# Eclogite Formation and the Rheology, Buoyancy, Seismicity, and H<sub>2</sub>O Content of Oceanic Crust

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A broad spectrum of variably altered igneous rocks with a wide range of grain sizes are compressed and heated over a wide range of pressure-temperature paths in subduction zones. Although experimental kinetic data cannot be extrapolated to predict the rates of blueschist and eclogite formation in nature, textural data from rocks indicate that transformation below temperatures of 150°C is minimal. Complete transformation of volcanic rocks occurs by ~250°C, but incomplete transformation of gabbroic rocks heated to 800°C has been observed. There are important consequences to the rapid transformation of volcanic rocks and the metastable persistence of gabbroic rocks into the blueschist and eclogite stability fields. Fast seismic velocities should be evident first in the upper oceanic crust and may be substantially retarded in the lower oceanic crust. The upper oceanic crust will be denser than asthenosphere before the lower oceanic crust. Early in the process of eclogite formation, volcanic rocks will be placed in deviatoric tension and the underlying coarser grained rocks in compression; with further reaction, the state of stress in gabbroic rocks will change from compressive to tensile. Earthquakes at shallow depths should be extensional in basalt and contractional in gabbro, changing at deeper levels to extensional throughout the crust.

## INTRODUCTION

Formation of eclogite is a key factor in plate tectonics, influencing the size and shape of plates as well as their age and rate of disappearance from the Earth's surface. Subduction zones have long been of interest because they are the builders of continents, collecting material from the ocean basins and constructing magmatic arcs. In spite of the dynamic nature of subduction zones, much of our understanding of their behavior stems from equilibrium thermodynamics. The equilibrium viewpoint has proven a valuable first means of addressing processes in subduction zones, but a more in-depth comprehension requires consideration of the rates and mechanisms by which phase changes occur during subduction.

Subduction: Top to Bottom Geophysical Monograph 96 This paper is not subject to U.S. copyright. Published in 1996 by the American Geophysical Union This paper summarizes the rates and textures of densification reactions in subduction zones by examining experimental and field studies, and then considers the effects of these findings on the rheology, buoyancy, seismicity, and  $H_2O$  content of subducting oceanic crust.

## WHAT ARE BLUESCHIST AND ECLOGITE?

Blueschist facies mafic rocks are characterized by the coexistence of sodic amphibole and lawsonite at low temperature or epidote at high temperature [*Ernst*, 1963]. They may also contain garnet or omphacite. Eclogite *sensu stricto* is mafic rock consisting of garnet and omphacitic clinopyroxene with or without additional minor phases [*Coleman et al.*, 1965]. The geophysical importance of eclogite is that it represents the highest density commonly attained by crustal rocks. Omphacite and garnet have densities of 3.2–3.4 gm/cm<sup>3</sup> and 3.6–4.0 gm/cm<sup>3</sup>, a marked elevation beyond the density of a typical basalt at 2.9 gm/cm<sup>3</sup> [*Carmichael*, 1989]. This shift from positive to negative buoyancy relative to asthenosphere, which has a

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density of roughly 3.23 gm/cm<sup>3</sup> [*Cloos*, 1993], is as important a factor in controlling plate subduction as are conductive cooling of the lithosphere and the olivine  $\rightarrow$  spinel transformation in subducting mantle [*Ahrens and Schubert*, 1975].

#### PRINCIPAL DENSIFICATION REACTIONS

A broad spectrum of reactions cause rocks to become more dense in subduction zones. At the high temperatures found in very hot subduction zones the transformation of basaltic rocks to garnet granulite and then eclogite involves the breakdown of plagioclase, addition of Na<sub>2</sub>O to cliopyroxene and the growth of garnet (Figure 1). PT conditions for the appearance of garnet have been determined by experimental study [*Green and Ringwood*, 1967; *Ito and Kennedy*, 1971; *Liu et al.*, 1993; *Poli*, 1993], and occursat lower pressures/higher temperatures in SiO<sub>2</sub>undersaturated [*Ito and Kennedy*, 1971] and MgO-rich rocks [*Green and Ringwood*, 1967]. The disappearance of plagioclase is less well defined by the same studies and depends more strongly on rock composition.

