The Oceanic Crust

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Summary
I attempt in this article to put in context recent studies of the oceanic crust, studies which led to the drilling of the uppermost few hundred metres of the igneous rocks of the crust on Leg 37 of the Deep Sea Drilling Project. We see that physical measurements – seismic studies, heat-flow measurements and magnetic studies in particular, combined with observations of rocks dredged from the sea floor, and rocks drilled on oceanic islands lead to the idea that the oceanic crust is rather similar to the ophiolite suites seen on land. However, the oceanic crust is altered by the interaction of sea water with hot rocks near the ridge crests, and by cold weathering as the rocks lie exposed to water for millions of years. These alterations affect the physical and chemical properties of the crust, and must play a large part in determining the composition of sea water itself. The thermal processes involved in the rock–sea-water interactions modify the simple model of a cooling plate, where heat transported by the lateral convection of sea-floor spreading is lost by vertical conduction into the oceans.

Introduction
Many of you will know that numbers of us at Dalhousie and at Bedford Institute of Oceanography have been concerned for some years with the floor of the oceans. This interest of ours culminated to the point where a proposal was made to the Joint Oceanographic Institutions for Deep Earth Sampling (JOIDES) to advise the Deep Sea Drilling Project (DSDP) that their vessel *Glomar Challenger* attempt to drill into the ocean floor one kilometre or so (if possible), below any sediments covering the floor, into the crystalline rocks forming the upper part of the oceanic crust. We asked the National Research Council for funds to support analysis of rocks that would be recovered; this would be expensive. A number of scientists gained experience in the problems to be encountered by drilling holes into the volcanic rocks of Bermuda and the Azores. The Deep Sea Drilling Project agreed to the proposal, and Aumento, Hyndman, and Ade-Hall sailed with scientists from the United States, the USSR, France and Germany on May 31st on *Glomar Challenger*. The Bermuda and Azores holes had worked out well, and NRC kindly granted us a Negotiated Grant for the work with DSDP. *Glomar Challenger* returned in late July having successfully completed holes at four sites, the longest 600 m into igneous rocks, and the total penetration being 3181 m. It seems sensible therefore to try to show why we were interested in being so involved in the ocean crust, and to attempt to give a summary of some aspects of our knowledge of it. I want in particular to look at its composition, and how it may change with time. This is clearly important to geologists working on land, observing ancient oceanic rocks, but it turns out, of course, to be of importance to oceanographers concerned with the chemistry of the oceans, and to geophysicists concerned with the properties of the interior of the earth.

There are far more measurements of physical parameters in the ocean basins than there are collections of rocks. So consequently we start with physical observations from seismic studies and heat flow, and ask what sorts of processes and what sorts of rocks can lead to these observations. We will turn first to rocks from land, such as those H. Williams at Memorial has described, and second to rocks dredged from the ocean floor. We will see that we can explain the seismic observations with a particular choice of rocks, and ask "can we fit the heat flow"? To answer this, and in particular the details, we again examine rocks on land, some from eastern Liguria, and some from the Azores. We find that the rocks postulated will do, so set about seeing how they change with time. They are exposed to sea water, and interact with it. This interaction can be measured in terms of physical and chemical properties, and so we have clues at least to the composition of rocks of the ocean floor some tens of millions of years old. The rocks are altered from their pristine state; they will not be the same when they are a part of a descending slab of lithosphere as when they started at a ridge crest. So they change the interior of the earth. The changes are caused by interaction with the sea, and they must play a part in keeping the ocean the way it is.

Physical Observations
We can see the need for physical measurements, and gain insight into the sorts of questions which should be asked by looking at a very altered lava dredged by D. H. Matthews (1971) in 1958 from small seamounts in the eastern Atlantic, far from the crest of the Mid-Ocean ridge. Some of its properties are listed in Table I, where they are compared with the average of analyses for fresh igneous rocks, fresh lavas, just formed at the Mid-Atlantic Ridge. The differences in properties are substantial. If *all* the new crust formed at mid-ocean ridges each year (about 10 km^3) were so enriched in potassium as Matthews'

Table I
Weathering of oceanic basalts

	On-axis [1] Mid-Atlantic Ridge at 45°N	**Off-axis** [2] Swallow Bank
SiO_2	50%	47%
K_2O	0.2%	3%
Velocity (P)	4.8 km/sec	2.8 km/sec
Heat production	1.5 cal/gm/my	10 cal/gm/my

(1) From Aumento and Gunn (1975), Barrett and Aumento (1970) and Hyndman and Rankin (1972).
(2) From Matthews (1971).

rocks were, extracted presumably from sea water, we would have an equivalent flow of heat of about 0.3 µcal/cm²/sec from the ocean floor. This is significant, because the average value is about 1.5. The difference in compressional wave velocities is substantial, and should be detected in seismic surveys, or at least be indicative of alteration of rock samples on the sea floor.

