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Soft Plate and Impact Tectonics

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lution may be irregular or even chaotic; spherical symmetry-breaking bifurcation will affect thermal convection in the Earth's mantle (Chossat and Stewart 1992).

3.12 Time-Dependent Mechanical Response of the Oceanic Lithosphere; Strain Partitioning and Deformation Mechanisms

Much of our present understanding of the major processes that control the complex evolution of the oceanic lithosphere is mainly the result of scattered direct observations, coupled with inferences based on thermal and mechanical models that consider both geochemical (including petrogenetic) and geophysical data, as well as estimations of physical properties deduced from experiments on monomineralic and/or non-hydrothermally altered rocks (e.g. Agar 1994; Fisher 1998, and references therein). That understanding is therefore chiefly conceptual rather than specific, and many key questions will remain unsolved until additional direct measurements are completed in a variety of ocean crustal ages and settings. This is particularly clear when one examines the available rheological models for the oceanic lithosphere, where major gaps in our knowledge severely constrain the boundary conditions to be imposed on the numerical modelling (e.g. Watts 1982; Bergman and Solomon 1984; Buck 1986; Chen and Morgan 1990; Lin et al. 1990; Hayes and Kane 1991; Tolstoy et al. 1993; Shaw and Lin 1996).

Uncertainties and the lack of knowledge related to the geological processes that strongly influence strain accommodation in space and time in the oceanic lithosphere (including strain partitioning and local variations in stress states) are critical in the understanding of its rheology. Therefore, the evaluation of the integrated effects of magmatism, fluid flow and crustal extension at large scales does not enable an adequate characterisation of the time-dependent mechanical response of the oceanic lithosphere, thus preventing the construction of high-resolution rheological models (especially for the upper oceanic crust). This means that further work is necessary in order to understand the rates and mechanisms by which thermal, fluid flow and chemical variations of the oceanic lithosphere are actually controlled over a continuum of scales. In other words, comprehensive examination of the magmatic, hydrothermal and crustal extension evolution should be considered in any realistic rheological model for the oceanic lithosphere. Microstructural studies, both in situ and experimental, are thus a complementary, but essential, tool in these investigations (e.g. see Agar 1994, for a detailed discussion).

In addition, detailed examination of oriented samples (to provide kinematic interpretations and palaeostress orientations) and more experimental data are needed, particularly on strain accommodation by texturally diverse mineral assemblages similar to those usually found in the heterogeneous suite of altered rocks recovered by dredging of the ocean crust and in their analogues in ophiolite complexes.

Some general comments about strain partitioning and the most probable active deformation mechanisms during oceanic crust evolution can, however, be made, considering the magmatic-hydrothermal crustal extension evolution that is believed to characterise both fast- and slow-spreading systems (for a comprehensive review, e.g. see Agar 1994), thus tentatively responding to the major challenge of the present book.

The key issue is that the oceanic lithosphere deforms by buckling as a consequence of its viscoplastic behaviour acquired after 10–20 Ma, regardless of the magmatic-hydrothermal crustal extension evolution experienced by the crust. As already discussed, buckling of the oceanic crust is preceded by the accommodation of 4–10% of homogeneous shortening, perhaps ascribable to different (brittle) failure mechanisms which converge to similar consequences, i.e. to prevailing aseismic creep. If so, strain partitioning in the oceanic crust as a whole, as well as its stress state variations in space and time, reflects the heterogeneous results of the intricate relationships established between magmatic and hydrothermal processes during crustal extension, which are expected to be notably distinct in fast- and slow-spreading systems.