The transformation of basaltic rocks to blueschist and then eclogite at the low temperatures that prevail in most subduction zones is more complicated because of the prevalence of disequilibrium in nature and experiment. One reaction inferred to have transformed blueschist to eclogite at several localities worldwide is epidote + glaucophane  $\rightarrow$  garnet + omphacite  $\pm$  paragaonite + quartz + H<sub>2</sub>O [*El-Shazly et al.*, 1990; *Ridley and Dixon*, 1984; *Schliestedt*, 1986].

Eclogite facies conditions exist at temperatures >500°C and pressures >1.2 GPa (Figure 1). Eclogite facies *rocks* do not necessarily exist wherever such conditions prevail in the Earth because of sluggish kinetics. All reactions require overstepping of equilibrium to overcome local free energy increases related to attachment or detachment of atoms at interfaces, and all reactions involve diffusion or atom-atom bond rearrangement. Perceptible reaction only occurs at sufficiently high diffusivities and reaction free energies.

#### HOW FAST ARE REACTIONS IN THE EARTH?

#### Experimental and Theoretical Inferences

How long can rocks remain out of equilibrium at elevated pressures and temperatures? Specifically, how long can rocks remain within the eclogite stability field before transforming partially or completely to the stable phase assemblage? Information relevant to this question derives from two sources: experiment and study of natural rocks [*Rubie*, 1990]. Experimental kinetic data collected on



Fig. 1. Metamorphic facies and reactions pertinent to eclogite formation, after [Liou, 1971], [Maresch, 1977], [Brown, 1977; Brown and Ghent, 1983; Evans, 1990], [Green and Ringwood, 1967; Ito and Kennedy, 1971; Liu et al., 1993; Oh and Liou, 1990; Poli, 1993]. Three- and two-digit number for each facies indicate density (g/cm<sup>3</sup>) and H<sub>2</sub>O content (wt%) calculated using NCMASH model of Peacock [1993] assuming equilibrium (i.e., no kinetic hindrance).

powdered materials may not be directly applicable to the majority of geologic situations where rocks are tightly packed crystalline aggregates—principally because the interfacial free energy of a rock is much less than that of a powder. For extrapolation to geologic conditions, data on phase transformations in unpowdered materials are required—information that is exceedingly sparse.

Recent experiments on reactions known to occur in subduction zones demonstrate that most kinetic data cannot be extrapolated to the Earth and that H<sub>2</sub>O has major effects on reaction rate and texture. For instance, *Holland* [1980] observed jadeite growth in 24 hr from powdered albite + quartz at 900°C and a pressure overstep of 50 MPa. In contrast, albite  $\rightarrow$  jadeite + quartz experiments by *Hacker* et al. [1993] on albite rock showed no transformation in 24 hours at 1100°C at a pressure overstep of 500 MPa!

Quantitative transformation-rate measurements are available for the calcite  $\leftrightarrow$  aragonite reaction. As with albite, the calcite  $\rightarrow$  aragonite transformation in Carrara marble [e.g., *Hacker et al.*, 1992] is much slower than in calcite powder. For example, at 600°C and 2.0 GPa confining pressure, powdered calcite transforms 50% to aragonite in <1 hour [*Brar and Schloessin*, 1979, p. 1409], whereas 192 hours were required for similar conversion in unpowdered marble. *Carlson and Rosenfeld* [1981] and *Liu and Yund* [1993] measured the rate of the aragonite  $\rightarrow$ calcite reaction; extrapolated to geologic timescales and grain sizes, their data imply that 1 mm aragonite grains will revert completely to calcite in less than 1 m.y. at temperatures as low as 200°C. *Hacker et al.* [1992] explored the reverse transformation in marble and found a similar result, that the calcite  $\rightarrow$  aragonite transformation occurs within less than 1 m.y. at 200°C for oversteps of <100 MPa. The reaction depends strongly on temperature, such that even several GPa overstep at temperatures of <100°C are unlikely to promote transformation.

The catalytic action of H<sub>2</sub>O is well known from kinetic studies of powders and rocks. *Hacker and Kirby* [1993] deformed polycrystalline, vacuum-dried albite rocks at strain rates of  $10^{-4}$ – $10^{-6}$  s<sup>-1</sup> at temperatures of 600–800°C and confining pressures of 1.0–2.0 GPa. In spite of extreme pressure overstepping and extreme sample strain, deformed anhydrous samples contained no jadeite. Differential stresses as high as 2000 MPa produced maximum normal stresses up to 4.0 GPa—all at temperatures where the equilibrium pressure is <2.0 GPa. Some of these samples were strained 70%, and individual grains attained aspect ratios of greater than 10:1. In marked contrast to this, the addition of 1 wt% H<sub>2</sub>O produced partial reaction at temperatures as low as 600°C, even in undeformed samples.

