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Geophysical and Geochemical Constraints at Converging Plate Boundaries—Part I: Dehydration in the Downgoing Slab*

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Summary

Geochemical constraints based upon compositions and mechanisms of formation and eruption of island arc volcanic rocks strongly suggest hydrous conditions in the region of primary magma generation. The source of most of the water is probably dehydration of the oceanic crust during underthrusting. Chemical reactions involving dehydration of representative oceanic crustal assemblages absorb enough heat to possibly negate any frictional heat which may be generated along the Benioff–Wadati zone in the upper 75 to 100 km of the underthrust zone. Calculations of thermodynamic parameters of relevant reactions at elevated pressures and temperatures show that ~ 50 cal of heat are absorbed per gramme of oceanic crust dehydrated. The result of this heat absorption is suppression of isotherms not only in the oceanic crustal section of the downgoing slab but also in the overlying mantle wedge. Therefore, low heat flow measured from the trenches to the volcanic arcs may be caused in part by the heat-sink effect of dehydration in the downgoing slab.

The water released by dehydration either remains in the slab promoting its melting and the generation of andesitic primary magmas and/or escapes into the overlying wedge of mantle promoting its melting and the generation of the basaltic or andesitic primary magmas. The major thermal effect of the released water is to cool the wedge overlying the slab. The progressive geochemical changes in arc volcanism both in space and time may be controlled by the thermal conditions at the downgoing slab-peridotite wedge interface.

Introduction

Within plate tectonics (Morgan 1968; Le Pichon 1968; Isacks, Oliver & Sykes 1968) perhaps the most complex of plate boundaries is that of convergence where oceanic lithosphere is underthrusting either continental or oceanic lithosphere. Many models for the geophysical structure within and around the downgoing slab have been proposed (McKenzie & Sclater 1968; McKenzie 1969; Minear & Toksöz 1970; Oxburgh

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& Turcotte 1970; Hasebe, Fujii & Uyeda 1970; Griggs 1972; Toksöz, Minear & Julian 1971; Andrews & Sleep 1974, among others) which are variously constrained to reproduce one or more of the following observations. (a) Melting must occur below volcanic arcs. Most underthrusting zones which have seismicity are associated with active volcanism above the downgoing slab. It is most likely, therefore, that this volcanism is intimately related to subduction, although subduction of the cold slab would not normally be considered a favourable process for the generation of high temperatures or melting. Geophysical models, thus, must provide a special mechanism for generation of melting either within the underthrusting slab or in the surrounding mantle. (b) The heat flow pattern in the region of subduction provides another boundary condition. Distinct zones of low and high heat flow respectively, exist between the oceanic trench and volcanic front and behind the arc. Low heat flow is consistently measured between the trench and the volcanic arc; high heat flow measurements dominate the region from the volcanic front through any marginal seas which might be active to the ' landward ' boundary of the subduction zone. This heat flow pattern suggests that the wedge above the downgoing slab and bounded by the front of arc volcanism and the trench is essentially ' cold ', and the region landward of the arc has high heat generation somehow associated with underthrusting. This zonal pattern of cold and hot regions seems to be a characteristic of orogeny associated with slab subduction or of the Pacific-type orogeny accompanied by paired metamorphic belts (Miyashiro 1961; Matsuda & Uyeda 1971; Oxburgh & Turcotte 1970). Further, the mechanism of transfer of heat to the surface below marginal basins must be dominated by some form of bodily transfer since ordinary conductive heat flow is too slow a process to account for high heat flow observed behind volcanic arcs (McKenzie & Sclater 1968). This bodily transfer of heat is most likely in the form of convection induced by magma generation at depth and subsequent intrusion to the surface. (c) Additional constraints are imposed by the seismicity of the subduction zone. The location of earthquakes within the slab not only delineates its general location between the surface and 700 km depth but demands that the slab be relatively cold in order for brittle failure to occur to such depths. This has been verified by studies of seismic attenuation and travel time anomalies (Utsu 1967, 1971; Oliver & Isacks 1967). On the other hand, surface wave studies have shown (Kanamori & Abe 1968; Abe & Kanamori 1970) that the upper mantle velocities are lower under the back-arc basins than under basins oceanward of trenches. Seismic attenuation also shows that the active zone from the volcanic front landward is an area of excessive thermal gradients and of probable partial melt (since low Q is found there), and additionally, that the wedge from the trench to the arc is cold and solid since this is a zone of normal Q (Barazangi & Isacks 1971). These observations are generally in agreement with the thermal structures inferred from heat flow data.

We consider here some geophysical implications of the geochemistry of the subduction process by noting that the oceanic crust is probably composed of a hydrous mineral assemblage with up to 6 per cent by weight water (Wyllie 1971). Raising the pressure and temperature of oceanic crust by underthrusting will result in dehydration. Geochemically, this dehydration has been considered the primary source of water necessary to melt the slab and/or the peridotitic wedge for the production of arc volcanism (Wyllie 1971; Green 1973; Kushiro 1972; Nicholls & Ringwood 1972, 1973). We have quantitatively evaluated the magnitude of heat absorbed by these endothermic chemical reactions and find that, though they have been totally neglected as a major heat sink by all geophysical models *except* Hasebe *et al.* (1970), they absorb large quantities of heat in the upper 75–100 km of the underthrusting zone. We then qualitatively consider the implications of this additional heat sink on the spatial and temporal variations in the thermal structure of the downgoing slab and mantle wedge, and we evaluate the effects of the newly-released water on melting in the slab and in the overlying wedge.

Volcanic arc magmatism and dehydration in the downgoing slab

It is probable that volcanic arc primary magmas have high H_2O contents for the following reasons:

(1) Island arc volcanoes are characteristically much more explosive than other volcanoes, conceivably because of high H_2O and volatile content of the magma: The percentage of tephra in the total volcanic products was defined as explosion index ' E ', by Rittman (1936). Circum-Pacific volcanoes have E generally higher than 80, whereas E rarely exceeds 3 in oceanic island volcanoes.

(2) Subduction requires the underthrusting of a cold lithospheric plate deep into the mantle. The geophysical setting thus suggests that this environment would be less conducive to magma generation than other regions since the slab is a very large heat sink. It is thus very difficult to generate melting temperatures along the slab even with greatly depressed melting points which would result under hydrous conditions. It is unlikely that melting could occur in a dry environment.

(3) Though experimental petrology suggests several possible source materials and source regions for volcanic arc magmatism, many of these experiments indicate hydrous conditions are necessary for the generation of the calk-alkaline rock series that is characteristic of well-developed arcs (see later for references).

A reasonable, at least partial source of this water is from dehydration of the oceanic crust during underthrusting since, while water may be present in the low velocity layer (LVL) and, indeed, may be the cause of the LVL, it may not be present in quantities greater than a few tenths of one per cent because too great a volume of partial melt would be generated (Wyllie 1971).

Theoretical models for underthrusting versus observations

Before reviewing the evidence that water for dehydration is available in the oceanic crust and before considering the implications of dehydration of this hydrous crust on the thermal structure of the slab, it is necessary to review the theoretical models and observational constraints which, to date, have often presented conflicting representations of the thermal structure of the downgoing slab.

Heat flow observations

Heat flow in the areas of trench-arc systems has been measured extensively for the Kurile–Japan–Izu–Bonin–Mariana system, Nankai Trough-Ryukyu system (Uyeda & Horai 1964; Vacquier *et al.* 1966; Yasui *et al.* 1968, 1970; Watanabe *et al.* 1970), Melanesian and Tonga-Kermadec systems (Sclater, Ritter & Dixon 1972), Aleutian system (Foster 1962; Langseth & Horai 1976; and the Java Trench (Vacquier & Taylor 1966).

Almost universally, high heat flow has been observed in the shallow marginal basins landward of arcs. Another interesting feature is the low heat flow in the zones oceanward of the arcs between the trench and the magmatic front. Fig. 1 shows some of the examples of heat flow profiles across volcanic arcs (Vacquier *et al.* 1966; Watanabe 1975). Though heat flow is very difficult to measure in the region of the landward slope of a trench, the existence of the low zone appears particularly well established in Japan where land measurements support the oceanic results.