Refraction velocities. Raitt (1963) analysed many of the results of seismic refraction surveys up to 1960 and showed clearly the division of the oceanic crust into three layers above the mantle, on the basis of compressional wave velocity measured indirectly by experiments at sea. Christensen and Salisbury (1974) have recently made a comparable analysis using data from experiments done since 1965; their study confirms Raitt's results. Layer 1 is composed of sediments and I will not concern myself with it at all. What is of concern are Layers 2 and 3. Suggestions have been made more recently that these "layers" can be subdivided in a meaningful way, and I show these results with Raitt's in Table II. What are the rocks responsible for these velocities? Are there really "layers" or are they an artefact of interpretation of the data?

Heat flow. The average heat flow from the continental crust and the oceanic crust is about 1.5 µcal/cm²/sec. In the case of the continental crust it is fairly clear that the heat arises from two separable sources, the contribution of radiogenic heat from the crustal rocks themselves, and

Table II
Compressional Velocities from Refraction Studies

Layer	Raitt's Compilation		Recent Suggestions
2A	3.3 - 6.0	Crust	2.3 - 3.8
2B			4.1 - 5.9
3A	6.1 - 7.0		6.5 - 6.8
3B			7.0 - 7.7
4	8.1	Mantle	8.1

Raitt's compilation from Raitt (1963). Recent suggestions from Fox *et al.* (1973). Values in km/sec.

the contribution by conduction from the mantle. The first measurements on the ocean floors showed that there was approximate equality between oceanic and continental heat-flow, and Sir Edward Bullard thought this was puzzling, because oceanic crustal rocks were thought to be much less radiogenic than continental rocks, and hence less hot. We will see by the end of the article that the puzzle now may be that oceanic heat-flow is not more than continental heat-flow! But the observation illustrates our interest in the content of the common radiogenic elements in oceanic rocks.

It is generally true that high heat-flow values are found near the crests of mid-ocean ridges, the spreading centres, and lower values away from the crests. A simple *model* (McKenzie and Sclater, 1969) says that newly formed hot rock is intruded from depths of tens of kilometres at ridge crests; being hot, there is a high temperature gradient. It loses the heat convected laterally through sea-floor spreading by conduction into the ocean as it spreads from ridge crests, and the temperature gradient falls (Anderson, 1972; Hyndman *et al.*, 1974; Lister, 1970). But in detail this model cannot be true as is shown rather clearly in Figure 1. We have over the crest proper, a *low*. We have for crust two to four million years (m.y.) old values near those predicted. We have beyond that low values, and finally, from 7 to 10 m.y., values again near those predicted.

We have therefore, in a sense two questions at this stage: what rocks account for the observed seismic velocities, and what rocks and processes account for the observed heat-flow? We will look first at rocks and processes which could account for the low values of heat flow at the crests of ridges.

Processes – Liguria and the Azores

Spooner and Fyfe (1973) have described a suite of rocks from eastern Liguria, Italy, which, for reasons which will become clear, they think were once part of the oceanic crust, formed at a mid-ocean ridge. The particular assemblages of minerals found at different depths in this complex of rocks led them to believe that the temperature gradient must have been some 1.3°C per metre, equivalent to a heat flow of 30 times the average sea-floor heat-flow, over some hundreds of metres. They point out that very elementary considerations lead us to account for much of the oceanic heat-flow; if we take only the latent heat of crystallization and the heat capacity of the oceanic crust formed per annum, a heat flow of several units arises if all this heat is concentrated in a 100 km wide strip across the ridge crests. Spooner and Fyfe suggested that a geothermal sea-water system was responsible for the thermal regime and for much of the mineralogy observed. Sea water enters the rocks from above, is heated by the hot rocks and ascends, incidentally stripping the rocks of a number of elements of interest, which are precipitated elsewhere.