More complicated experiments measured the olivine  $\rightarrow$  spinel transformation in hot-pressed Mg<sub>2</sub>GeO<sub>4</sub> [Burnley, 1990] and Ni<sub>2</sub>SiO<sub>4</sub> [Rubie et al., 1990] at pressures and temperatures of 800–1900°C and 1–15 GPa. When extrapolated to geologic conditions these investigations suggest that transformation may be suppressed to depths 150–200 km below the equilibrium boundary in rapidly subducting slabs.

In summary, existing experimental data indicate that  $H_2O$  has a major catalytic affect on reaction rates and that some reactions are geologically fast (e.g., calcite $\rightarrow$ aragonite) whereas others are probably slow (olivine  $\rightarrow$  spinel). We have essentially no experimental information on reactions forming blueschist and eclogite.

#### Metamorphic Textures

Observations on rocks show that transformation to blueschist and eclogite may be suppressed long after rocks have passed into the respective stability fields of these assemblages. Figure 2 was constructed from studies worldwide that reported not only peak metamorphic pressures and temperatures but also textural information about the degree of reaction. Rocks reported as feebly recrystallized or whose high-pressure phases were so minor as to require detection by electron microscopy are shown in Figure 2 as "feeble, 25%." Rocks containing abundant relict phases (usually clinopyroxene) or partially developed



Fig. 2. Reported extents of transformation, pressures, and temperatures, for blueschist and eclogite facies metamorphism worldwide. References and data used to compile figure are available from the author on request.

reaction coronas are denoted as "moderate, 50%", and those with rare relict phases are labeled "extensive, 75%". Metamafic rocks specifically described as lacking relict phases are shown as "complete, 100%".

There are some important limitations to Figure 2. 1) Reported peak pressures and temperatures may not have occurred at the same time and, at best, represent only a single way-station on the PT path experienced by the rocks. 2) How long the rocks were at the reported pressures and temperatures is unknown. 3) Pressure estimates for many rocks are minima because of the high variance of the phase assemblages; these are denoted by U-shaped lines rather than rectangles. 4) Many high-pressure rocks experienced substantial post-high pressure alteration that likely consumed existing relict phases; studies reporting severe alteration of this type were excluded from consideration. 5) Fine-grained rocks react faster than coarse-grained rocks because of increased interfacial free energy; gabbroic rocks are only partially transformed at conditions where volcanics are completely recrystallized. Numerous papers that do not specifically state whether relict phases are present or absent had to be excluded from Figure 2, but this may simply mean that the absence of relics was not significant for the purposes of the study. Thus Figure 2 should be taken as an illustration of the transformation behavior of coarsegrained, i.e., gabbroic rocks.

With these limitations in mind, Figure 2 indicates that feebly metamorphosed rocks generally persist to temperatures of 200–250°C. Moderate transformation begins at 200–250°C and is found in rocks that were heated to just above 600°C. There are a few cases of extremely

limited reaction at high temperature shown by dashed boxes in Figure 2. Incomplete conversion of gabbro to eclogite is known to have happened at temperatures >700°C and pressures of 1.5-2.0 GPa in the presence of aqueous fluid [Mørk, 1985]. Suppression of transformation of gabbro to eclogite was reported from rocks that reached 800-900°C and pressures of >2.8 GPa in the absence of fluid [Zhang et al., 1995]. Oddly enough, there is little difference in the PT range over which extensive to complete transformation have been found. Both have been observed from rocks heated to less than 300°C and yet partial transformation of gabbro to eclogite has occurred at 760-850°C and 1.6-2.0 GPa [Indares and Rivers, 1995]. In general, 500-600°C marks the transition to extensive or complete reaction for most rocks. A few cases, however, indicate that the transformation of gabbro to eclogite may be suppressed to much higher temperatures.

