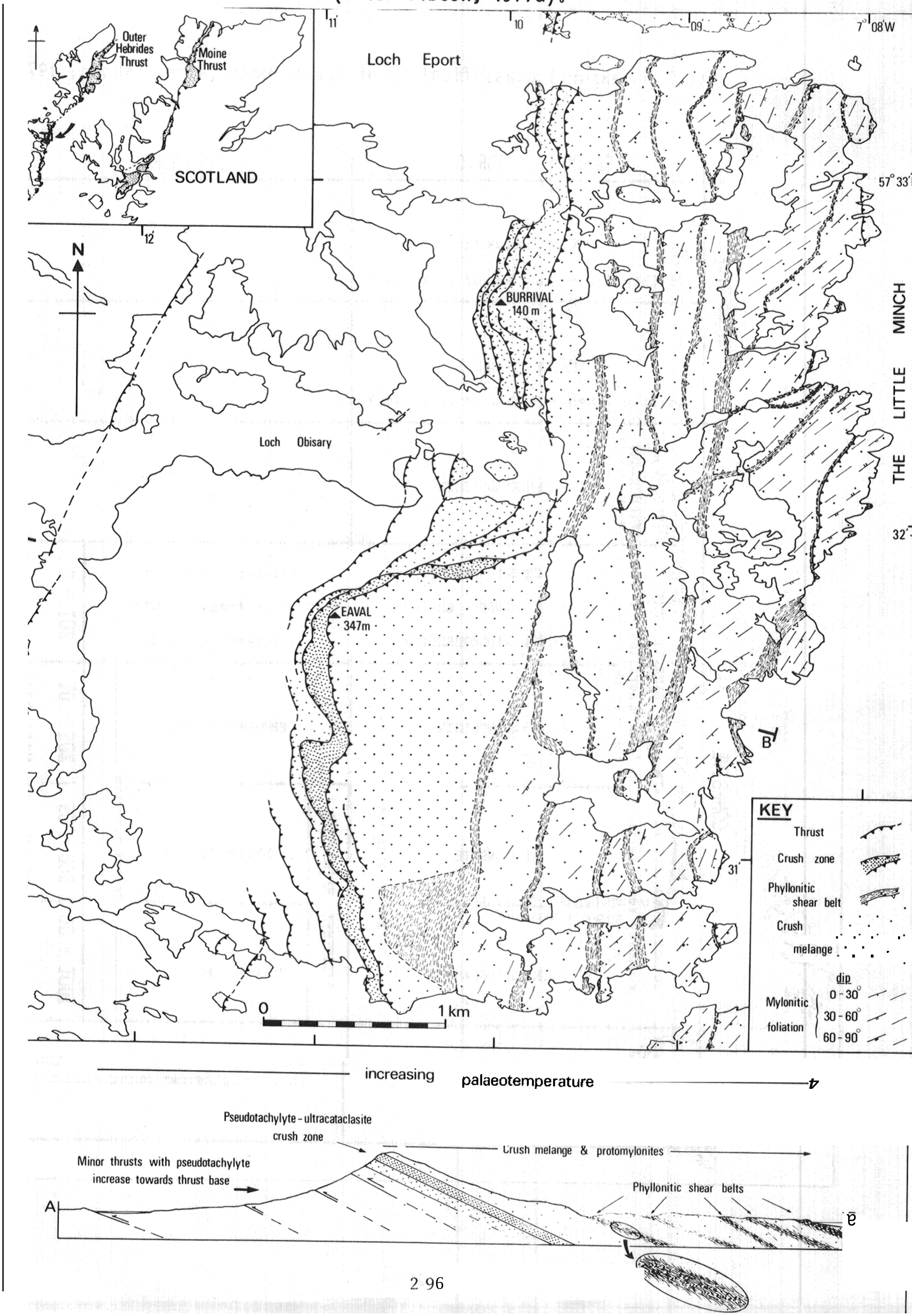
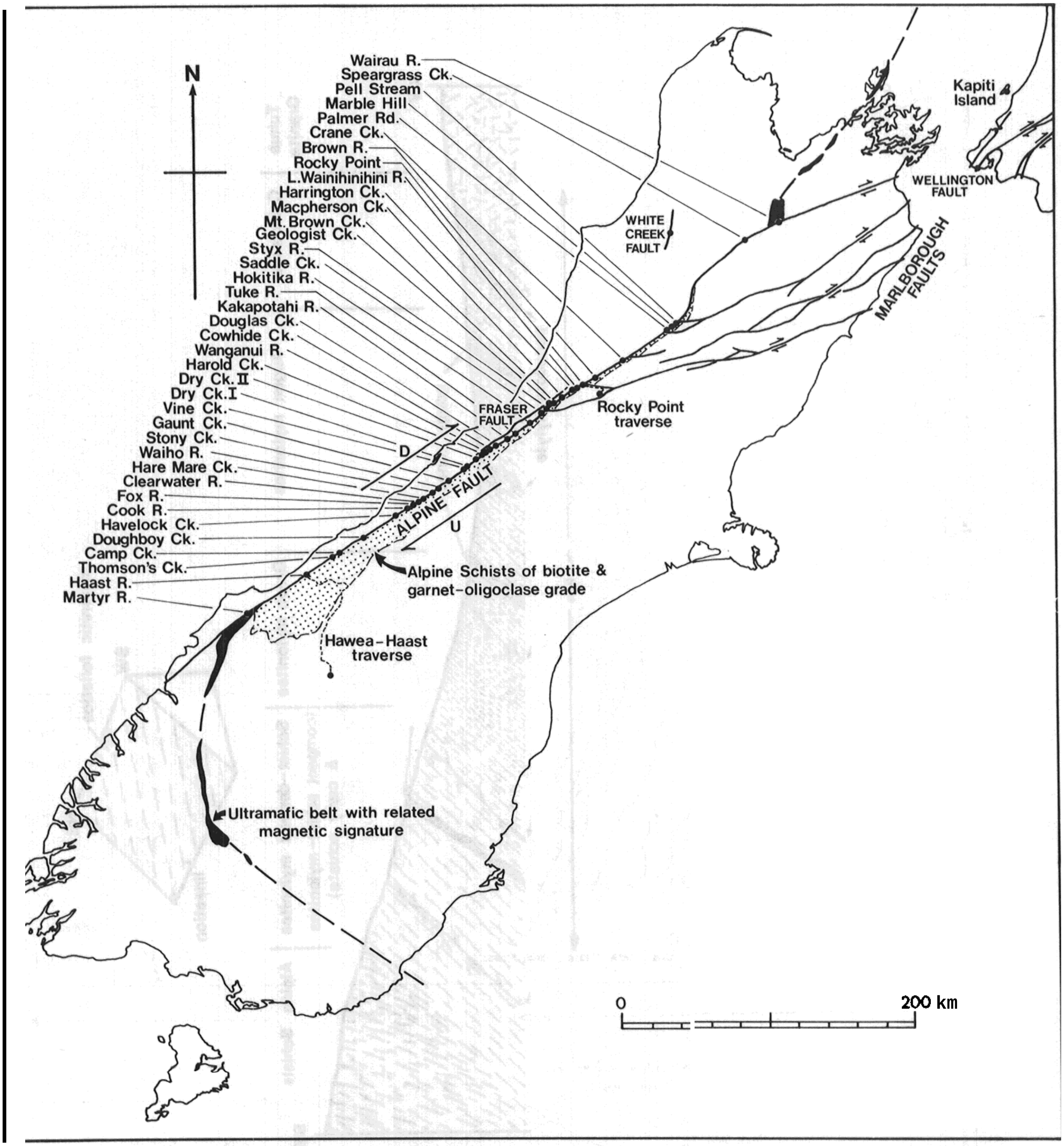
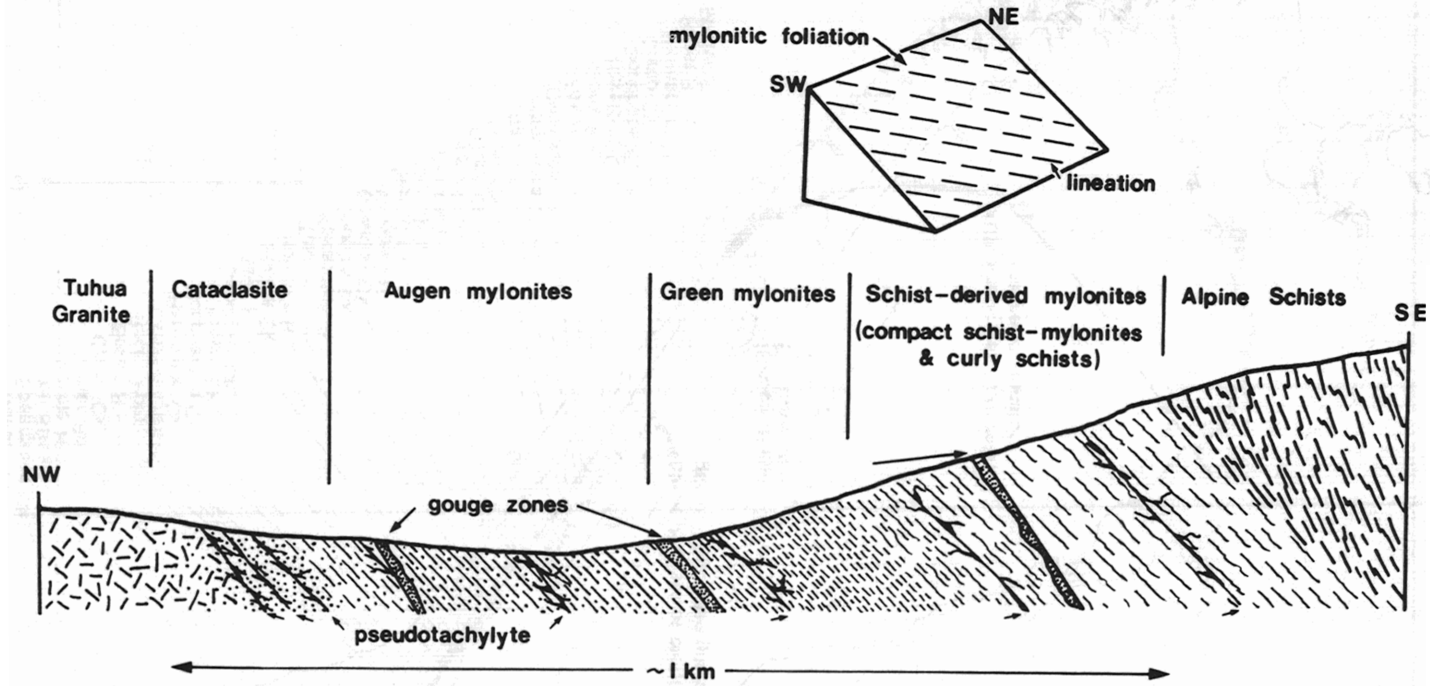


Fig.5 - The Alpine Fault Zone: sampling