The architecture of the oceanic crust is, indeed, far more puzzling than commonly assumed in general simple layered models. It reveals strong lateral and vertical lithological and structural heterogeneities, mostly due to multiple magmatic injections subsequently overprinted by polyphase hydrothermalism (e.g. Cann 1974; MacDonald et al. 1988; Karson and Rona 1990; Jacobson 1992; Smith and Cann 1992; Harding et al. 1993; Detrick et al. 1994; Head et al. 1996). These heterogeneities, also well documented in ophiolites (e.g. Moores and Vine 1971; Coleman 1977; Christensen 1978; Christensen and Smewing 1981; Varga and Moores 1985), may therefore be responsible for a significant and variable lithospheric weakening, since they reflect variations of many physico-chemical parameters (such as mineralogy, grain size, permeability, cohesion, etc.) that dramatically alter the time-dependent mechanical response of rocks, and/or changes in local stress configurations (mainly dependent on pore fluid pressure buildups and on the geometry of discontinuities mostly formed at lithological contacts), thus controlling strain accommodation.

According to the available data, narrow, but long-lived magma conduits lead to multiple intrusive events at fast-spreading systems (e.g. Hayes and Kane 1991; Chen 1992; Wilcock et al. 1992; Christeson et al. 1994; Su et al. 1994). A higher background geothermal gradient and a relatively swift translation of the new oceanic lithosphere further from the ridge axis are therefore predicted. In addition, possible deep melt accumulations over a relatively wide region (from 5 to 20 km; Garmany 1989; Barth et al. 1991) will favour aseismic slip during fast spreading, causing strain concentration at the base of the young oceanic crust. In these circumstances, magmatic over-pressuring related to variations in magma supply will tend to restrain normal fault development, thus explaining the low seismicity and relief of these oceanic settings (e.g. Bratt et al. 1985; Parsons and Thompson 1991). However, even in the presence of important tectonic faults, similar to those often found in slow-spreading systems, seismicity also remains relatively low and is usually concentrated at the ridge segment ends, mainly at the inside corners of ridge-offset intersections (e.g. Forsyth 1992; Shaw 1992). This means that the cause of the aseismic behaviour of the fast-spreading ridges must be sought elsewhere.

Episodic magmatic activity is believed to characterise slow-spreading systems, thus enhancing the relative importance of crustal extension by faulting (e.g. Harper 1985; Karson and Rona 1990; Cannat et al. 1991; Hayes and Kane 1991; Chen 1992; Smith and Cann 1993; Su et al. 1994). Changes in fault geometry, fault spacing and displacement (often increasing towards ridge-offset discontinuities) are common and presumably ascribable to variations in lithosphere thickness and fault dips (e.g. Chen and Morgan 1990; Neumann and Forsyth 1993; Shaw and Lin 1993). These intricate structural configurations, strongly influence along-axis valley morphology and often lead to thinner and discontinuous ridge segment ends, where strongly altered gabbros and peridotites outcrop (e.g. Cannat 1993; Cannat et al. 1995; Escartín et al. 1997a,b). Once in a position to be hydrated, peridotites will be serpentinised if the average temperature remains below ca. 400° C, and this concentrates fault slip on the inside corner of ridge-offset intersections, assuming that hydration is first restricted to fault planes (Raleigh and Paterson 1965; Francis 1981; Janecky and Seyfried 1986; Rona et al. 1987, 1992; Escartín et al. 1997a; Moore et al. 1997). However, given the distribution of serpentinites at slow-spreading systems, not only at ridge segment ends, but also at many localities along the axial valleys (e.g. Bonatti and Michael 1989; Tucholke and Lin 1994; Cannat et al. 1995), the chemical softening processes related to non-serpentinising peridotite hydration will clearly influence the rheological evolution of the oceanic crust.