Heat flow measurements in continental arc margins (i.e. Western South American margin) are much less complete (Diment *et al.* 1965; Uyeda & Watanabe 1970; Sclater, Vacquier & Rohrhirsch 1970), so that distribution of high heat flow in the back-arc area has not been clearly delineated if it exists, but the low heat flow in the

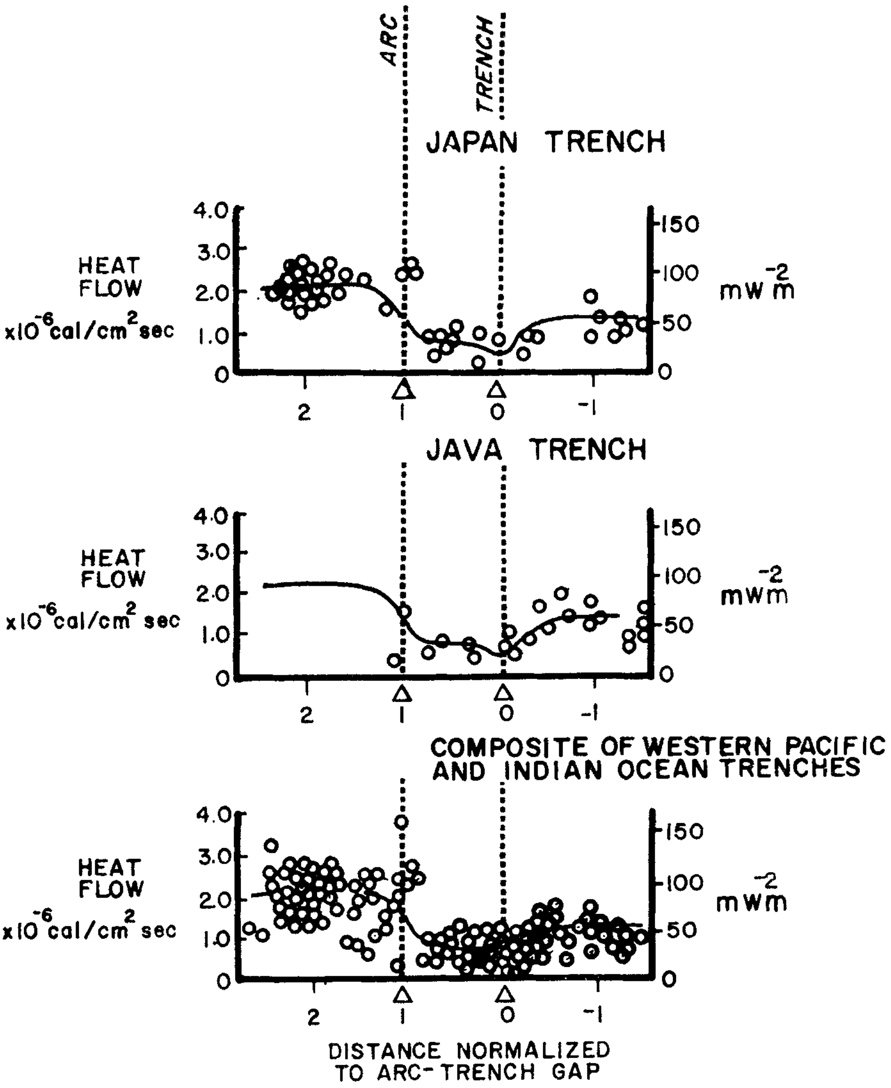


Fig. 1. Heat flow profiles in and near the trenches of the western Pacific and Indian Ocean. These are the only trench systems for which we have adequate coverage between the trench and volcanic arc. Data sources are listed in text.

zone between trench and magmatic front seems to be real. These and other geophysical observations permitted the construction of a general cross-section of the north-east Japan arc (Sugimara & Uyeda 1973), of the Java Trench (Vacquier & Taylor 1966), and a composite heat flow profile from all information available near the western Pacific and Indian Ocean trenches (Fig. 1).

Geophysical models

The first geophysical models of the downgoing slab (McKenzie & Sclater 1968; McKenzie 1969) noted that the introduction of the cold lithospheric plate into the asthenosphere would (a) cause heat to conduct into the slab, (b) cause frictional heating on the boundaries of the cold slab and/or (c) induce secondary flow in the

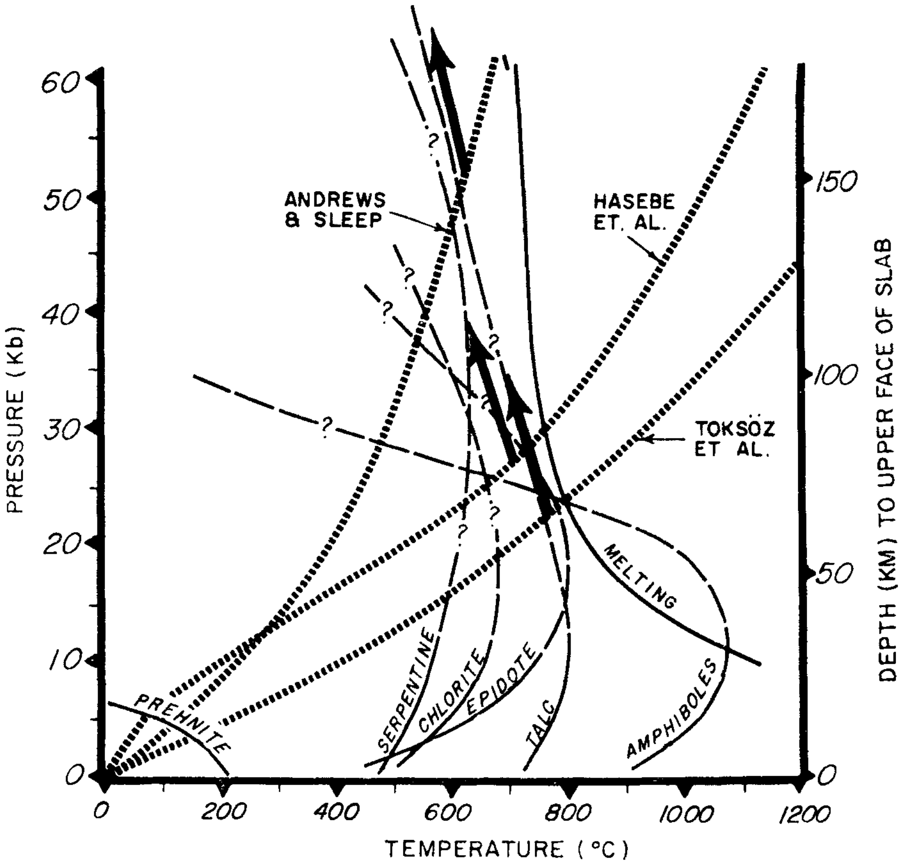


FIG. 2. Thermal structure of the upper face of the underthrusting lithosphere predicted by geophysical models of Toksöz *et al.* (1971), Hasebe *et al.* (1970) and Andrews & Sleep (1974). Also on the depth-temperature plot are dehydration temperatures of various crustal hydrous minerals discussed in the text. The dehydration range is from approximately 200 °C to approximately 900 °C. The melting range is from approximately 650 °C (for hydrous assemblage, Lambert & Wyllie 1972) to approximately 1200 °C (for anhydrous assemblage). Dashed portions of the univariant curves for dehydration are still speculative at this stage (Lambert & Wyllie 1972; Ringwood 1974). Arrows indicate effect of dehydration on geotherms from the various models.

mantle above the slab. McKenzie & Sclater (1968) and McKenzie (1969) estimated that the frictional heating term was small and unimportant, and noted that the slab would remain cold to great depths because conductive heat transfer is such a slow process. McKenzie (1969), Griggs (1972), and more recently, Andrews & Sleep (1974) have calculated the thermal structure of the slab with no frictional heating term at all. All these models generate high heat flow and volcanism in island arcs and marginal seas by the secondary convective flow induced by the downgoing slab in the asthenosphere. In the Andrews & Sleep (1974) model, water in the oceanic crust of subducting slabs will remain in the slab to depths of > 200 km or until pressure dehydration takes place, because the temperature is kept so low (Fig. 2).