localities (atter Sibson et al. - in press).





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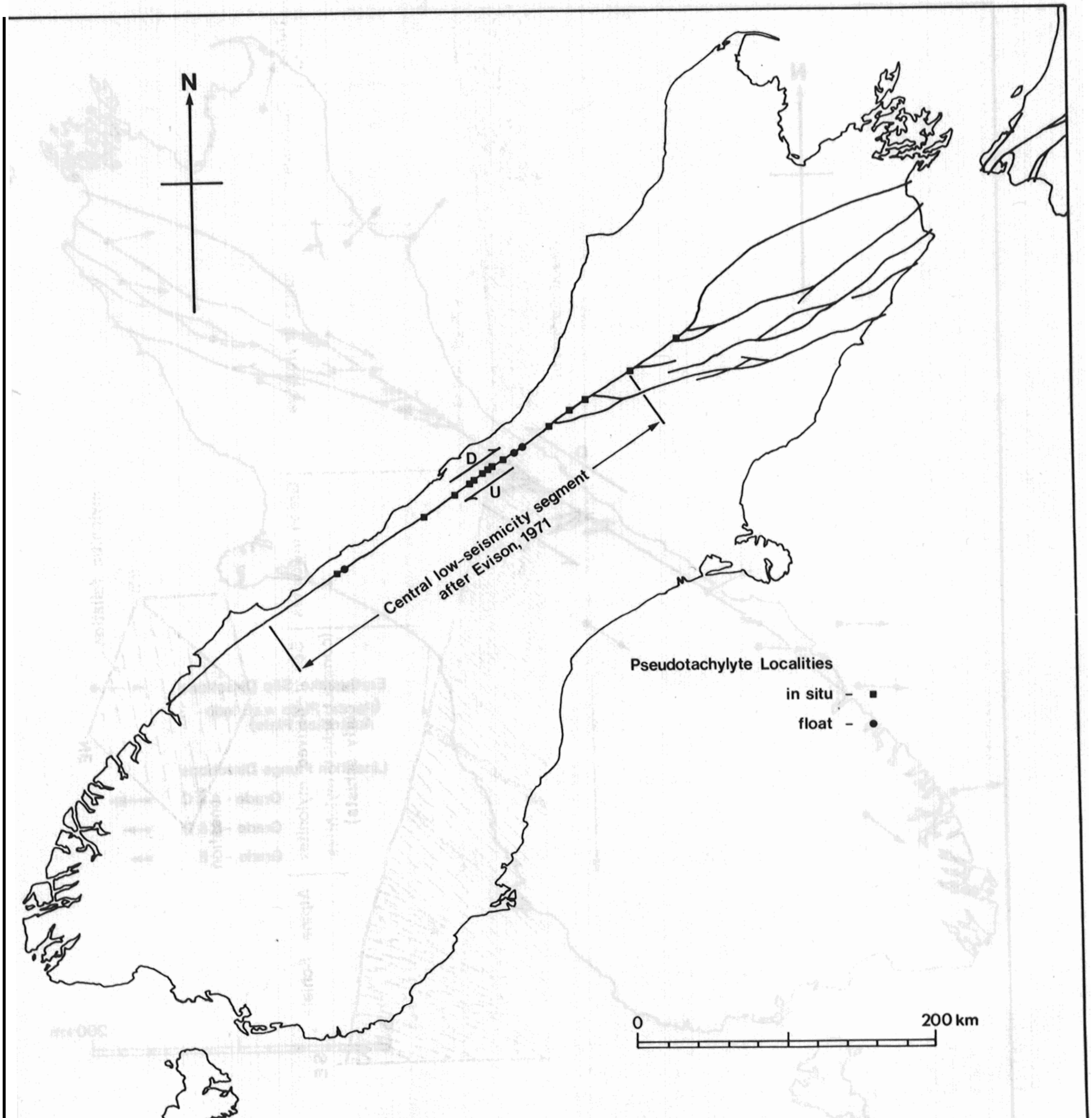