Serpentinisation appears to proceed as a non-equilibrium process that takes place mainly under temperatures ranging from ≈350° to 400° C if the stability field of the most common serpentine polytype in these geological settings (lizardite) is considered (e.g. Janecky and Seyfried 1986 and references therein). Consequently, the maximum depth of serpentinisation will grossly coincide with the position of the 400° C isotherm, which is positioned ca. 6 km below the seafloor at the end of the ridge segments in slow-spreading systems (e.g. Escartín et al. 1997a). This corresponds also to the depth of brittle deformation indicated by seismic activity, strongly suggesting that the maximum depth of seismicity in the oceanic lithosphere coincides with the serpentinisation front (e.g. Bergman and Solomon 1990). However, a general aseismic creep is expected, since the deformation of lizardite aggregates is largely accommodated by shear cracks along cleavage planes, probably suppressing dilatancy (Escartín et al. 1997a,b). As a result of this mechanical behaviour, fluids will be confined within already serpentinised rock domains, enhancing subsequent weakening and aseismic slip of fault zones. This will promote strain concentration as well, contributing to the increase in fault spacing and fault throw at the end of ridge segments (e.g. Escartín et al. 1997b).

Crustal extension along listric faults that detach in low angle ductile shear zones at depth are believed to represent a major structural feature on ridge segmentation at slow-spreading systems (e.g. Mutter and Karson 1992). Thus, accepting that these faults provide suitable paths for deep seawater circulation after significant crustal cooling, hydration should affect deep domains of the oceanic lithosphere, further controlling their mechanical response (e.g. Rowlett 1981; Bergman and Solomon 1990; Kong et al. 1992; Wolfe et al. 1995; Escartín et al. 1997a). Late breccia zones and late chemical transformations assisted by fracturing processes are therefore to be expected in deep plutonic units, overprinting many of the early microstructures either caused by syn-magmatic deformation or ascribable to high-T (>600°C) brittle failure (e.g. Agar 1994), and favouring the stable slip of the fault zones. A serpentinised, weakened upper mantle may as well be developed, assuming that the pathways for fluid circulation are not completely sealed as the crust moves away from the spreading centre. This will provide the means for the accommodation of significant aseismic creep in the cooled upper mantle, preventing also the development of critical detachments between this layer and the lower oceanic crust above. Consequently, the assumption that, far from the ridge, the upper mantle responds as a strong, potentially brittle, layer is not free of controversy.

Although the multiple magmatic events in fast-spreading systems and the presence of abundant serpentinites in slow-spreading systems may explain the relative scarcity of seismicity in near ridge ocean domains, the fact is that other processes must be envisaged in order to justify the crustal homogeneous shortening prior to its buckling. The extensive chemical-softening of primary mineral aggregates during crust alteration by means of hydrothermalism is the major candidate, because:

1. it modifies significantly the mechanical characteristics of the rocks, being often tightly related to stress corrosion mechanisms and to subcritical inter- and intragranular fracturing processes (e.g. E.H. Rutter 1976; Fyfe 1976; White and Knipe 1978; Kranz 1983; Atkinson 1984; Carlson and Herrick 1990); and
2. it is dependent on geological processes that are an inevitable consequence of ocean-spreading magmatism and exert a major influence on the chemical and thermal evolution of seawater and ocean crust (e.g. Wolery and Sleep 1976; Fehn and Cathles 1979, 1986; Sclater et al. 1980; Lowell 1980, 1991; Rona 1988; Cann and Strens 1989; Cathles 1990; Rona et al. 1990; Baker et al. 1991; Travis et al. 1991; Alexander et al. 1993; Johnson et al. 1993; Lowell et al. 1995). Hydrothermal activity starts when the crust is young and will continue for tens of Ma (Staudigel et al. 1981; Gallahan and Duncan 1994). However, the nature of hydrothermal activity in seafloor spreading systems is the result of a complex combination of different parameters. These include the bulk permeability, the chemical composition of fluids and the effect of magma chamber geometry on the distribution of near-critical P–T conditions for fluid flow (which are roughly reached near the transition zone between magma and its crystalline envelope; see e.g. Fyfe et al. 1978; Brikowski and Norton 1989; Mottl and Wheat 1994; Fisher 1998).