Now a number of people (Anderson, 1972) have suggested that geothermal systems account for the *low* heat flows observed *at* the crests of ridges (see Fig. 1), because they are convective, not conductive, but Spooner and Fyfe suggest the geothermal system accounts for the *high* thermal gradient their rocks demand. This seems paradoxical, but a system which behaves this way was observed in the Azores bore-hole in 1973 (Muecke *et al.*, 1974). The temperatures found during drilling are

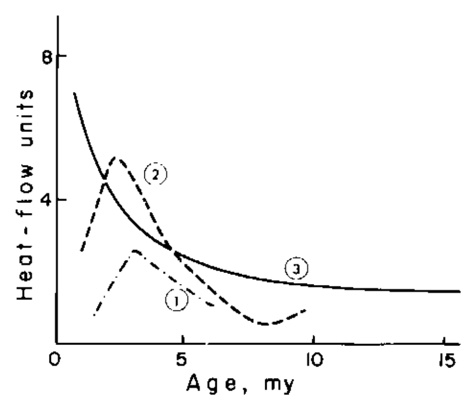

Figure 1
Profiles of heat-flow against age of (1) the Mid-Atlantic Ridge at 45°N (from Hyndman and Rankin, 1972), (2) the Juan de Fuca Ridge (from Lister, 1970), and (3) theoretical (from McKenzie and Sclater, 1969). Redrawn from Anderson (1972). One heat-flow unit is 1 µcal/cm² sec.

shown in Figure 2. We observe (1) very low gradients – very low heat-flow – in the uppermost 150 metres, (2) very high gradients between 150 metres and 550 metres, about 0.25°C per metre. This is just what is needed; low gradients near the surface (low heat-flow) and high gradients beneath to alter the minerals appropriately. Incidentally, this hole was closed when steam gushed higher and higher, causing concern to the man on top of the rig. But the Portuguese Electric Company are re-opening it for geothermal power studies, an interesting offshoot from research sponsored by the International Decade of Ocean Drilling, the Research Corporation, NRC and Dalhousie.

Newfoundland Ophiolites

Our colleagues at Memorial have told us that rocks like those of eastern Liguria are to be found in Newfoundland, and they too have features which suggest they were formed as a part of the oceanic crust. "Ophiolites" are a complex of pillow basalts (basalts poured out into water where the characteristic pillows develop) underlain by more massive basalts; intrusive rocks (as a sheeted complex of dykes); coarse grained intrusives such as gabbros and rocks

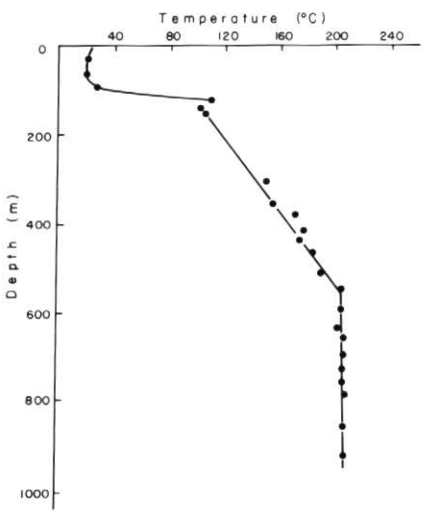

Figure 2
Temperature measurements in the Azores bore-hole. Above 140 m: temperatures 1 to 3 hours after termination of water circulation. Below 140 m: temperatures during drilling. (From Muecke et al., *1974).*

where there has been an accumulation by gravity settling of minerals such as pyroxenes, i.e., cumulates; and ultramafic rocks chemically low in silica, mineralogically dominated by olivines and pyroxenes. Examples are to be found in Troodos, Cyprus (Gass and Smewing, 1973), and Ming's Bight, Newfoundland (Peterson *et al.*, 1974). A summary is given in Figure 3. The important point for us here is that velocities of samples from the Ming's Bight complex are directly comparable with the velocities

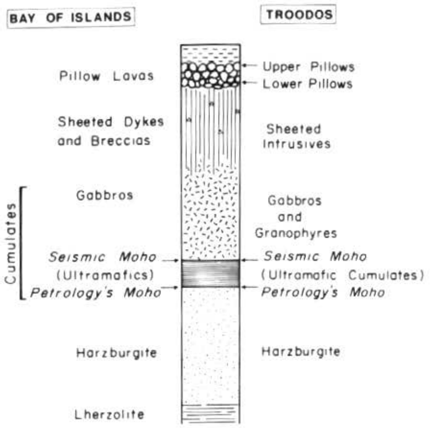

Figure 3
Two ophiolite complexes. The left hand diagram is a sketch of the Bay of Islands complex, redrawn in a simplified fashion from Figure 3 of the G.A.C./M.A.C. 1974 Field Trip Manual A1 (after Williams et al., *1972). The right hand diagram is a sketch of Troodos, Cyprus, redrawn from Gass and Smewing (1973).*

deduced from refraction studies at sea, seen in Table II. But what are the ocean-floor rocks?