In order to generate sufficient temperatures to produce volcanism on the Benioff–Wadati zone at the appropriate points (Fig. 2), several authors introduced frictional heating as a significant term in the thermal balance calculations. Oxburgh & Turcotte

(1970), Minear & Toksöz (1970), Toksöz *et al.* (1971), Turcotte & Schubert (1973) calculated that shear stresses of 1 to 2 Kbars were necessary to heat the slab sufficiently to produce melting at 100–200 km. Stresses of this magnitude are much larger than stress drops estimated from earthquakes occurring in the downgoing slab (of the order of a few hundred bars), but may not be unreasonable when seismic efficiency is taken into consideration (Wyss 1970). Because of such large frictional heat generation, these models fail to reproduce the low heat flow between the trench and the magmatic front.

Hasebe *et al.* 1970, recognizing this problem, constructed a thermal model constrained to fit the heat flow data around Japan. Besides the need for frictional heat generation and bodily transfer of heat beneath marginal seas (to account for the high heat flow there), they were forced to constrain the frictional heating to begin only at depths greater than 60 km in order to fit the low heat flow data in the zone between the trench and the volcanic front (Fig. 1). They suggested that dehydration in the oceanic crust may absorb enough heat to ‘cancel out whatever heating may occur there’ (Hasebe *et al.* 1970, p. 349). They, however, offered no quantitative explanation for this anomalous 0–60 km region. It is the physics and chemistry of the downgoing slab between the trench and arc which we wish to investigate below.

Water in the oceanic crust

In order to propose dehydration as a major source of water for magma production and of the low heat flow in the zone between trench and volcanic front, it is first necessary to establish that the oceanic crust is indeed hydrated. Widely diversified geological and geophysical evidence has accumulated in the past couple of years to make the existence of a hydrated oceanic crust almost a certainty.

Metamorphosed basaltic and gabbroic rocks are commonly dredged from the walls of the median valley and transverse fracture zones and sometimes even on the tops of crestal mountains of the Mid-Atlantic and Mid-Indian Ocean Ridges (see below). A similar rock suite has been recovered from the south-eastern Pacific (Nishimori & Anderson 1973). It appears that, except for a thin surface layer, the basaltic-gabbroic crust of the ocean floor is metamorphosed. The grade of metamorphism is diverse, including zeolite, greenschist and amphibolite facies (Melson, Thompson & van Andel 1968; Miyashiro *et al.* 1971). Prehnite was found in some rocks from the equatorial Atlantic, though it is not clear yet how widespread rocks of the prehnite-pumpellyite facies are (Bonatti, Honnorez & Ferrara 1971).

Recent heat flow results from Lister (1972), Williams *et al.* (1974), Sclater *et al.* (1974), Detrick *et al.* (1974), and Anderson & Hobart (1976), argue strongly for widespread circulation of sea water deep into the oceanic crust at spreading centres. Williams *et al.* (1974) observed an oscillatory heat flow pattern on the Galapagos Spreading Centre which is best modelled by massive penetration of hydrothermal fluids $>$ 3 km into the oceanic crust. Anderson (1972) argued that such circulation would produce extensive metamorphism.

Seismic refraction evidence points to velocities in the oceanic crust which are best fitted by velocities of a hydrated basaltic layer 2 and a hydrated gabbroic upper layer 3 (Schreiber & Fox 1973) or a partially serpentinized layer 3 (Woollard 1975). The seismic velocity below the MOHO is that of dry peridotite.

JOIDES leg 37, site 334 drilled $\sim$ 400 m into the oceanic crust on a fault scarp near the crest of the Mid-Atlantic Ridge. In addition to tholeiite, they also sampled fresh gabbro, serpentinized olivine gabbro, serpentinized plagioclase peridotite, and tectonic breccia (Melson *et al.* 1974).

Dehydration is endothermic, and the magnitude of heat absorbed is approximately proportional to the weight per cent H_2O in the reaction assemblage. It is, therefore, important to establish the water content of the hydrous oceanic crust. The most

Table 1

Chemical composition of rocks representative of layers of oceanic crust (weight %)

	2a	2b	2c	2d	3a	3b
SiO_2	50·0	48·6	47·48	49·37	47·73	49·22
TiO_2	1·4	1·4	1·47	0·68	0·20	0·33
Al_2O_3	16·1	15·8	15·79	15·09	20·09	17·01
Fe_2O_3	1·4	5·2	5·73	1·74	1·00	1·28
FeO	8·0	4·2	4·21	8·82	4·77	4·34
MnO	0·2	0·2	0·17	0·20	0·10	0·14
MgO	8·0	7·8	7·37	9·34	9·59	10·53
CaO	11·0	10·8	10·97	8·89	11·51	12·73
Na_2O	2·8	2·7	3·08	2·23	2·23	2·20
K_2O	0·2	0·2	0·17	0·04	0·03	0·06
H_2O^-	0·2	1·5	1·32	0·16	0·05	0·23
H_2O^+	0·5	1·5	2·31	3·48	2·98	2·06
P_2O_5	0·1	0·1	0·13	0·06	0·01	0·02
Total	99·9	100·0	100·20	100·10	100·29	100·15

2a. Fresh abyssal tholeiite, representative of fresh rocks in the upper part of layer 2.

2b. Weathered abyssal tholeiite.

2c. Metamorphosed abyssal tholeiite (V25–RD6–T30) in the zeolite facies with $FeO^*/MgO = 1·27$ from the Mid-Atlantic Ridge (23°44·7′N, 45°33·6′W).

2d. Metamorphosed abyssal tholeiite (V25–RD5–T28) in the greenschist-like facies with $FeO^*/MgO = 1·11$ from the Mid-Atlantic Ridge (23°31·7′N, 45°07′W).

3a. Metagabbro (V25–RD5–T48) with $FeO^*/MgO = 0·59$ from the Mid-Atlantic Ridge (23°31·7′N, 45°07′W).

3b. Olivine gabbro, an average of nine analyses from the Mid-Atlantic Ridge (V25–RD5 and 6). From Miyashiro *et al.* (1970), Table 2.

reliable pieces of evidence in this regard are probably the chemical analyses of metamorphic rocks dredged from mid-ocean ridges. Miyashiro, Shido & Ewing (1970) reported average metagabbroic compositions from the Mid-Atlantic Ridge with ~2 per cent H_2O. Chernysheva (1970) reported altered diabases and gabbros from the Indian Ocean ridges with ~3 per cent H_2O. Thompson & Melson (1972) reported metabasalts from the Mid-Atlantic Ridge averaging 4 per cent H_2O. Engel & Fisher (1969) presented metagabbroic rock analyses from the central Indian Ridge with ~3 per cent H_2O. Miyashiro, Shido & Ewing (1969) reported Mid-Atlantic Ridge serpentinites with >12 per cent H_2O. Melson & Thompson (1971) reported serpentinite with almost 13 per cent H_2O.

Lambert & Wyllie (1972) have experimentally investigated melting of gabbros to 35 kb with excess water present. They find that the melting curve in *P–T* space is depressed from the anhydrous condition of melting at 1200 °C to a hydrous melting temperature of 600–700 °C (Fig. 2). Dehydration of prehnite, chlorite, epidote, and serpentine will occur before melting, but amphiboles may melt before they dehydrate. (The univariant curve for hornblende is basically isobaric at 25–30 kb; this is a complication which will be elaborated upon below.)

A model of the oceanic crust

Miyashiro *et al* (1971) made a comprehensive study of metamorphic rocks from the Mid-Atlantic Ridge near 24° and 30 °N. They recovered metabasalts in the zeolite and greenschist (or greenschist-like) facies as well as metagabbros in the greenschist (or greenschist-like) and amphibolite facies (Table 1). Thus, the temperature range of metamorphism tends to be higher in metagabbros than in metabasalts. This suggests that metagabbros were metamorphosed generally at greater depths than metabasalts,

Table 2(a)

Theoretical assemblages for dehydration of layer 2B of the oceanic crust.