The uppermost part of the oceanic lithosphere in fast-spreading systems is mostly composed of pillow basalts, lava flows and breccias, often forming complex interfingerings of variable thickness. Permeability values are thus expected to be high, favouring the expansion of the hydrothermal alteration zones during the earlier stages of off-axis migration. During this high fluid/rock interaction, large and successive, although transient, variations in pore fluid pressure may be developed at distinct, but widespread discontinuities (e.g. microcracks having narrow widths and limited lateral extent, features associated with pillow boundaries, and collapse structures). These enable the nucleation/propagation of distinct fracture arrays at different scales that weaken the crustal strength and promote strain concentration. Discontinuous sealing of fractures would also cause irregular redirection in fluid flow throughout the (altered) rocks. As metasomatic processes evolve, fluid-enhanced deformation increases, and pressure solution creep may play an important role in strain accommodation, alternating with fracturing mechanisms (e.g. Gratier and Gamond 1990; Gratier 1993; Gratier et al. 1999).

At slow-spreading systems, differences in fracture geometry and/or hydrothermal alteration in the uppermost crustal domains are also responsible for strong vertical and lateral heterogeneity. Indeed, fracture configurations related to crustal extension and to serpentinisation can be locally reinforced by means of hydrothermal alteration, and the repeated syn-kinematic fluid flow along pre-existing faults may determine their subsequent reactivation, leading to distributed cataclasis or to distinct brecciation patterns. Contacts between gabbros and serpentinised peridotites may evolve into discrete faults with minor vertical displacements or into distributed fracturing along prominent sets of microcracks formed during fluid-rock interaction. Also, as in fast-spreading systems, a general viscoplastic rheology will be expected after significant fracturing and fluid-enhanced deformation (which may also explain the seismic anisotropy usually identified in the upper oceanic crust; e.g. Stephen 1991; Fisher 1998).

Bulk permeability is, however, expected to decrease in depth, as documented by the progressive shrinkage of the hydrothermal alteration intensity in ophiolite complexes (e.g. Richardson et al. 1987; Gillis and Robinson 1988; Nehlig and Juteau 1988; Nehlig et al. 1994; Gillis and Roberts 1999), and predicted by different conceptual and numerical models (e.g. Lister 1983; Lowell and Germanovich 1995). Therefore, the vertical and along-strike widespread fracturing in the sheeted dyke complex would mainly reflect the orientation of stresses developed during spreading, although other mechanisms might be involved in fracture generation, such as thermal craking, volatile-rich magma expansion and differential thermal expansion of pore fluids (e.g. Knapp and Knight 1977; Nehlig and Juteau 1988; Nehlig 1993, 1994; Fisher 1998; Gillis and Roberts 1999). How can magma chambers influence the distribution of rock permeability at and below the sheeted dyke – high-level gabbros transition?

The contact between the sheeted dyke complex and the high-level gabbros hardly represents a real barrier to deep hydrothermal circulation, as revealed by geological and geochemical evidence from ophiolites (e.g. Gregory and Taylor 1981; Nehlig and Juteau 1988). The fact that hydrothermal circulation may reach the plutonic levels is also supported by fluid inclusion data (Vanko 1988), the geobarometry of hydrothermal vent fluids (Campbell et al. 1988), the characteristics of the seismic reflector present at many spreading systems and thought to represent the top of the magma chamber (e.g. Calvert 1995), and the requirement of a thin boundary layer above the magma chamber in order to supply enough heat to support ridge crest vents (e.g. Lister 1977; Cann and Strens 1982; Converse et al. 1984; Gillis and Roberts 1999). This means that chemically modified seawater is able to penetrate the roof of the magma chamber, its relative abundance being sometimes high enough to promote metasomatism in the underlying layered gabbros (e.g. Nehlig et al. 1994; Nehlig, 1993, 1994). However, (long?) before being reached by hydrothermal fluids, these deep crustal domains may already be highly fractured, due to processes of thermal contraction and subsequent volatile-rich phase expansion, which allow efficient transfer of magmatic heat (e.g. Burnham and Davis 1971, 1974; Baker et al. 1987; Nehlig 1994). One may, consequently, presume that the mechanical behaviour of the sheeted dyke complex high-level gabbros transition may change between brittle and ductile, being extremely sensitive to local thermo-elastic stresses related to dyke injection and magma chamber cooling. According to Alexander et al. (1993), the presence of a non-steady state magma chamber would also be important in controlling late development of extensional faulting, causing as well the tilting of the dykes, and certainly related to retrograde metasomatic processes under higher fluid/rock ratios. These events overprint the early alteration processes which, evolving from diffuse grain scale flow, lead to epidosite formation, an hydrothermal product thought to represent the geological record of the root zone of the hydrothermal circulation in fossil fast-spreading systems (e.g. Nehlig et al. 1994, and references therein).