Barrett and Aumento (1970) showed that an igneous rock "stratigraphy" could be deduced from a careful study of dredge-hauls on the Mid-Atlantic Ridge. They concluded that the igneous rocks were characteristically "layered" with vesicular and pillowed basalts on top, underlain by massive basalts, metamorphosed basalts and metamorphosed gabbros. This is, of course, similar to the rock types and stratigraphy of the Newfoundland ophiolites. Christensen and Salisbury (1974) and Fox *et al.* (1973) studied the velocities of dredged igneous rocks. The picture we get – a synthesis of the evidence from refraction studies at sea, the stratigraphy of igneous rocks dredged from the sea floor, the stratigraphy of ophiolites on land, and velocities measured in the laboratory of the dredged rocks and the ophiolites – is shown in Figure 4.

Alteration under Hot Conditions

The igneous rock assemblages produced at the crests of mid-ocean ridges cannot stay in their pristine state. They will be initially under thermal gradients of at least 0.15°C/m. They will fracture, which metamorphoses the rocks involved in the development of the fracturing, allows sea water access to the hot rocks, and exposes them to view as

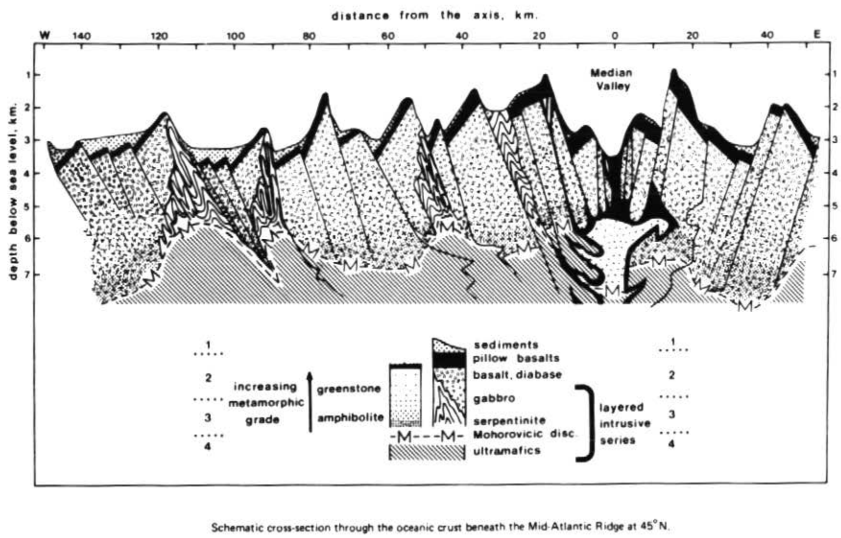

Figure 4
The oceanic crust. A synthesis by Aumento et al. *(1971) of the Mid-Atlantic Ridge at 45°N.*

it were, available to sampling by dredge hauls. As the rocks move away from ridge crests they are covered with sediment, and exposed to different thermal regimes where water has less ease of access. Little systematic attention appears to have been devoted to metamorphism within the oceanic crust as a function of distance from ridge crests, and to the inevitable effects of metamorphism on physical properties, because of the lack of samples. By contrast, we do seem to know a good deal about cold processes – the interaction of basalts exposed to cold sea water on the ocean floor – and we return to this later. We want here to look at metamorphism and heat flow, and comment on velocities so far as we can.

Two Sorts of Metamorphism

First, we have metamorphism under thermal gradients of, say, up to 0.15 °C/m, with only limited access of sea water. This leads to metamorphism of the sort described by many authors for mid-ocean ridge dredge hauls, and for ophiolites from land. Consequently, we see assemblages of minerals characteristic of the zeolite, greenschist and amphibolite facies. A question arises immediately about seismic velocities; is the change from Layer 2A to 2B, or from Layer 2 to 3 a change in rock type? Although the studies of metamorphosed rocks from the sea floor by Fox *et al.* (1973) suggested this is so, on Troodos a change in velocity was associated with the change from zeolite to greenschist facies, and the question may deserve more study than it has had.

Second, we have metamorphism under the sort of conditions described by Spooner and Fyfe (1973) in Liguria. Sea water interacts with very hot rocks, leading to massive changes in rock type. Thermal gradients are known in the Azores to be 0.25°C per metre, and 1.3°C per metre in Liguria. The sorts of reactions involved are illustrated in Table III. Oxidation leads to the introduction of magnetite and hematite, and the release of silica into solution. The introduction of sodium from sea water alters the plagioclase feldspars so they are richer in albite. Carbonates precipitate because the solubility of calcite and aragonite decreases with increasing temperature. And so in Liguria *and* in the drill core from Bermuda we see mineral assemblages such as albite-calcite-hematite, replacing the original calcic plagioclase-pyroxene-olivine assemblage. In Liguria the geothermal system was hotter at the bottom than at the top and mineral assemblages change vertically. Except for magnetite replacing hematite at depth, this has not been detected in the 800 m hole in Bermuda (Aumento and Gunn, 1975).