Theoretical assemblage number	% H_2O (weight)	Total % hydrous minerals	% Individual minerals	Total ΔH_R (cal g^{-1})
2–1	1·7%	33%	46% Albite 20% Epidote 9% Chlorite 4% Actinolite 21% Quartz*	17·7
2–2	2·5%	54%	34% Albite 30% Epidote 13% Chlorite 10% Actinolite 13% Quartz*	27·7
†2–3	3·5%	73%	24% Albite 33% Epidote 20% Chlorite 20% Actinolite 3% Quartz*	35·1
2–4 (85% model 2–3+ 15% serpentine mix)	4·9%	77%	20% Albite 28% Epidote 17% Chlorite 17% Actinolite 3% Quartz* 15% Serpentine	40·0

* And other anhydrous minerals to preserve isochemicallity with model 2–3.

† Best estimate.

Table 2(b)

Theoretical assemblages for dehydration of layer 3A of the oceanic crust.

Theoretical Assemblage number	% H_2O (weight)	Total % hydrous minerals	% Individual minerals	Total ΔH_R (cal g^{-1})
3–1	1·6%	19%	76% Anorthite 8% Actinolite 11% Chlorite 5% Enstatite*	9·9
†3–2	2·5%	32%	63% Anorthite 16% Actinolite 16% Chlorite 5% Enstatite*	15·7
3–3	4·3%	60%	30% Anorthite 32% Actinolite 28% Chlorite 10% Enstatite*	28·1
3–4 (mechanical mixing model of 85% model 3–3+15% serpentine)	5·8%	66%	26% Anorthite 27% Actinolite 24% Chlorite 8% Enstatite* 15% Serpentine	34·0

* And other anhydrous minerals to preserve isochemicallity with model 3–2.

† Best estimate.

and that the temperature during metamorphism tended to be higher with increasing depth. Of course, such a horizontal layering cannot be complete since magma intrusions and vertical movements of H_2O must be occurring beneath mid-oceanic ridges which would result in a considerable irregularity in the distribution of rock types and temperatures.

The correlation of metamorphic facies with temperature is considerably uncertain because of (1) the low accuracy of experimentally determined equilibrium curves for metamorphic reactions, particularly at low temperatures, and (2) our uncertain knowledge of the behaviour of H_2O beneath mid-oceanic ridges. Based on some selected equilibrium curves (Table 4) and oxygen isotope geothermometry, we may assign a temperature range including 400 °C to the greenschist facies, and a range including 600–700 °C to the amphibolite facies (Miyashiro 1973, Fig. 3–12).

Metamorphic rocks in mid-oceanic ridges may be classified into two groups. Group I includes rocks whose chemical composition remained virtually unchanged during metamorphism except for the addition of H_2O, and Group II includes rocks whose chemical composition suffered a marked change during metamorphism. The commonly-observed compositional changes are the addition of Na_2O to zeolite-facies metamorphic rocks, and the removal of CaO and either the removal or the addition of SiO_2 in greenschist-facies metamorphic rocks. Such intense chemical migration is possible only by the dissolution of such components in a moving aqueous fluid (Miyashiro *et al.* 1971). Group II rocks belong to the zeolite and greenschist facies and Group I rocks to the greenschist and amphibolite facies. Thus we may propose that H_2O as an aqueous fluid could move (circulate) through the upper part of the mid-oceanic ridge crust and may not have actively moved in the lower part.

Table 3(a)

Theoretical reactant and product assemblages upon dehydration of layer 2B, model 2–3 (Table 2(a)) during subduction of the oceanic crust

Reactant assemblage	Product assemblage*
Albite	Anorthite
Epidote	Albite
Chlorite	Diopside
Actinolite	Hedenbergite
Quartz	Enstatite
	Quartz
	Sillimanite
	Fayalite
	H_2O

Table 3(b)

Theoretical reactant and product assemblages upon dehydration of layer 3A, model 3–2 (Table 2(b))

Anorthite	Anorthite
Actinolite	Enstatite
Chlorite	Hedenbergite
Enstatite	Diopside
	Fayalite
	Quartz
	Sillimanite
	H_2O

* Sillimanite from reactions in Table 4 will react with diopside to form enstatite and anorthite. In addition, the product assemblage is a low pressure, high temperature assemblage. In the slab, anorthite will react with diopside to form garnet and albite will form some jadeite. Heats of reaction of these solid–solid reactions are small relative to dehydration reactions and are neglected in this paper (Robie & Waldbaum 1968).

As indicated by deep drilling at the Mid-Atlantic Ridge (Melson *et al.* 1974), the upper part of the oceanic crust is made up of an accumulation of basaltic volcanic materials (lava flows and pyroclastics). Such a layer should have originally contained a large amount of sea water in the interspaces between rock masses and fragments. The amount of interstitial water would commonly be as much as 30 per cent by volume and 10 per cent by weight.

On the other hand, the lower part of oceanic crust may be mainly composed of compact gabbroic rocks. The original water content of such a layer would be small, and diffusion of H_2O also may be limited. Movement of aqueous fluid is probably very limited in the layer, so metamorphic reactions involving addition of H_2O may be greatly hindered by the limited availability of H_2O.

Fox, Schreiber & Peterson (1973) presented many measurements of compressional wave velocities of rocks from the ocean floor, as well as a concise summary of seismic observational data bearing on the layered structure of the oceanic crust. It may be assumed that such a layered structure was created at mid-oceanic ridges when the oceanic plate was created. The temperature of the oceanic crust should be generally high beneath the axial zone of mid-oceanic ridges, and the major features of the metamorphic layer structure of the oceanic crust are probably formed in response to the temperature distribution at that time. During the lateral movement of oceanic crust away from the axial zone, the temperature should gradually decrease, and some retrogressive metamorphism should occur. Such retrogressive metamorphism is commonly observed in some metagabbros (Miyashiro *et al.* 1971). However, the extent of such

Table 4

Estimated thermodynamic parameters of dehydration reactions in layers 2 *and* 3 *of the oceanic crust* (*from Helgeson, Nesbitt & Delany* 1976)

Reaction	Pequil† (kb)	Tequil (°C)	ΔH_R (cal g^{-1})	ΔV† (cm^3 mol^{-1})
(1) Prehnite + chlorite + 2 qtz $\rightleftharpoons$ 4 epidote + tremolite + 6 H_2O	3	160	6·8	−31·3
(2) 6 Epidote + chlorite + 7 qtz $\rightleftharpoons$ tremolite + 10 anorthite + 6 H_2O	10	734	40·3	206
(3) Fe epidote $\rightleftharpoons$ Haematite + 2 anorthite + 2 wollastonite + H_2O	4*	994	81·1	66·5
(4) 2 Mg chlorite + qtz $\rightleftharpoons$ 2 sillimanite + 5 forsterite + 8 H_2O	10	900	99·7	47·9
(5) 2 Fe chlorite + qtz $\rightleftharpoons$ 2 sillimanite + 5 fayalite + 8 H_2O	10	408	40·1	10·6
(6) Serpentine $\rightleftharpoons$ enstatite + forsterite + 2 H_2O	10	518	67·5	2·7
(7) 5 Serpentine $\rightleftharpoons$ talc + forsterite + 9 H_2O	10	501	59·7	17·3
(8) Serpentine + brucite $\rightleftharpoons$ 2 forsterite + 3 H_2O	9	414	74·1	7·2
(9) Talc $\rightleftharpoons$ 3 enstatite + qtz + H_2O	10	860	28·2	2·6
(10) Tremolite + forsterite $\rightleftharpoons$ 2 diopside + 5 enstatite + H_2O	10	860	10·6	−6·6
(11) Tremolite $\rightleftharpoons$ 2 diopside + 3 enstatite = qtz + H_2O	10	991	19·6	−0·4

* Highest pressure at which accurate determination of thermodynamic properties is possible at this time.

† Preliminary values for the thermodynamic calculations were used which are in the process of being revised. In most cases, uncertainties are of the order of 100 cal mole^{-1} or less (Helgeson, private communication).