Fe-rich rocks transform to eclogite along high P/T trajectories more readily than Mg-rich rocks because reactions involving Fe-rich phases generally occur at lower temperatures [Bohlen et al., 1983]. Field evidence of this—MgO-rich rocks with abundant igneous relics adjacent to Fe-rich rocks with rare relict phases—has been noted at Bardoney Valley in the Alps [Reynard and Ballevre, 1988], at Flemsøy, Norway [Mørk, 1985], and in Piedmont ophiolites [Pognante, 1991; Sandrone et al., 1986].

Eclogite formation is faster in finer grained rocks. For example, in the Bardoney Valley of the Alps, abundant relict augites are present in gabbros, but absent in metavolcanics [*Reynard and Ballevre*, 1988]. In the Sesia Lanzo zone farther east, amphibolite transformed to eclogite contains relics in coarse rocks but none in finegrained rocks [*Lardeaux and Spalla*, 1991].

Not only is  $H_2O$  proven to accelerate reactions in the laboratory, numerous localities demonstrate that this is true in nature as well. Mafic granulite in the Musgrave Ranges of Australia cooled into the eclogite stability field and, though the bulk of the rock is unaltered, local shear zones contain omphacite + garnet + minor zoisite [*Ellis and Maboko*, 1992]; the presence of zoisite implies that the reaction was fluxed by H<sub>2</sub>O. At Holsnøy in the Bergen Arcs, zoisite-bearing eclogite haloes in undeformed rock surrounding deformed veins also indicate that fluid drove reaction [*Boundy et al.*, 1992; *Klaper*, 1990]. Deformation is involved in these fluid-mediated transformations inasmuch as the fluid must make its way to the reaction site via cracks.

### COMPOSITION OF IGNEOUS OCEANIC CRUST

To understand the transformation of oceanic crust to blueschist and eclogite we must know the composition and structure of oceanic crust. Oceanic crust has been studied by dredging and drilling in ocean basins, and by the examination of ophiolites. The Samail ophiolite in Oman is the best-exposed, largest, least-deformed, and perhaps most-studied ophiolite in the world. Plutonic rocks of the Samail ophiolite are divisible into a lower layered sequence and an upper isotropic section. The lower 2.3 km (avg.) is well-foliated and lineated, layered ultramafic (chiefly wehrlite, dunite, and clinopyroxenite) through mafic cumulates [Pallister and Hopson, 1981]. Olivine predominates at the base of the section, whereas plagioclase is dominant at the top [Juteau et al., 1988b]. Three-quarters of the section is gabbro-typically olivine-clinopyroxene or clinopyroxene gabbro with minor hornblende and FeTi oxides-but at rare localities it is gabbronorite [Juteau et al., 1988a]. The gabbro consists of mm-size grains of 55% plagioclase, 35% clinopyroxene, 10% olivine, and rarely Ti-pargasite [Browning, 1984]. Overlying the layered section are 200-900 m of isotropic high-level mafic rocks, chiefly hornblende-clinopyroxene gabbro [Pallister and Hopson, 1981]. Grain sizes range from mm to pegmatitic [Juteau et al., 1988a]. Sheeted dikes contain abundant to sparse plagioclase and augite [Hopson et al., 1981]. The volcanic rocks are non- to moderately vesicular, almost aphyric pillows and rare brecciated flows and massive flows [Ernewein et al., 1988].

Mid-ocean-ridge basalts (MORBs) typically are glassy or contain plagioclase with less olivine [Hekinian, 1982], while rocks crystallized in intraoceanic magmatic arcs generally contain plagioclase and augite [Ewart, 1982]. Midocean-ridge basalts (MORBs) are subalkaline tholeiites that typically contain plagioclase with less olivine [Hekinian, 1982]. Oceanic plateaus, which make up a sizeable proportion of oceanic crust, are similar in their major element chemistry to MORB, but are 10–40 km thick [Cloos, 1993]. Hotspot lavas and propagating rifts are enriched in Fe and Ti relative to MORB [Flower, 1991]. Lavas produced at fast spreading ridges such as the Pacific have lower Mg/Fe ratios than MORB erupted at slow spreading ridges [Batiza, 1991].