Table III

Oxidation reactions suggested by Spooner and Fyfe

Fe_2SiO_4 Fayalite	+	$\frac{1}{2}O_2$			=	Fe_2O_3 Hematite	+	$\frac{1}{2}(SiO_2)$ solid Quartz			+	$\frac{1}{2}(SiO_2)$ solution		
$11Fe_2SiO_4$	+	$2SO_4^{--}$	+	$4H^+$	=	$7Fe_3O_4$ Magnetite	+	FeS_2 Pyrite	+	$11SiO_2$	+	$2H_2O$		

A principal point that emerges from studies of such metamorphic assemblages is that they involve hydration, which is exothermic. We have another source of heat to account for the heat flow of the ocean floor. R. A. Hart (1973), using data of Helgeson, estimates that the contribution from this source could be up to 0.5 μcal/cm²/sec.

We are now in a position to return to our original heat-flow profile (Fig. 1). At the ridge axis, low conductive heat-flow values are found, reflecting the presence of convective systems. As the rocks move away from the axis fractures may be sealed by metamorphic processes, or sea water may be denied access to the rocks because sediments seal the igneous rocks beneath. The convective systems cease. Consequently, the normal geothermal gradient will be re-established, and temperature gradients and heat flow must rise to the values predicted by the conduction model for a cooling plate. If temperatures rise enough, it is possible that dehydration may occur. Such reactions are endothermic; so heat is locked up and until the reactions are completed heat flow will be low. This is the model put forward by R. N. Anderson (1972), and by Hyndman and Rankin (1972). We do not seem to have any proof that the dehydration takes place. Presumably we should look to results of refraction studies and to laboratory measurements of velocity on appropriate rocks. The right measurements have not so far been made.

Alteration: Cold

Igneous rocks sitting for millions of years on the ocean floor exposed to cold sea-water will suffer a sea change. We can ask questions such as: how far down do these cold reactions go? how fast do they operate? how do they operate – along cracks and fissures, or through atom-by-atom substitution only? what happens to the physical properties? what happens to the chemistry? We seem to know quite a lot about this.

(1) *"Velocities and weathering".* Measurements of velocities in basalts collected to depths of several tens of metres have been made in holes drilled on several Legs of *Glomar Challenger*. The velocity generally increases with depth to values around 5.5 km/sec, and there values at depth are higher than we might expect from the results mentioned so far from dredged rocks, and from refraction results. Now a plot of velocity against density from rocks of the ocean floors shows clearly that the two are related (Christensen, 1972; Christensen and Salisbury, 1972; Christensen and Shaw, 1970); some results are shown in Figure 5. Obviously, weathering of basalts as they sit on the floor of the ocean is taking place. We can ask: how fast is it happening, and how is it happening? Knowing the age of the rocks, Hyndman calculated the rate of progression of weathering downwards, and it is much too fast for simple atom-by-atom diffusion. We see that this is so shortly, but it tells us what we might have guessed, that sea water attacks the rocks by

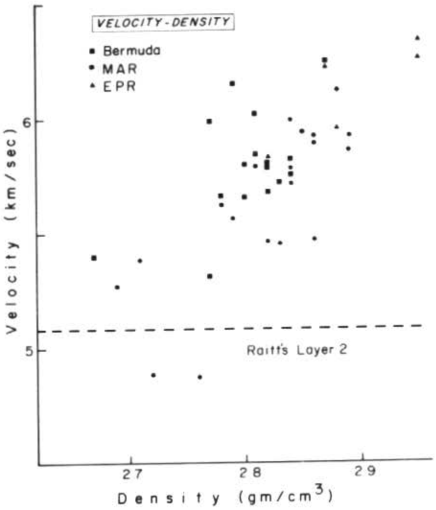

Figure 5
Compressional wave velocity and density for oceanic igneous rocks. Dots – Bermuda lavas at 1 atmosphere (unpublished measurements by R. D. Hyndman); vertical bars – Mid-Atlantic Ridge (after Christensen and Shaw, 1970); squares – East Pacific Rise (after Christensen, 1972).

percolation through cracks and fissures. This may be an important point. Velocities in drilled, unweathered basalts are higher than interpreted from many refraction experiments. The presence of cracks and fissures and the presence of rubbly basalts associated with more massive basalts must reconcile the two. Velocities of dredged rocks are comparable to refraction velocities, and lower than velocities in unweathered, drilled rocks. "Weathering" reconciles these anomalies. In a sense, cracks and fissures must lower the overall bulk density, and weathering must lower the density of individual rock fragments. Suppose you want to interpret marine gravity values – you need density. Should you estimate density from individual sample density or from velocity-density relationships on individual unweathered samples? Probably not, because the values that result will be too high. It may be better to estimate density from refraction velocities, as has usually been done. The velocity studies tell us weathering has occurred. I will illustrate the processes involved in two ways, physically and chemically.