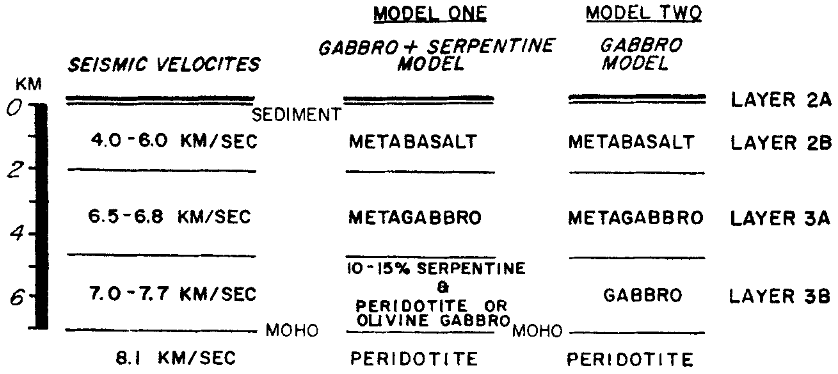


FIG. 3. Seismic velocity structure of the oceanic crust from seismic refraction results of Schreiber & Fox (1973). Model 1 is geochemical model of the oceanic crust which predicts an assemblage with seismic velocities matching observations (Woollard 1975). Model 2 is an alternative geochemical model proposed by Schreiber & Fox (1973) which fits the seismic velocity structure equally well.

changes is probably relatively small. Outside mid-oceanic ridges, the temperature at the base of oceanic crust is probably very low, (say <150 °C).

As observed by Christensen & Salisbury (1973), weathering of a near-surface layer of seismic layer 2 should advance progressively with time during the lateral movement of oceanic crust. However, the thickness of such a weathered layer is probably very small, and negligible in our discussion. So, we may consider that the igneous and metamorphic layer structure of the oceanic crust created in and near the axial zone of mid-oceanic ridges is preserved in all essential features until the crust reaches a subduction zone.

On the basis of the above discussions, we present one possible mineral and chemical model with composition typical of rocks in layers of the oceanic crust. The upper part of layer 2 (layer 2A) is mainly composed of only slightly altered abyssal tholeiites. A representative composition is given in Table 1, column 2a. A possible composition of weathered rocks in the upper part of layer 2 is obtained by increasing Fe_2O_3 H_2O^- and H_2O^+, as shown in Table 1, column 2b.

Two analyses of basalts metamorphosed at low temperatures are shown in Table 1, columns 2c and 2d, to represent rocks from layer 2B. Though quantitative estimation of mineral compositions of such rocks is very difficult, a possible mineral composition (mode) of such rocks was calculated from the norm (Table 2(a), model 2–3). An average composition of olivine gabbro from the Mid-Atlantic Ridge is given in Table 1, column 3b, in order to give an idea of relatively unaltered gabbros possibly from layer 3B. Some gabbros used in the calculation of this average were practically free from metamorphic recrystallization, and others were only slightly metamorphosed. Table 1, column 3a gives an analysis of a more completely-metamorphosed gabbro in a greenschist-like facies (with calcic plagioclase) from the Mid-Atlantic Ridge. An estimated mineral composition of such a gabbro is given in Table 2(b), model 3–2.

Serpentinite occurs usually along transverse fracture zones across mid-oceanic ridges. However, it is also possible that some serpentinite-peridotite masses are intruded into ordinary parts of the oceanic crust.

In mid-oceanic ridge serpentinites, brucite is scarce and may be neglected in our discussion, but talc and pargasite are more common. Quantitative estimates of the average amounts of these two minerals is not possible. Here, we may assume that as part of layer 2 and the upper part of layer 3, serpentinites are composed of 99 per cent serpentine minerals, 0 per cent talc and 1 per cent pargasite by weight (besides magnetite).

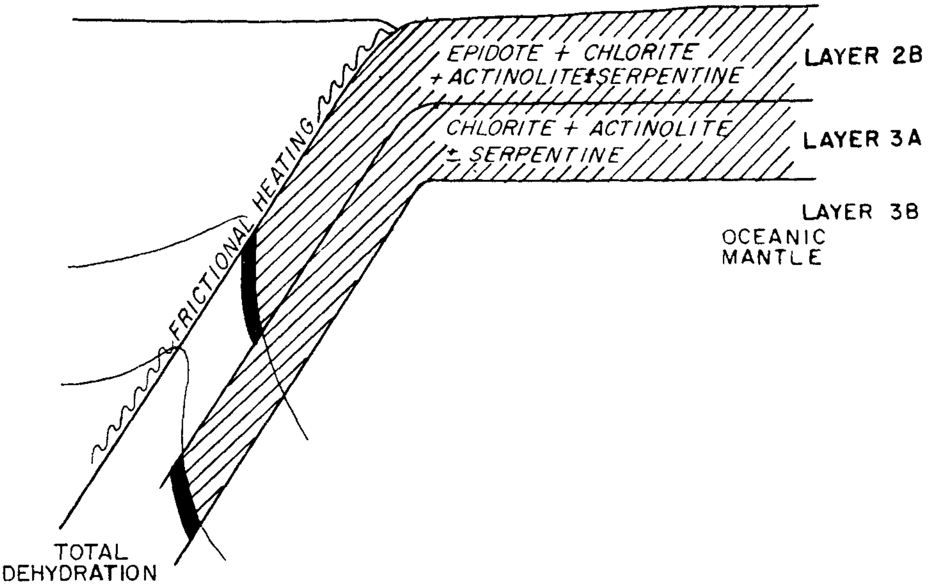


Fig. 4. Hydrous artificial assemblages involved in step-by-step dehydration of Model 1. Solid black lines are theoretical reaction boundaries.

In the lower part of layer 3, serpentine minerals might be decomposed to produce talc and olivine, and so the mineral composition of the rock might become 70 per cent olivine (forsterite) + 29 per cent talc + 1 per cent pargasite. Toward the base of layer 3, the talc should be decomposed by the reaction number 9 of Table 4.

We thus can construct the following models for the hydrated layers of the oceanic crust which are consistent with the seismic structure of a four-layered crust from Schreiber & Fox (1973) (Fig. 3). We assume that the major portion of layer 2, layer 2B, is composed of metabasalt. Layer 3 consists of a metamorphosed gabbroic upper layer, and a lower layer of either olivine gabbro or a mix of 10–15 per cent serpentinite and peridotite or olivine gabbro. In addition, a mechanical mix may occur in which up to 15 per cent of either layer 2B or 3A is serpentine. Most of this would reside in fractures zones. We then construct theoretical assemblages which we propose to be representative of the two hydrous layers of the oceanic crust (Table 2). Because we are not certain of the exact water content of the oceanic crust, we have constructed theoretical assemblages which are isochemical with the exception of mechanical mixing of serpentine and which contain varying water contents of from 2 to 5 per cent H_2O (by weight) in layer 2B and from 1 to 6 per cent in layer 3A. In the theoretical assemblages for layer 2B, prehnite, chlorite, epidote. actinolite ± serpentine, constitute the hydrous mineral assemblage (Table 2(a), Fig. 4) whereas in the theoretical assemblages for layer 3A, chlorite and actinolite ± serpentine constitute the hydrous components of the crust. We have evaluated prehnite reactions such that they add to the total content of epidote and tremolite but contribute little themselves to the thermal regime of the oceanic crust (Table 4).