### ALTERATION OF IGNEOUS OCEANIC CRUST

Because phase transformations are strongly influenced by the availability of H<sub>2</sub>O and the extent and type of preexisting alteration, understanding the alteration of oceanic crust is critical to evaluating when and where phase transformations occur. Variable alteration of oceanic crust to lower grade mineral assemblages occurs by active hydrothermal circulation at spreading centers and by later weathering on the seafloor [*Alt et al.*, 1986; *Humphris and Thompson*, 1978]. Some rocks escape alteration entirely, while others are wholly recrystallized to new minerals. Most typically, lavas are variably metamorphosed to subgreenschist facies, dikes change partially to greenschist facies, and gabbro recrystallizes weakly to amphibolitefacies minerals. The pressures at which these metamorphic assemblages form are relatively low ( $\leq 300$  MPa; determined by ocean depth and the thickness of oceanic crust) and the formation temperatures may range up to ~650°C.

Our best information about the alteration of oceanic lavas and dikes comes from DSDP hole 504B [Alt et al., 1989; Alt et al., 1986]. There the lavas contain ~10% alteration haloes around veins and cracks, leading to about 1 wt% H<sub>2</sub>O in the bulk rock. Veins and vugs contain Fehydroxides, clay, celadonite, smectite, zeolites, and carbonate; groundmass minerals are altered to Fehydroxides and clay; plagioclase is partly altered to smectite: olivine is partly to totally replaced by smectite and minor calcite; glass is partly to completely replaced by clay; and pyroxenes are unaltered. Alteration in the dikes increases down section to a maximum of ~50 vol%. Olivine is replaced by chlorite, clay, and talc; plagioclase is partially to totally replaced by albite + zeolites, chlorite, or clay; and augite is unaltered high in the section and rimmed by actinolite deeper in the section. The H<sub>2</sub>O content of the dikes averages 2 wt% [Alt et al., 1989; Alt et al., 1986].

Good information on the alteration of plutonic rocks comes from the Samail ophiolite and ODP core holes. Alteration of the Samail layered plutonic rocks is limited to rare replacement of clinopyroxene by actinolite [Lippard et al., 1986] and alteration of the isotropic plutonic rocks is limited to the growth of up to 10% actinolite and FeTi oxides, and the infilling of cavities with epidote + sphene + prehnite + quartz [Ernewein et al., 1988]. In contrast, gabbroic rocks drilled from ODP hole 735B in the Indian Ocean show widespread amphibolite-facies alteration from 0-100%, with an average of perhaps 20-30% [Dick et al., 1991]. Olivine and pyroxene are replaced by hornblende, there is partial to complete alteration of olivine to colorless amphibole and talc, and 2.4 vol% of the rock consists of hydrothermal and late magmatic veins composed of hornblende or oligoclase + diopside + epidote. H2O contents of the 735B gabbros are mostly 0.5-1.5 wt% [Robinson et al., 1991].

## EFFECTS OF SLOW TRANSFORMATION ON DENSITY AND H<sub>2</sub>O CONTENT

When oceanic crust is subducted it is compressed and heated along one of a wide variety of possible PT paths (Fig 3). The highest-temperature path into the eclogite





2.89

2.90

0.8

0.6

stability field is from the granulite facies. In the hottest of subduction zones involving lithosphere less than a few m.v. old, it is possible for oceanic igneous rocks to pass into the eclogite stability field by this path [Hacker, 1991]. Rocks that recrystallized in the subgreenschist facies during hydrothermal alteration at the mid-ocean ridge will pass through the greenschist and amphibolite facies prior to entering granulite conditions, and may transform variably to greenschist, amphibolite, or granulite facies assemblages prior to entering the stability field of eclogite. At the opposite end of the temperature spectrum, rocks that previously recrystallized in the prehnite-pumpellyite facies will pass through the blueschist facies prior to entering eclogite conditions [Peacock, 1990], and may thus transform variably to blueschist prior to entering the eclogite stability field.

*Peacock* [1993] used the restricted compositional system NCMASH (Na-Ca-Mg-Al-Si-H-O) to calculate the mineral modes and maximum H<sub>2</sub>O contents of subducting mafic rocks. He demonstrated that moderately altered oceanic crust containing 1-2 wt% H<sub>2</sub>O begins to dehydrate at the onset of eclogite or amphibolite facies metamorphism, and suggested that the transition from blueschist to eclogite facies, associated with the breakdown of lawsonite or clinozoisite, releases the most H<sub>2</sub>O during subduction.