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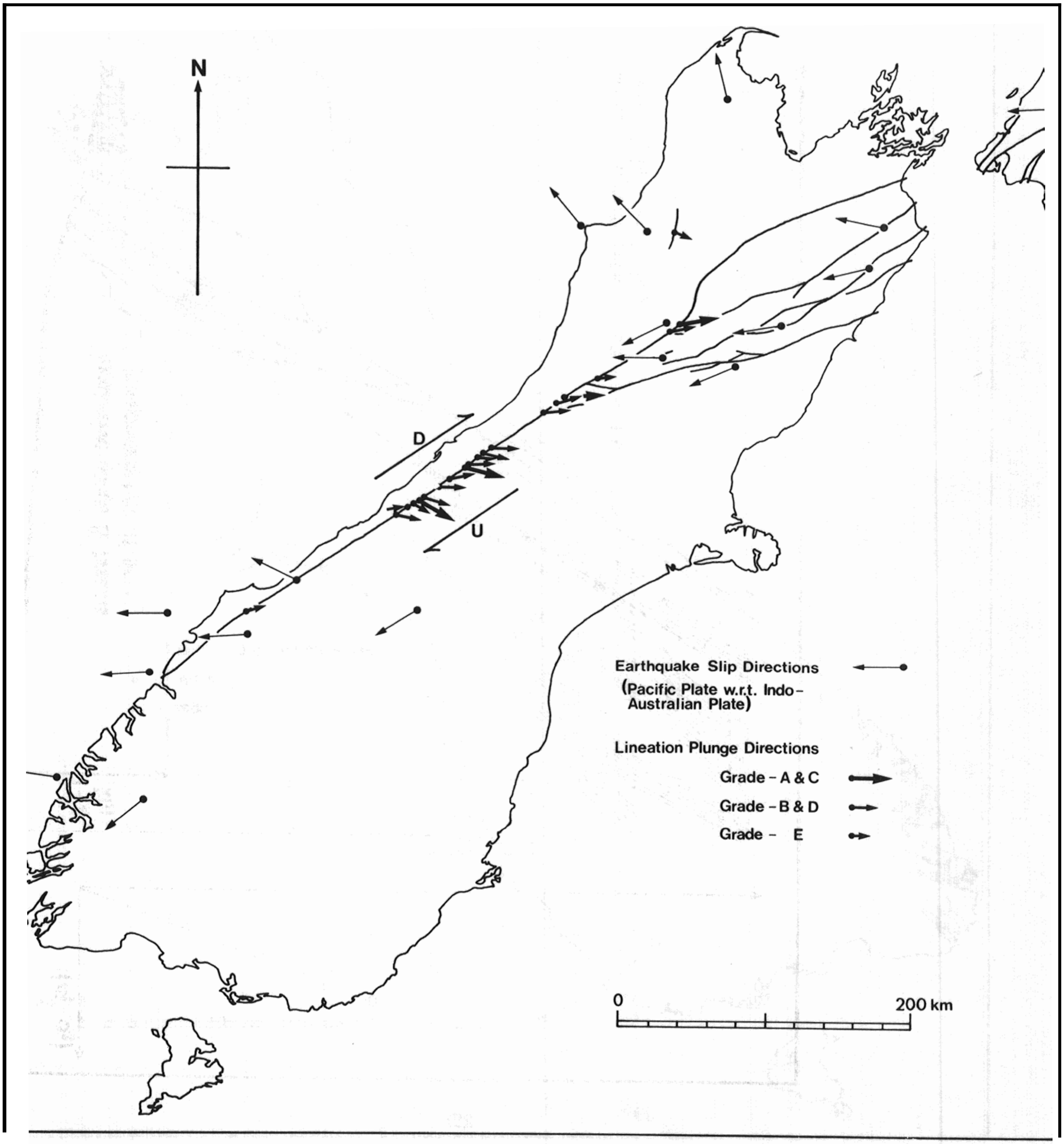

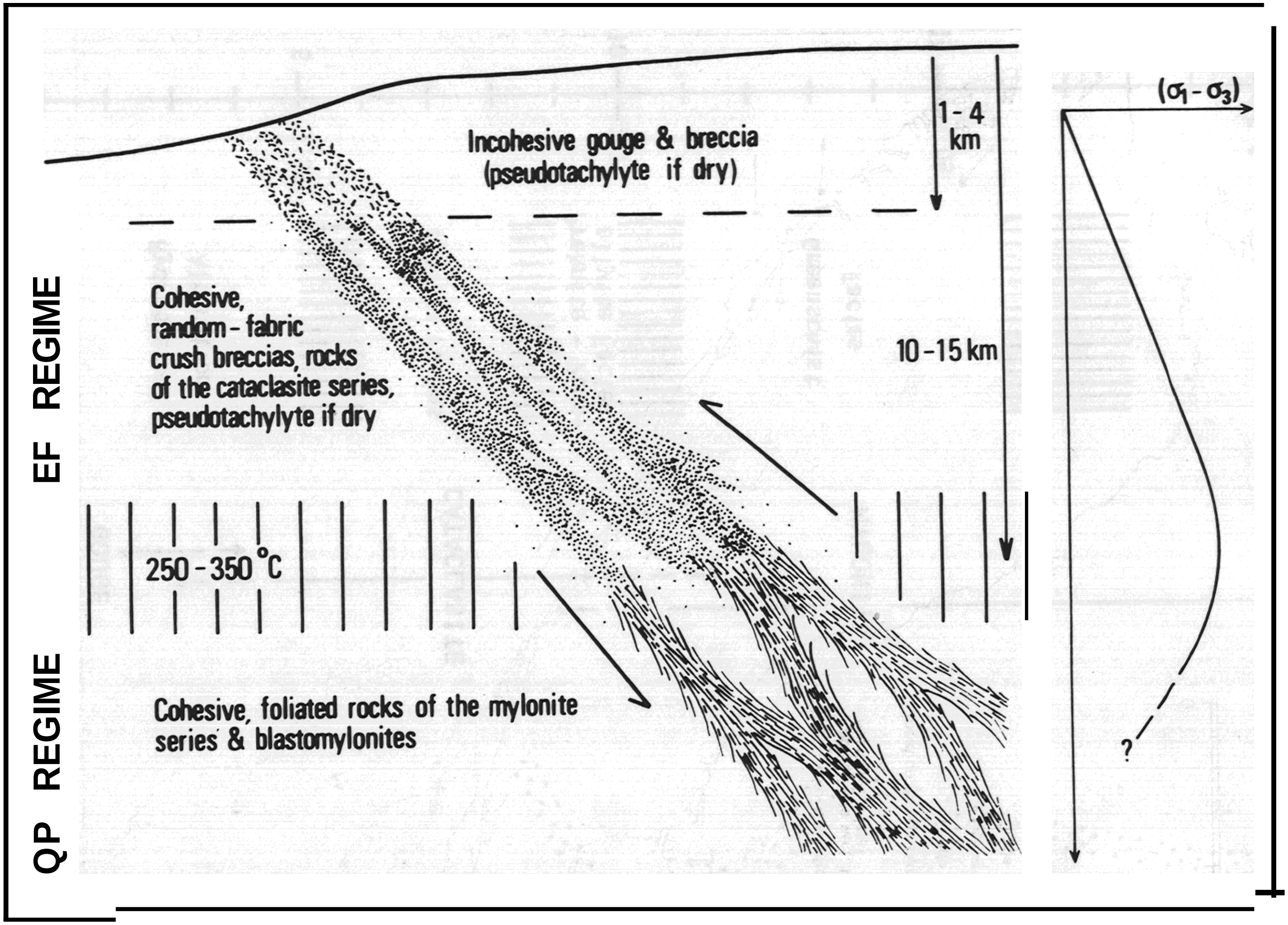

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