We would like to use actinolite and hornblende as the amphibole but the tremolite end-member is the only amphibole for which thermodynamic data exists at this time. Our set of minerals in the theoretical assemblages presented below is close enough to reality to represent the basic physical principles involved in dehydration of the oceanic crust during the subduction process. Thus models 2–3, Table 2(a), and 3–2, Table 2(b),

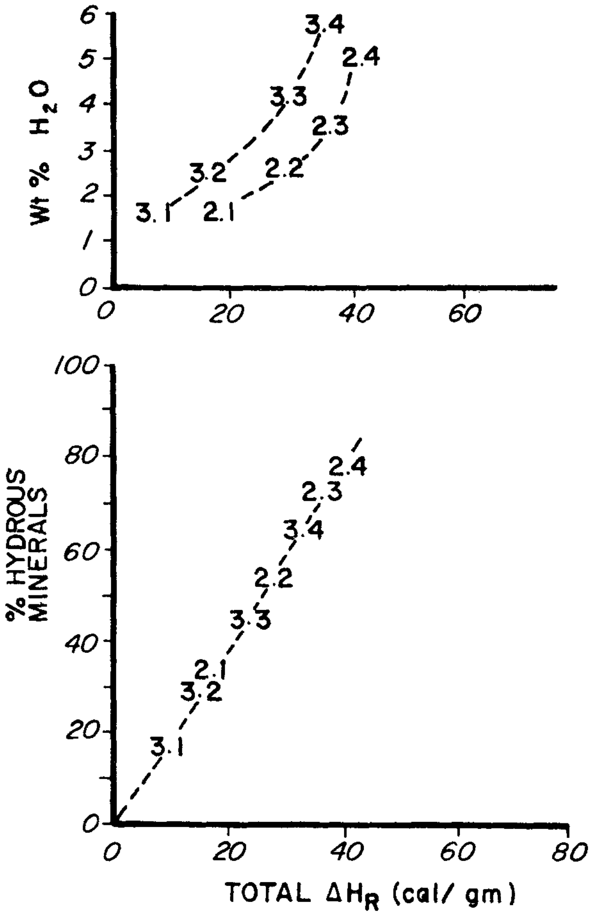


Fig. 5. Variation in total enthalpy of reaction as a function of % H_2O and % hydrous minerals in theoretical assemblages of Table 2. The percentages are for that layer only, so the percentage for the total oceanic crust would be much smaller. Numbers are various models of Table 2.

were constructed to best represent the analyses presented in Table 1 for layers 2B and 3A, respectively. The water content was then incremented and the isochemical changes in percentages of hydrous minerals at various steps noted, these becoming theoretical assemblage 2–1 through 2–4 and 3–1 through 3–4. The effect of increasing the water content of layer 2B from 2 to 5 per cent H_2O is to increase the total percentage of hydrous minerals from 19 to 66 per cent.

Heats of reaction for dehydration of the oceanic crust during underthrusting

In order to evaluate the heat budgets of these theoretical assemblages, thermodynamic parameters of 11 reactions were calculated at elevated pressures and temperatures. The enthalpy, entropy, free energy, heat capacity, and equilibrium constant of each reaction were evaluated as pressure and temperature were incremented using the Maier–Kelley heat capacity power function and/or average heat capacities for the reactants and products. The standard state used was that of unit activity of solids and H_2O at all pressures and temperatures (Helgeson & Kirkham 1974). The method of calculation of the thermodynamic parameters of these reactions, standard state input parameters for each reaction considered, and detailed results at varying pressures and temperatures of equilibrium are given in Helgeson, Nesbitt & Delany (1975, 1976). Results at 10 kb and equilibrium temperature are given in Table 4.

The dehydration of each layer was then evaluated in total to facilitate the calculation rather than to represent the actual dehydration process in the downgoing slab. Actual metamorphic reactions are continuous and *cannot* be divided into discrete layers, partly because of metastability but, more importantly, because of the formation of solid solutions of constituent minerals. The different end members of a solid solution mineral are dehydrated at different temperatures, and the actual dehydration of the mineral occurs over a wide range of temperatures between the lowest and highest temperatures of end-member dehydration. Thus, dehydration-temperature ranges of various minerals overlap one another, resulting in continuous and gradual changes.

Based upon the simplistic concept of the oceanic crust incorporated in the theoretical assemblages of layers 2B and 3A, we can conclude that the amount of heat absorbed by the dehydration of the oceanic crust is probably very near the ΔH_R of one of the theoretical assemblages presented in Table 2. The total ΔH_R of dehydration of the oceanic crust is not a simple linear function of the percentage of H_2O in the oceanic crust but appears to be more linearly related to the total percentage of hydrous minerals in the oceanic crust (Fig. 5). It is interesting to note that even though we have chosen two quite different chemical models for the two layers involved, the total quantity of heat absorbed by dehydration appears to have almost a one-to-one correspondence to the percentage of hydrous minerals in the ocean crust for both models.

Based upon our presentation above, we believe the most likely candidate for the proper model for the oceanic crust is the combination of assemblages 2-3 and 3-2. The dehydration of layer 2B would then require the absorption of 35 calories per gram of layer 2B dehydrated, whereas that of layer 3A would require the absorption of 16 cal g^{-1}. However, if future deep drilling or dredging, for example, show that the composition of the oceanic crust is different from that chosen above, the data in this paper should allow the calculation of the heat budget involved in dehydration simply by the construction of a new theoretical assemblage, if another from Table 2 does not fit.

We can identify four ' errors ' in the above analyses which lead us to believe that our heat-sink estimates are minimal values for that likely to occur upon underthrusting of oceanic crust. First, the ΔH_R at 30 kb may be at least as great as that at 10 kb based upon the thermodynamic behaviour of H_2O at high pressure (Tödheide 1972). Though thermodynamic calculations are constrained to 10 kb at this time because of uncertainties in the determinations of the thermodynamic properties of water at higher pressures, the 10 kb values in Table 4 suffice to demonstrate the physical theory involved in this paper. Second, melting could begin at 600–700 °C in the hydrated oceanic crust (Lambert & Wyllie 1972). Thus, some amphiboles could melt as soon as dehydration occurs. Though our collection of silicate latent heats of melting are more primitive than our data set of enthalpies of reaction, we can make a crude estimate of the magnitude of this heat absorption by considering the heats of fusion data of Goranson (1942). Melting of anorthite absorbs 105 cal g^{-1}; diopside, 106 cal g^{-1}; and albite, 48 cal g^{-1}. No numbers exist to date for melting of hydrous minerals. Thus, if melting occurs in the slab during or after dehydration of the oceanic crust, even more heat will be absorbed than that calculated for the dehydration of the oceanic crust. Also if a significant volume of sediment is subducted, the ΔH_R from dehydration of micas and clays will absorb significant quantities of additional heat.

Another possible source of ' error ' is that the newly-released water may escape from the slab, carrying heat with it. Uyeda & Horai (1964) have calculated the quantity of heat removed by advection of heat by water transport. Assuming the same heat removal as that calculated by Uyeda & Horai (1964) (for water advection through geosynclinal sediments approximately 50 km thick; the oceanic crust and the mantle wedge are less permeable, so heat removal should be smaller in our case than in that of Uyeda & Horai 1964) we arrive at $\Delta H < 5$ cal g^{-1} of water advecting from the slab. So we can conclude that the amount of heat liberated by water escaping from the downgoing slab, though small compared to the quantity absorbed during dehydration, again makes our estimate of the heat absorbed during underthrusting minimal.

Now, realizing the nature of the minimal estimate from our ΔH_R calculation, we can evaluate the quantity of heat absorbed during dehydration relative to the quantity of heat generated by any frictional heating which may be occurring along the slab interface.

Net effects of dehydration heat absorption and frictional heat generation on the downgoing slab

We now wish to calculate the rate of ' heat generation ' by the endothermic dehydration reactions in the oceanic crust of the downgoing slab. This term will be negative since we are talking of a heat sink.

A dipping slab converging at a velocity of 8 cm/yr reaches a depth of 75–100 km in approximately 2 My. It is probable that dehydration will have been completed by these depths (Fig. 2). We will treat dehydration as occurring in two layers, as above. The time for dehydration of each layer is critical to the heat budget calculation because the heat generation term A is given as follows:

$$A = \frac{\Delta H_R \cdot \rho}{t} \qquad \text{where}$$

ρ = density and t = time for dehydration to be completed in this case. Dehydration of our theoretical assemblages will not begin until temperatures > 400 °C are reached along the downgoing slab interface (Fig. 2). Regardless whether this heating is by conduction or from friction along the surface, it is probable that dehydration will occur during the last My of the 2 My time interval required for the slab to reach a depth of 100 km (Fig. 2). If $t = 0{\cdot}5$ My for dehydration of layer 2B, A becomes $-6{\cdot}7 \times 10^{-12}$ cal cm^{-3} s^{-1}; if $t = 0{\cdot}5$ My for dehydration of layer 3A, A is -3×10^{-12} cal cm^{-3} s^{-1}.