(2) *"Magnetic properties and weathering".* Irving (1970) first suggested as a result of experiments on rocks collected at sea that the high magnetic field values measured over mid-ocean ridge crests arise because the older basalts to either side have been weathered and their magnetization has decreased. Patrick Ryall (1974) at Dalhousie has looked at this process in detail in two ways. He has studied the change in

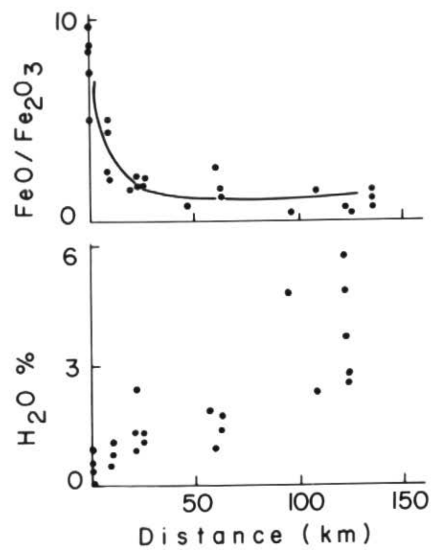

Figure 6
Changes in weathering with distance from a ridge axis. Oxidation ratio FeO/Fe_2O_3 (upper), and % water (lower), versus distance from the axis of the Mid-Atlantic Ridge at 45°N (from Irving, 1970).

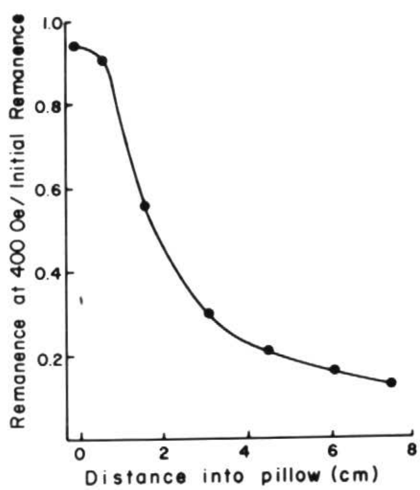

Figure 7
Change in magnetic hardness – a measurement of oxidation state – from the outside to the inside of a pillow from the Mid-Atlantic Ridge (from Ryall, 1974).

magnetization within individual pillows of lava, where one should see changes in one place with time, and he has studied the differences between pillows of different ages. Weathering is oxidation: Figure 6, taken from Irving's work, shows that the iron-titanium minerals at increasing distances from the ridge crests are progressively poorer in FeO by comparison with Fe_2O_3. The oxidation takes place from outside inwards in any one pillow. Figure 7, a graph of the change in magnetic hardness (a measure of oxidation state) shows this. Weathering takes time. Ryall also cooked his rocks in sea water under pressure at 150°C for thousands of hours, measuring their changes in magnetization as they cooked. This is an attempt to simulate processes at 0°C, which take millions of years, in the time available to a Ph.D. student. There is a decrease in magnetization with increased cooking, as is found on the sea floor.

Oxidation processes affect other properties such as resistivity. We need to know resistivity of ocean rocks because one day we may want to seek economic minerals on the floors, using electrical methods.

(3) *"Chemistry and weathering".* The changes in chemistry as basalts interact with sea water can be seen in a study of flows from Bermuda. Figure 8, work of Aumento and Gunn (1975), shows us that there may be spectacular changes across one flow, reflecting intensity of interaction in time with sea water. I show here only the results for two elements: the flow is enriched at top and bottom in

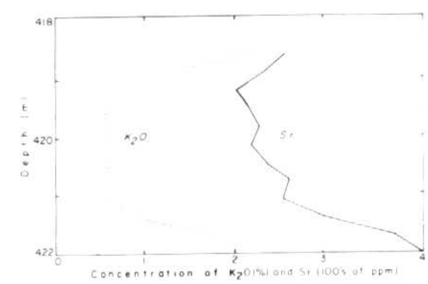

Figure 8
Change in chemistry across one lava flow from Bermuda. Potassium as K_2O (%) and strontium (100's of ppm) against depth in the bore-hole. (From Aumento and Gunn, 1975).

potassium by a factor of four. If you had sampled in the wrong place you might not think it was an oceanic tholeiite at all.