The size and the geometry of oceanic ridge magma chambers are presumably quite variable (e.g. Brikowski and Norton 1989; Sinton and Detrick 1992; Nicolas et al. 1993). Present-day investigations enable the inference of discrete and relatively narrow magma chambers (<2 km; e.g. Detrick et al. 1986; Langmuir et al. 1986; Shirey et al. 1987), although evidence from some ophiolite complexes strongly suggests the development of much larger chambers and/or their gradual widening in depth (e.g. Nicolas et al. 1988). More or less distinct variations in size and geometry of magma chambers are, however, to be expected and may simply represent distinct stages of their evolution, since a wide chamber will quickly reach a confined morphological configuration as magma cools and differentiates. However, apart from this convergent effect, it should be noted that the influence of the size and geometry of the magma chamber on the time-dependent mechanical response of the oceanic crust is probably not negligible. In fact, these parameters explicitly control the heat flow regime, constraining the transfer of the latent heat of magma crystallisation and, subsequently, the rate of removal of the heat of magma cooling by the hydrothermal circulation of seawater, thus indirectly governing the development of metasomatic processes responsible for the major lateral and vertical heterogeneities usually identified in the uppermost oceanic crustal levels (e.g. Brikowski and Norton 1989; Cathles 1990; Mottl and Wheat 1994; Fisher 1998).

A relatively confined near-critical zone for fluid flow is therefore expected for narrow magma chambers, thus favouring the focusing of hydrothermal fluids into very restricted rock domains. Wider magma chambers will promote the development of a broad near-critical domain for fluid flow, leading to widespread hydrothermal activity that will migrate ridgewards with time as magma liquids shrink, evolving naturally to configurations indistinguishable from those related to narrow chambers. The implications of these two expected evolution trends in terms of heat flow and of the time-dependent mechanical response of the oceanic crust are enormous. More so if we consider calculations by Brikowski and Norton (1989) which show that the hydrothermal activity will remain for nearly twice the lifetime of the magma chamber. One may, therefore, conclude that multiple intrusion events at fast-spreading systems will generally reinforce hydrothermal activity patterns, thus enhancing vertical and lateral heterogeneities of the oceanic crust, especially in the presence of wider magma chambers.

From what was previously said, it seems that the accommodation of significant (<10%) homogeneous shortening by the upper oceanic crust in fast-spreading systems is mainly ascribable to the combined effects of renewed magmatic activity, subsequently complemented by vigorous hydrothermal alteration, which also restrain seismicity and support strain partitioning between seismic and aseismic processes. The time-dependent mechanical response is apparently similar for slow-spreading systems, the decrease in the upper crustal strength being extremely variable and largely controlled by faulting and by the relative abundance of serpentinised peridotites. Serpentinisation processes and subsequent hydrothermal alteration of gabbroic and basaltic rocks, coupled by extensive fracturing at all scales, will thus lead to a mechanically weakened crust, which easily accommodates the homogeneous shortening, delaying buckling onset. Deformation by buckling should, therefore, be earlier and more intense in fast-spreading systems, even after lower rates of homogeneous shortening. Accepting this general interpretation, an evolution from (thermo-)viscoelastic to viscoplastic behaviour is expected for the upper oceanic crustal domains as intense fracturing and hydrothermal alteration proceed. Since the heat flow regime that controls those chemical and physical modifications is mainly active-advective (the heat source being magmatic and crustal cooling mainly controlled by the advection of seawater), a thermo-viscoelastic rheology, locally or intermittently viscoplastic, should characterise the time-dependent mechanical response of deeper intrusive units. This is acceptable for the sheeted dyke complex – high-level gabbros transition, as demonstrated before. However, could we extrapolate that mechanical behaviour for the plutonic units?