What effect on heat flow at the slab interface will this heat absorption from dehydration have compared to possible heat generation from friction? The problem may be approximated as heat flow at the plane of contact between two slabs with internal heat generation.

Q (heat flow) at the slab interface may be approximated as follows:

$$Q = Q_f + A_2\, l_2 + A_3\, l_3$$

where Q_f is the heat flow produced by frictional heating; and l_2 is the dehydration layer 2B thickness, and A_2 is heat generation from dehydration of layer 2B; l_3 is the layer 3A thickness, and A_3 is the heat generation from dehydration of layer 3A. The thicknesses of dehydration layers 2B and 3A are taken to be 2 and 3 km, respectively (Fig. 3).

Q_d (heat flow ' absorbed ' by dehydration) then is $-2{\cdot}3 \times 10^{-6}$ cal cm^{-2} s^{-1}.

How does this heat flow compare with Q_f? The quantity of energy available for frictional heating can be estimated as follows:

$$Q_f \sim \Delta\rho g v l$$

where $\Delta\rho$ is the density contrast between the lithosphere and asthenosphere. $\Delta\rho \approx 0{\cdot}03$ g cm^{-3} if the asthenosphere is composed of < 6 per cent partial melt (Birch 1969). This quantity is probably minimal and *uncertain by at least a factor of* 2 since $\Delta\rho$ would be larger because of the temperature difference between the slab and the asthenosphere. l is the thickness of the lithosphere (taken to be 100 km here). v is the underthrusting velocity (taken to be 8 cm/yr here). g is the gravitational acceleration.

Q_f must then be $\sim 2{\cdot}0 \times 10^{-6}$ cal cm^{-2} s^{-1}.

If this estimate were correct, dehydration would absorb 100 per cent of the heat generated by friction.

Alternatively, we might calculate Q_f by assuming that all the energy required to elevate the temperature along the slab interface to the melting temperature (T_m) comes from shear stress heating. From Turcotte & Schubert (1973), the heat flow per unit area entering the plate at the slip plane from frictional heating along that plane is given by:

$Q_f = \sigma\bar{v}$ where σ = shear stress along the slip zone

$\bar{v}$ = velocity of convergence between the two plates involved.

In order to evaluate the importance of this term relative to the dehydration term, we must select a stress appropriate for the slab interface. Values proposed in the literature range from 0 by Andrews & Sleep (1974) to $> 2{\cdot}5$ Kbar for Toksöz *et al.* 1971. This term is constrained only by the *assumption* that this must be the source of heat for melting below island arcs.

From Turcotte & Schubert (1973)

$$\sigma = \frac{K\,T_m}{2}\left(\frac{\pi\ \sin\theta/\cos\phi}{\bar{v}\,d_m\,k}\right)^{\frac{1}{2}}$$

where

K = thermal conductivity = $0{\cdot}007$ cal cm^{-1} s^{-1} °C^{-1},

T_m = temperature of melt of slab = 700 °C,

θ = subduction angle = 30°,

ϕ = component of oblique subduction = 0°,

$\bar{v}$ = velocity of convergence = 8 cm/yr,

d_m = depth of magma generation = 100 km,

k = thermal diffusivity = $0{\cdot}01$ cm^2 s^{-1}.

The above parameters approximate the convergence geometry of Japan Trench. T_m is taken from Fig. 2 and Lambert & Wyllie (1972).

σ then is $19{\cdot}2$ cal cm^{-3} or 800 bars.

Frictional heating would generate $Q_f = 4{\cdot}9 \times 10^{-6}$ cal cm^{-2} s^{-1}. Then, in order to produce melting, additional Q_d would have to be generated by friction. This requires an additional 400 bars of shear stress along the slip zone. By this mechanism, 1/3 of all heat generated along the slip zone goes into dehydration of the oceanic crust rather than into heating the slab.

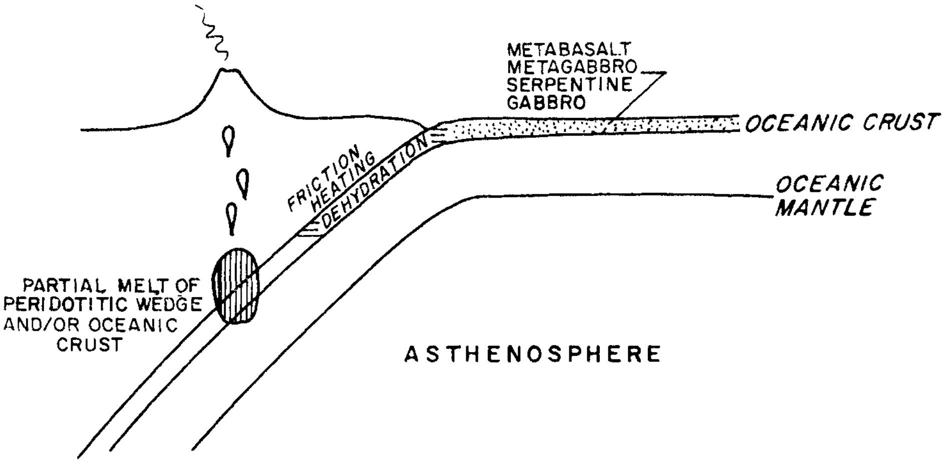


FIG. 6. Schematic of petrological regime in and around downgoing slab. Volcanic arc magmas are generated in the periodotitic wedge overlying the slab and/or by partial melting of the subducted oceanic crust.

Thus, it is clear that Q_d is a quite important heat sink which should be included in detailed quantitative calculations of the thermal regime of the downgoing slab. We are currently in the process of making these calculations. It is obvious to us though that heat absorbed by dehydration is a prime mechanism for generation of the low heat flow zone measured in the arc-trench gap.

The effect of dehydration on the thermal regime of the downgoing slab assuming frictional heating is important will be to suppress the heating of the oceanic crust as shown by the arrows in Fig. 2. All theoretical models published to date therefore overestimate the rate of heating of the slab. Frictional heating thus does not control melting below island arcs, dehydration does.

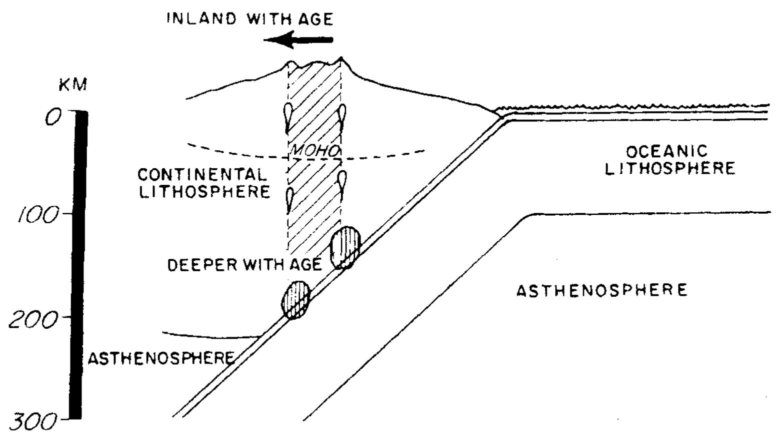
If on the other hand no frictional heating occurs upon subduction, the problem becomes whether there is enough energy available for dehydration to occur. As we have shown, more than 2×10^{-6} cal cm^{-2} s^{-1} of heat flow, in addition to that needed simply to heat the slab, are needed for the dehydration reactions to occur. It is not obvious that that much energy is flowing into the oceanic crust in thermal models with no frictional heating term. Otherwise the hydrous assemblage would be subducted deep into the asthenosphere, and water would not be available for generation of volcanic arc primary magmas. Much more quantitative work is needed before we have a clear idea of the thermal structure of the downgoing slab.

Petrological implications of dehydration

We have so far demonstrated that large quantities of heat are absorbed upon dehydration of the oceanic crust during subduction. We now consider the petrological effects of this newly-released water. This and other related problems have been extensively discussed by many authors (Yoder 1969; Kushiro 1972; Ringwood 1974; among many others).