Aumento and Hart have worked out what happens overall in cold weathering between sea water and basalt. Table IV is a summary based on Aumento's work. As Aumento points out, the net effect is to produce a rock with (slightly) less silica and more alkalis than its parent. This is a similar trend to that observed in the production of alkali-rich igneous rocks.

We saw in Figure 5 that velocities measured in the laboratory correlate directly with density: low velocity, low density, and vice-versa. We saw also that weathered basalts have low velocities, and that weathering takes time. Clearly then we may in some crude way be able to relate refraction velocities measured in the ocean basins to chemical processes of weathering and metamorphism in general, and to the age of the crust – as Christensen and Salisbury (1972) showed (Fig. 9). If this is true then there may be some hope of tracing the changes in the igneous rocks of the ocean floor as they spread from the ridges, and eventually answer the question: what does the crust look like as it plunges down a subduction zone?

Roger Hart (1973) has gone some way to seeing if the changes in the crust can be found indirectly, from velocities found in refraction studies. We see in Figure 10 changes in velocities measured in the laboratory with changes in the oxides of some major elements, for samples recovered by dredging and drilling. Weathering is hydration – velocity goes down with increase in water content – and weathering is gain of potassium from sea water, and so on. Hart also estimated the changes to be expected in metamorphism and asked the question: how thick a layer of crust must interact with sea water to account for the known input of elements from rivers, and the known composition of sea water? He found for example, that only 0.7 km thickness of low temperature weathering would account for most of the potassium.

Table IV
Weathering changes

	Rock	Sea Water
K	Up	Down
Na	Up	Down
H_2O	Up	Down
Sr	Up	Down
Ba	Up	Down
Bo	Up	Down
Rb	Up	Down
U	Up	Down
Ce	Up	Down
SiO_2	Down	Up
Ca	Down	Up
S	Down	Up
Al_2O_3	Same	Same

From a compilation by Aumento (1974).

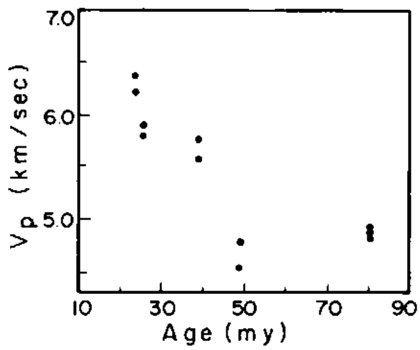

Figure 9
Change in compressional wave velocity with age of igneous rocks from the Mid-Atlantic Ridge. (From Christensen and Salisbury, 1972).

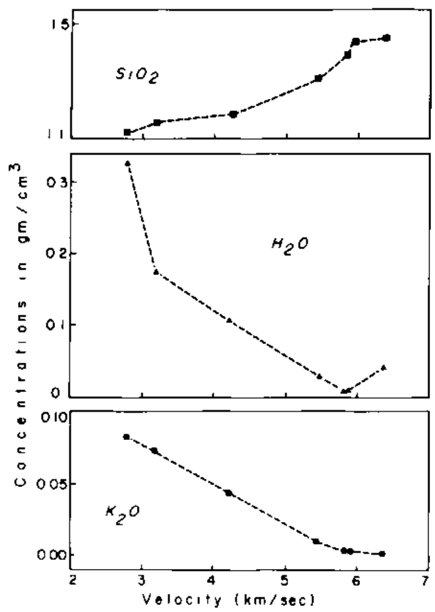

Figure 10
Change in compressional wave velocity with composition of oceanic igneous rocks (From Hart, 1973).

Problems

Our questions can start from the point we have just left. How far from the ridge crests is there evidence that the processes of weathering and metamorphism described extend? Figure 11 reproduces Hart's compilation of refraction velocities for Layer 2, reported by various workers. Although we should treat such plots with a good deal of skepticism it does seem that velocities on slow-spreading ridges, such as the Mid Atlantic Ridge and the Carlsberg Ridge, decrease systematically for some tens of millions of years, and may increase again beyond that. But on the East Pacific Rise such a trend is not at all obvious – is the Rise, as reported from Deep Sea Drilling Results, a much less weathered ridge, which would account for the rather higher velocities beyond 400 kms on the Mid-Atlantic Ridge? Is the *increase* real? Christensen and Salisbury noted it too. If real, what processes or rock types does it reflect?