As mentioned before, narrow, but long-lived magma conduits should lead to multiple intrusive events at fast-spreading systems, while episodic magmatic activity is believed to characterise slow-spreading systems. Accordingly, the new deep crust will tend to have monotonically decreasing temperatures in fast-spreading systems, and to experience successive re-heating events as it slowly moves off-axis in the other case. From this, wider strength crustal variations at slow-spreading systems may be expected, mostly caused by subsolidus deformation far from steady state conditions that will favour heterogeneous high-T (≥600°C) diffusion creep and intracrystalline slip, subsequently overprinted by annealing (e.g. Karson 1990; Agar 1994). Mineralogical and textural adjustments ascribable to diffusion controlled mechanisms (such as grain boundary sliding) and to the effective competition between work hardening and recovery processes are, conversely, expected in younger deep oceanic crust generated at fast-spreading systems, thus suggesting enhancing of synmagmatic viscous flow and quasi-plastic flow under near steady state conditions at temperatures above 600°C (e.g. Karson 1990; Agar 1994). We may, therefore, conclude that, when young, the rheology of deep plutonic units at fast- and slow-spreading systems is also expected to evolve from thermo-viscoelastic to viscoplastic.

As a result of the magmatic-hydrothermal-crustal extension evolution that is believed to characterise both fast- and slow-spreading systems, the oceanic crust is thought to behave as viscoplastic some time after its formation (10–20 Ma, according to previous discussions in this volume). At this time, the heat flow regime in the seafloor at fast-spreading systems should become mainly passive-conductive, remaining largely active-advective at slow-spreading systems, grossly agreeing with the age of transition from advective to conductive heat flow estimated for the Juan the Fuca Ridge (<3 Ma), Galapagos Rift (5 Ma), East Pacific Rise (12 Ma), Indian Ocean (40 Ma) and Mid-Atlantic Ridge (70 Ma; see e.g. Mottl and Wheat 1994, and references therein). Nevertheless, as comprehensively reviewed by Fisher (1998), uncertainties regarding the distribution and evolution of permeability within the ocean crust significantly determine the predictions based on the available conceptual models. Consequently, all we can say is that the age of heat flow regime transition from active-advective to passive-conductive differs notably from one ridge segment to another, being mostly dependent on the sedimentation rates and on the roughness of the basement topography (the latter, in turn, a function of the spreading rate, as discussed before). However, despite all these uncertainties, a general tendency exists for a uniform temperature at the basement interface and for homogenised pore-water composition as the heat flow regime become passive-conductive with time. In terms of mechanical behaviour, this means that a gradual evolution from high-T to low-T diffusion creep is expected for both the strongly hydrothermally altered uppermost crustal domains

and their deeper counterparts. Within the sediment cover formed in the meantime, progressive dehydration facilitates strain accommodation by means of intense (macro-)microfracturing and low-T diffusion mechanisms, locally assisted by differentiated thermal expansion of pore fluids if non-homogeneous conditions at basement enable the development of heat flow anomalies (e.g. Sclater et al. 1980; Stein and Stein 1994). From this, a prevailing aseismic creep is to be expected at off-axis settings, thus favouring the subsistence of a viscoplastic rheology.