Peacock calculated a range of paths that traverse most of the region of PT space relevant to subduction zones. Figure 3 illustrates two paths for the uppermost and lowermost parts of subducted 7-km thick oceanic crust-calculated for a subduction velocity of 50 mm/a and shear stresses of 100 and 33 MPa. Most of the PT paths followed by the uppermost layer of the crust (basalt) are limited by the paths identified by upright triangles, and most of paths followed by the lowermost layer of the crust (gabbro) are limited by the paths with inverted triangles. Intermediate levels in the crust follow intermediate PT paths. For path 1, the upper crust passes progressively through the greenschist to amphibolite to eclogite facies conditions, while the lower crust evolves from blueschist to eclogite facies. Along path 3, all the crust remains at blueschist facies conditions to pressures of 2.5 GPa. For paths intermediate between 1 and 3, most of the crust passes through the blueschist and eclogite facies.

In the spirit of Peacock's calculations and using the information presented in Figure 2, assume that coarsegrained rocks (i.e., gabbro) transform 25%, 50%, 75%, and 100% at 150°C, 250°C, 500°C and 550°C, respectively, in accordance with field observations described earlier. Further assume that fine-grained rocks (i.e., basalt) transform twice as fast—such that they reach 100% transformation at 250°C. Assume also that oceanic crust is made of 3 km of basalt and 4 km of gabbro. Given these assumptions, what conclusions can be drawn about the rates of eclogite formation in subduction zones and its effect on rock physical properties? Possible variables to be considered include the amount of  $H_2O$  and the composition of the crust.

Figure 3A and 4A illustrate the predicted densities and H<sub>2</sub>O contents of NCMASH assemblages determined by *Peacock* [1993] for the various metamorphic facies. Figures 3B and 4B repeat this theme, using the kinetic inferences discussed above. Figures 3C and 4C use the same kinetic hindrance considerations but begin with altered protoliths. Because these figures are based on modeling real assemblages with a subset defined by the NCMASH system, the predicted densities and H<sub>2</sub>O contents cannot be considered exact; however, several interesting features are apparent.

Figures 3B and 4B predict the behavior of subducted basalt and gabbro using the mineral assemblage olivine + orthopyroxene + clinopyroxene + plagioclase as a protolith. In the upper, volcanic crust, 50% reaction is assumed to occur at 150°C and complete reaction at 250°C; in the lower plutonic crust, 25, 50, 75, and 100% reaction is modeled as happening at 150, 250, 500, and 550°C, respectively. For example, the density of 2.97 g/cm<sup>3</sup> for lawsonite-blueschist facies gabbro in Figure 3B was calculated using unaltered gabbro (2.89 g/cm<sup>3</sup>) transformed 50% to lawsonite blueschist (3.10 g/cm<sup>3</sup>). The H<sub>2</sub>O required for hydration in Figures 3B and 4B is assumed to derive from sediments or hydrated mantle. In the absence of sufficient H<sub>2</sub>O, all facies other than CE, EC, GG, and GN have a density of 2.89 g/cm<sup>3</sup> and 0 wt% H<sub>2</sub>O.

The results for volcanic rocks are close to the predictions of Peacock's equilibrium model (Figure 1) because the transformation rate is rapid. Basalts traveling via path 1 undergo only slight changes in density prior to the appearance of garnet and the disappearance of plagioclase. In contrast, Path 3 indicates that 5-10% volume loss occurs in basalts upon entering both the blueschist and eclogite facies. As in the equilibrium case (Figures 3A and 4A), the biggest step in dehydration occurs at the blueschist/eclogite facies boundary; dehydration along path 1 is split subequally between the greenschist/amphibolite and amphibolite/granulite transitions. Basalts moving along path 3 or slightly warmer trajectories lose roughly equal amounts of H<sub>2</sub>O when entering and leaving the epidote blueschist stability field. Compared to basalt, where volume loss occurs fairly evenly over the 150-450°C temperature range, half the volume loss of gabbro occurs in the 500-550° interval. In contrast to basalt, which may undergo substantial hydration by 250°C, gabbro is predicted to



undergo hydration to temperatures up to  $550^{\circ}$ C—although at any given condition the amount of H<sub>2</sub>O in gabbro is less than in basalt. Again, as in the equilibrium model, the bulk of the dehydration occurs at the blueschist/eclogite transition.