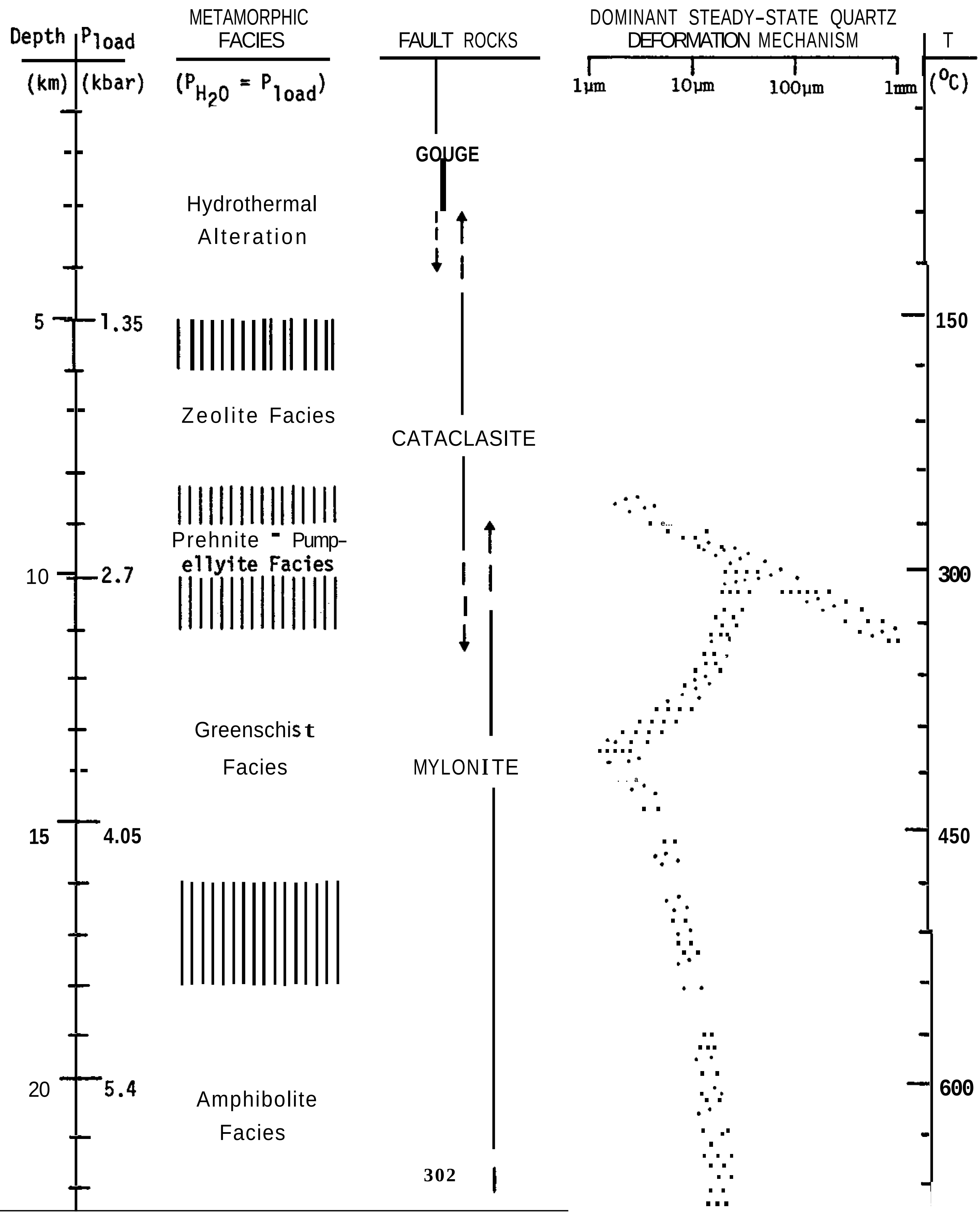

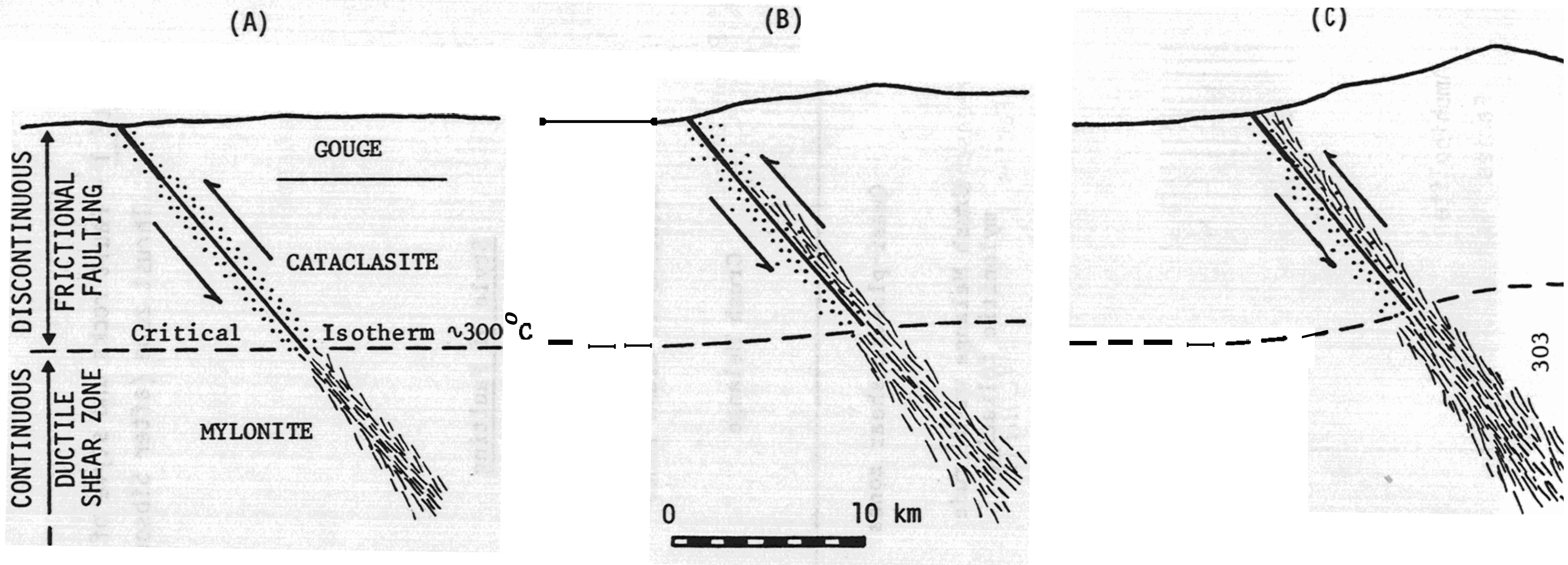

Fig.11 - Evolutionary model for a 1-strand reverse fault zone.

303

TABLE 1 - Fault rocks and style of faulting in the Outer Hebrides Thrust zone (after Sibson, 1977a).

	Style of Faulting	Fault Rocks
ELASTICO-FRICTIONAL	Brittle shearing of intact rock Sliding on existing planes	Pseudotachylyte
	Cataclastic crush zones	Cataclasite-Ultracataclasite (& some pseudotachylyte)
	Crush Melange	Crush breccias, microbreccias & protocataclasite
QUASI-PLASTIC	Quasi-plastic shear zones	Phyllonitic mylonites & ultramylonites
	Crush Melange with crude mylonitic foliation	Protomylonites