Before any given section of oceanic crust begins subduction, it has lost most of the heat it had at the time of plate creation and metamorphism. (We will assume it is old ocean crust.) As underthrusting proceeds, frictional heating along the plate boundary or simple conduction of heat from the hot asthenosphere to the cold lithosphere raises the temperature along the upper surface of the slab. Dehydration is probably near completion by the time the slab has reached approximately 75–100 km depth, where dehydration of amphibole will be completed because of its near isobaric equilibrium curve at high P (assuming enough energy is available). Serpentine and low P–T prehnite, epidote, and chlorite will also have dehydrated by this depth (Fig. 2). At this point, free water will occupy the interstices of the former oceanic crust. Whether the water then diffuses upward or is confined will determine when and where melting will occur when the temperature reaches approximately 650 °C (Lambert & Wyllie 1972). If confined, the melting depth of the slab will be approximately 100–150 km (Fig. 2) below the surface. The crustal melt will form andesitic primary magma which will migrate upwards. When it reaches the peridotitic upper mantle wedge which is much hotter than 650 °C, the crustal melt will provide a hydrous environment which will significantly lower the melting point of the wedge. Heat will also flow from the wedge to the crustal melt. If the wedge has a higher temperature than approximately

(a) Continent–ocean convergence



(b) Ocean–ocean convergence

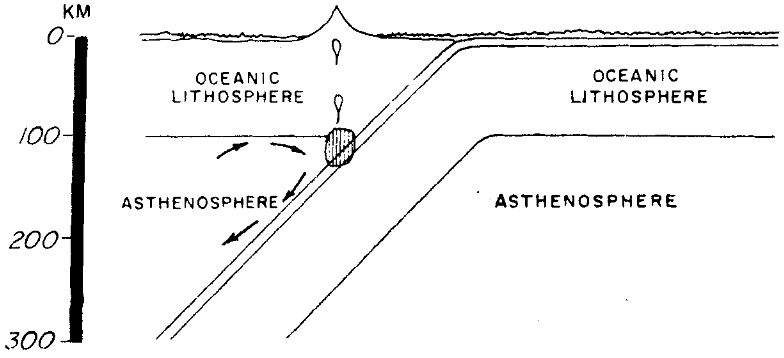


FIG. 7(a). Schematic of proposed magma source temporal migration during continent–ocean collision. Migration deeper into the continental lithosphere is required to tap new heat sources. (b) Schematic representing resupply of heat to magma source during ocean–ocean collision from asthenospheric convection. No large-scale migration is necessary because heat is continually resupplied from below.

1000 °C, melting will occur there also. The resulting wedge melt will either be tholeiitic primary magma (Nicholls & Ringwood 1972) or tholeiitic or calc-alkaline andesitic primary magma (Kushiro 1974). The resulting hybrid melt of both ocean crust and wedge components will then rise to the surface, undergo differentiation and deliver derivative magmas to the surface (Fig. 6).

If the water immediately escapes from the oceanic crust upon dehydration, no melting of the oceanic slab will occur until great depths when the anhydrous melting temperature of approximately 1200 °C is reached. The water will migrate upward into the hot wedge, and melting will occur forming either basaltic or andesitic primary magma. Differentiation on the way to the surface and extrusion of derivative magmas will follow. No oceanic crustal component will exist in these magmas except for water, however.

It is also interesting to note that all models require significant heat removal to the surface by advection of magma. Thus either the source region gets colder with time or this heat must be replaced as quickly as it is convectively removed.

Since magmas colder than 1000 °C are not observed in arc regions, we can conclude that heat is probably being supplied as quickly as it is being removed. We can only see two ways this may happen. Either convection in the asthenosphere or migration of the source region deeper with time will provide the additional heat required.

Temporal changes of arc volcanism

This leads us to an interesting speculation regarding the difference between arc volcanism on ocean–continent plate boundaries as opposed to ocean–ocean boundaries. We observe that progression with age of the volcanic front landward occurs on ocean–continent boundaries (Kawano & Ueda 1967; Matsuda & Uyeda 1971; James 1971; Hutchinson 1970) while ocean–ocean boundaries display no such clear cut large scale age progression (Schmidt 1957; Stark 1963; Ewart & Bryan 1972; Kaneoka, Isshiki & Zashu 1970). We concede that extension in marginal basins confuses the oceanic case (Karig 1971). We further note that old volcanic arcs such as the Cascades, Japan, or Aleutians, contain dominantly calc-alkaline adesites, whereas the oceanic arcs are dominated by tholeiites (Miyashiro 1974). Miyashiro (1974) suggests that this change in the abundance of tholeiitic and calc-alkaline rocks is due to secular depletion of the wedge in basaltic contents, whereas Ringwood (1974) proposes a two-stage petrological model to explain the same phenomenon. Here we point to yet another possibility. Perhaps the basic structural difference between the continental and oceanic lithospheres of the plate above the subduction zone provides different mechanisms for the supply of new heat required to sustain volcanism for a long time.

The total thickness of the continental lithosphere may be more than 200 km (Jordan 1974). The only available heat source for continental arc magma sources is for the magma chamber to migrate deeper along the slip zone with time (Fig. 7(a)). Replenishing of the source by convection occurs only with addition of new slab melt which must be andesitic primary magma (e.g. Green & Ringwood 1968; Taylor *et al.* 1969; Fitton 1971; and Marsh & Carmichael 1974). Thus one might speculate that the continental arc volcanism migrates deeper with time and becomes more calc-alkalic with time for this thermal reason (Fig. 7).

Oceanic lithosphere above a subduction zone is thin however (perhaps <60 km thick, above a subduction zone. Kanamori & Abe 1968; Abe & Kanamori 1970). Asthenospheric convection is the prime mechanism for resupply of both heat and new source material at depths greater than the lithospheric thickness. Thus possible tholeiitic primary magma derived from the mantle wedge (Green & Ringwood 1968; Yoder 1969; Kushiro 1972; Miyashiro 1974; and Arculus & Curran 1972), may stay at a fixed depth and maintain a fixed composition with time (Fig. 7(b)). There will, of course, be gradation between purely oceanic to purely continental cases.

Secular change of arc volcanics

If there is no convection leading to the exchange of material between the upper mantle wedge and the surrounding parts of the upper mantle, continued generation of tholeiitic magma in the wedge results in the depletion of the wedge in the basaltic component. So, the wedge becomes gradually more refractory and harder. This means an increase in the thickness of the lithosphere overlying the wedge. If we may assume an evolutionary trend: immature arcs to mature arcs to continental margins, the thickening of the lithosphere should occur in the late stage of this evolution. This is particularly true under continental margins such as the central Andes (James 1971). In such a place, convective rise of material effectively leading to the generation of tholeiitic magma in the wedge no longer occurs, because such a rise would have occurred at depths less than 100–200 km which are already hardened. So, in this latter stage only calcalkalic magma would form from the descending slab.

The petrologic models presented here need further testing from both observational and theoretical points of view. Model calculations on the secular change in the thermal regime of the downgoing slab, taking full account of both magma production and asthenospheric flow, are presently underway.

Summary

Previous geophysical models treating the thermal structure of subduction zones are inaccurate because they fail to consider the thermodynamics of dehydration of the oceanic crust upon underthrusting. This first generation of thermal models has been extremely useful in demonstrating the basic physical processes involved in subduction and in back-arc volcanism, but geochemical models for generation of volcanic arc primary magmas should not be based upon their predicted thermal structures. It is not even clear if frictional heating is an important or viable process for generating heat along the subduction zone. Even if it is an important source of heat, it does not control melting. Alternatively it is not clear if non-frictional-heating models provide a mechanism for release of water necessary for melting at the proper depths below volcanic arcs. It is clear to us, however, that dehydration is occurring upon subduction of the oceanic crust, that dehydration controls the thermal environment in the subducted oceanic crust, and that the newly-released water controls melting below volcanic arcs.

Second generation quantitative models for the thermal structure of subduction zones must consider the thermodynamics of dehydration, and they promise to yield new insights into the thermal controls of volcanic arc magma generation.

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