Figures 3C and 4C predict the behavior of subducted altered basalt and gabbro using the prehnite-pumpellyite facies mineral assemblage chlorite + albite + pumpellyite + quartz + calcite for altered basalt and the amphibolite-facies assemblage hornblende + clinozoisite + chlorite + quartz for altered gabbro. The predictions are qualitatively similar to Figures 3B and 4B, except that the H<sub>2</sub>O content at any chosen P and T is higher.

C Kinetic model with prehnite-pumpellyite and amphibolite facies protoliths



Fig. 4. Densities (g/cm<sup>3</sup>) and H<sub>2</sub>O contents (wt%) of subducted crust, based on Figure 3. A) Equilibrium model with essentially unaltered crust. B) Kinetic model with unaltered crust. C) Kinetic model with crust altered to subgreenschist-facies and amphibolite-facies minerals. Compare with [*Peacock*, 1993] Figure 4. Thicker lines are facies boundaries and thinner lines represent changes from 25% to 50% to 75% to complete transformation. The thickness of the crust is stretched by a factor of 5 for clarity. Elongate rectangle at bottom of each panel shows density difference relative to asthenosphere (3.23 gm/cm<sup>3</sup>).

#### DISCUSSION: GEODYNAMIC IMPLICATIONS

The retardation of reaction during subduction illustrated in Figures 3 and 4 will affect 1) how the structure of subducted crust is interpreted from seismic data, 2) the buoyancy forces that produce sinking and bending of the slab, 3) the distribution of stresses induced by volume changes, 4) seismicity in the slab induced by volumechange-related stresses, 5) the mechanical behavior of the slab, 6) the distribution of H<sub>2</sub>O in the slab, and 7) the generation of arc magmatism. All these factors differ from the equilibrium state because the densification reactions in descending slabs depend on grain size, H<sub>2</sub>O content, bulk composition, and temperature. Relative to the equilibrium model, transformation to blueschist and eclogite is always delayed to greater depths in the kinetic model.

The first rocks to transform will be glassy basalts, followed by holocrystalline basalts, then diabases, and finally gabbros. Note that this derives from thermal as well as kinetic causes, as the upper crust is warmer than the lower crust as well as being finer grained. The seismic velocity signature of dense rock should be evident first in the upper crust and may be substantially retarded in the

# 344 ECLOGITE FORMATION IN OCEANIC CRUST

lower crust. The absence of fast velocities at eclogite facies conditions cannot be taken to indicate the absence of oceanic crust. High densities will appear first in the upper crust and may be considerably suppressed in the lower crust. In Figure 4, the lower crust is not only less dense than asthenosphere, but also less dense than the basaltic upper crust. In these situations the upper crust may sink in the asthenosphere but the lower crust cannot. Moreover, Kirby et al. [1996] calculated that densification reactions in subducting crust can place the crust in deviatoric tension and the mantle lithosphere in deviatoric compression. The difference in reaction rate between basalt and gabbro means that their calculations are oversimplified. The initiation of densification in the volcanic rocks will initially place those rocks in tension and the underlying coarser grained rocks in compression. As reaction progresses into the coarser lower crust, the state of deviatoric stress in the gabbroic rocks will change from compressive to tensile. If there were no other forces at work in subduction zones, earthquakes at shallow depths would be extensional in basalt and contractional in gabbro, changing as the rocks descend deeper in subduction zones to extensional throughout the oceanic crust.

Little is known about the mechanical behavior of blueschist facies minerals, but clinopyroxene and garnet are two of the strongest phases in the crust [Ji and Martignole, 1994; Kolle and Blacic, 1983] and eclogite should be strong compared to altered basalt. The replacement of finegrained volcanic phases by blueschist or eclogite facies minerals may result in strengthening, whereas the alteration of coarse-grained minerals in gabbro to blueschist or eclogite phases may enhance deformation [Rubie, 1983]. The volcanic layer, though hotter, may actually become stronger than the gabbroic layer as a result of eclogite formation. It is conceivable that such rheological layering could result in a downward jump of the subduction decollement from the volcanic into the gabbroic layer, resulting in underplating of eclogitized upper crust to the mantle hanging wall and continued subduction of the gabbroic layer.