The observation may be supported by Bermuda results, where the average velocity has been found to be 5.8 km/sec, in agreement with high values found in refraction studies. Perhaps the phenomenon is associated with the other anomaly of Bermuda (Aumento and Gunn, 1975): 30 per cent of the rocks recovered consisted of extremely alkaline rocks, the like of which have seldom been found

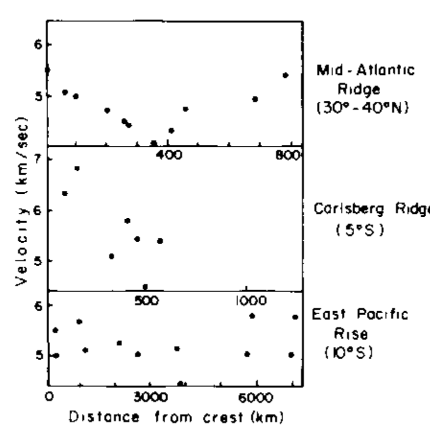

Figure 11
Layer 2 seismic refraction compressional velocities versus distance from the crests of three mid-ocean ridges. (From Hart, 1973, where the sources of data will be found).

(Fig. 12). It is curious too that it is likely that the alkaline rocks were intruded into the tholeiites, making a large part of the island some tens of millions of years after the tholeiites were extruded.

The central problem perhaps is this: can we map the composition of the oceanic crust throughout the ocean basins? By themselves studies of compressional wave velocities in experiments at sea and in the laboratory have been most helpful, but Christensen and Salisbury (1974) in a very thoughtful paper point out that measurements of shear wave velocities as well as compressional wave velocities, from which Poisson's ratio can be calculated, may be much more diagnostic of rock type. The difficulty is that shear waves do not propagate in a fluid, and in marine experiments where observations are made at the sea surface, shear waves have to be identified by using compressional waves generated by conversion of shear waves in the solid crust. This is difficult to do. One can turn to observations from ocean bottom seismometers, but there are only a few of these. Developments along these lines may be fruitful.

There are many problems – the reader can think of dozens, no doubt. *Glomar Challenger's* holes of Leg 37 this year were along one line of spreading away from the Mid-Atlantic Ridge axis at 37°N, in oceanic crust ranging in age from 3.5 to 16 million years. A great variety of igneous rock types were recovered; as an example,

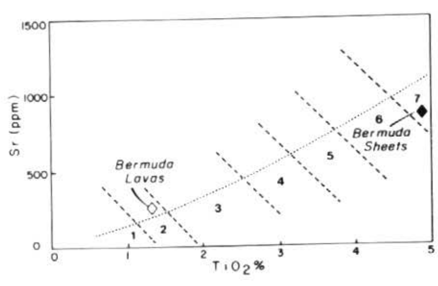

Figure 12
Changes in Sr and TiO_2 for various groupings of lavas, and for the Bermuda lavas, and intrusive sheets. (From Aumento and Gunn, 1975). 1 – island arc andesites, 2 – mid-ocean ridge basalts, 3 – island tholeiites, 4 – mildly alkaline lavas, 5 – highly alkaline basanitic lavas, 6 – limburgites.

the scientists stated on return that the major element composition of all the lavas recovered correspond to tholeiitic basalt, but the *range* of their composition is almost as great as that for *all* tholeiitic lavas dredged from the world's mid-ocean ridge system. Nevertheless, the holes drilled should be a good start towards comprehending the crust of the ocean floor and its interaction with the rest of the solid and fluid earth.

Summary
Seismic velocities measured in refraction studies at sea can be compared with laboratory studies on dredged and drilled rocks and rocks from ophiolite suites. They lend credence to the view that ophiolite suites are comparable to the igneous rocks of the ocean floor, and that the ocean crust is a complex of fresh basalts and gabbros, their metamorphosed equivalents, ultrabasic rocks and their serpentinised equivalents. Measurements of heat flow tell us that the simple model of a cooling plate developed by McKenzie (1967) has to be modified, because it appears that sea water interacts with hot rocks of the ridge crests, altering the rocks, and changing the thermal regime. The rocks formed at ridge crests are altered also by weathering as they interact with the ocean, while sitting for millions of years on the sea floor. These changes can be measured by changes in chemistry, magnetic properties and elastic properties. The interactions must play a substantial part in determining the chemical composition of sea water, and perhaps in changing the composition of the interior of the earth, because the lithospheric plate which descends at subduction zones is not the same as the plate formed at ridge crests.

Note:
This article is based on a talk given to the Royal Society of Canada in June 1974, and depended heavily on discussions with Drs. Ade-Hall, Aumento, Clarke, Hyndman, Muecke, Reynolds and Ryall at Dalhousie, and Dr. C. E. Keen of the Atlantic Geoscience Centre. Jan Aumento drafted the diagrams. I am grateful.

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