Crust that undergoes slow transformation is less capable of containing H<sub>2</sub>O. Figure 4 shows that a slowly transforming gabbro layer has a maximum H<sub>2</sub>O content ~1/2 that of the equilibrium model. The flux of H<sub>2</sub>O carried into the mantle beyond a depth of 70 km is  $1.3 \times 10^8$  grams per meter per year in the equilibrium model, but only 0.48 x  $10^8$  grams per meter per year in the kinetic model with unaltered protolith. Note that if the only source of H<sub>2</sub>O in a subduction zone is the igneous crust, situations such as Figure 4C, where the H<sub>2</sub>O content increases with depth, are impossible. Only if there is another source of H<sub>2</sub>O in the subduction zone, from subducted sediments or mantle, can the H<sub>2</sub>O content of igneous crust increase in subduction zones.

Kirby and Hacker [1991] proposed a link between arc volcanism, intermediate depth earthquakes, and eclogite formation. They noted that hypocenters of subduction zone earthquakes in the depth range 50-250 km frequently coincide with arc volcanoes, and proposed that these intermediate-depth earthquakes might be caused by delayed eclogite formation within subducting crust. This hypothesis requires that eclogite does not form at shallower depths because low temperatures and low H<sub>2</sub>O activity hinder transformation, and that the devolatilization of amphibole in the 50-250 km depth range triggers eclogite formation and releases volatiles that lead to the overlying volcanoes. Figure 2 suggests that transformation to eclogite occurs at much shallower depths in most subduction zones. For Kirby and Hacker's hypothesis to be correct, the subducting crust must be essentially anhydrous.

Calculations and conclusions presented above apply principally to normal mid-ocean ridge crust. Ocean crust that is anomalously thick or of unusual composition will transform differently. Oceanic plateaus, which are up to 6 times thicker than mid-ocean crust, but compositionally similar, will have somewhat retarded transformation because of the additional time required for the subducting material to heat; this may be partly offset however by the greater thickness of volcanic rocks, which will transform more rapidly. Field observations discussed above indicate that Fe-rich rocks transform notably faster than MgO-rich rocks. Hydrothermal alteration at spreading centers often produces marked MgO enrichment in lavas [Humphris and Thompson, 1978], which would be expected to slow transformation. Oceanic islands built by hotspots and crust from propagating rifts are both enriched in Fe relative to MORB, and will transform more rapidly than standard midocean-ridge crust. Crust erupted at fast spreading ridges (>60 mm/a) is also richer in Fe than crust erupted at slow spreading (<50 mm/a) ridges [Niu and Batiza, 1993; Sinton and Detrick, 1992], and is expected to transform more rapidly.

#### CONCLUSIONS

The transformation of mafic rocks to blueschist and eclogite in subduction zones is complex, encompassing a wide variety of igneous mineral assemblages and grain sizes, variably developed and equilibrated metamorphic mineral assemblages, all pressurized and heated via a broad range of PT paths. Existing experimental kinetic data cannot be extrapolated to actual petrotectonic settings in order to predict where rocks transform to blueschist and eclogite. However, textural data from exhumed subduction zones worldwide indicate that little transformation occurs at temperatures below 150°C. Volcanic rocks are completely transformed by perhaps 250°C. Coarser gabbroic rocks rarely avoid complete eclogitization at temperatures above 550°C, although examples of incomplete transformation are known from rocks heated at temperatures as high as 800°C.

The rapid transformation of volcanic rocks and the metastable persistence of gabbroic rocks into the blueschist and eclogite stability fields has several implications. The seismic velocity signature of dense rocks should be evident first in the upper crust and may be substantially retarded in the lower crust. The buoyancy forces causing the slab to sink will appear first in the upper crust and may be considerably suppressed in the lower crust. Earlier formation of garnet in the upper crust may cause the upper crust to become stronger than the gabbroic layer, and may lead to underplating of eclogitized upper crust and continued subduction of the lower crust. The initiation of densification in the volcanic rocks will initially place those rocks in deviatoric tension and the underlying coarser grained rocks in compression. As reactions progress into the coarser lower crust, the state of stress in the gabbroic rocks will change from compressive to tensile. If there were no other forces at work in subduction zones, earthquakes at shallow depths would be extensional in the basalt and contractional in the gabbro, changing at deeper levels to extensional throughout the crust. All these effects will be minimized in Fe-rich crust produced at oceanic islands and fast spreading ridges and most pronounced in Mg-rich crust formed at slow spreading centers.

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### 346 ECLOGITE FORMATION IN OCEANIC CRUST

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