Geometry of plate accretion

J. F. DEWEY
W.S.F. KIDD
Department of Geological Sciences, State University of New York at Albany, Albany, New York 12222

ABSTRACT

A steady-state model for plate accretion at oceanic ridges, developed using constraints mainly from western Newfoundland ophiolite complexes, involves four main rheological components: (1) a convectively cooling lid accelerating from the ridge axis across a zone of decreasing dike-injection rate and thickening by the addition of extrusive basalts above and gabbro underplating beneath; (2) a wedge-shaped magma chamber with a flat floor; (3) a differentially subsiding wedge of cumulates; and (4) a narrow axial partially melting lherzolite-derived diapir from which basalts are liberated over a narrow axial welt into the magma chamber and from which residual harzburgites are plated near the axis below the base of the subsiding cumulates. The axial depth of the magma chamber determines the thickness of plated gabbro. The basal width of the magma chamber and the height of the partially melted welt control the thickness of cumulates and the final attitude of cumulate banding.

INTRODUCTION

Models for the geometry of plate accretion at oceanic ridges have been developed by several authors using a variety of data bases and constraints. Adopting the view that ophiolite complexes (Penrose field conference participants, 1972) represent detached slices of oceanic crust and lithosphere, Church and Riccio (1974), Dewey and others (1973), Greenbaum (1972), and Moores and Vine (1971) have used geometric, petrologic, and structural data to construct models. The models of Cann (1974), Moore and others (1974), and Sleep (1975) were developed using petrologic and theoretical geometric and thermal constraints from oceanic data. Numerous workers have pointed out that the thickness and petrology of the internal units that compose well-developed ophiolite sequences match our present understanding of the oceanic crust very well in a general way and, although the thickness of individual ophiolite complex units may vary considerably, the range of thickness variation is comparable with layer thickness variation recorded in the oceans (Coleman, 1971; Dewey and Bird, 1971; Moores and Jackson, 1974).

The general sequence of rock types in ophiolite complexes, particularly in those in western Newfoundland with which we are most familiar, consists, from base to tope, of harzburgite–minor dunite–rare lherzolite, with metamorphic textures and a complex and pervasive high-temperature deformation history; cumulate dunite below a cumulate transition sequence of dunite, feldspathic dunite, wehrlite, feldspathic wehrlite, clinopyroxenite, and gabbro containing zones affected by penetrative high-temperature deformation; banded cumulate gabbro, clinopyroxenite, and anorthositic gabbro; more isotropic and homogeneous, somewhat leucocratic gabbro with irregular pods and veins of trondhjemite near the top and with diabase dikes increasing in abundance toward the top; a sheeted diabase dike complex with minor trondhjemite bodies near the base; and a pillow lava complex with an increasing percentage of dikes, and sometimes sills, toward the base.

We present here a steady-state model for the geometry of plate accretion, with constraints drawn mainly from our observations in the following ophiolite complexes in western Newfoundland: Bay of Islands Complex (Williams, 1973), Mings Bight Complex (Norman and Strong, 1975), and Betts Cove Complex (Upadhuay and others, 1971). The model is generalized and does not consider the topography and shallow structural morphology of ridges but rather attempts to account for, in particular, the thickness ratio of isotropic gabbro to cumulates, the geometry of cumulate layering, and the progressive geometric development of the pillow and sheeted complexes. Furthermore, we avoid a discussion of layer thicknesses and the problems of precise correlations of ophiolite complex layering with oceanic crustal layering.

CONSTRAINTS FROM OPHIOLITE COMPLEXES

The following list of constraints imposed on the geometry of plate accretion is drawn mainly from our own and published observations on the Bay of Islands, Mings Bight, and Betts Cove Ophiolite Complexes in western Newfoundland but include published observations on other ophiolite complexes, including Troodos (Gass and Smewing, 1973; Greenbaum, 1972; Kidd and Cann, 1974; Moores and Vine, 1971) and Vourinos (Jackson, and others 1975; Moores, 1969).

1. In most ophiolite complexes, from 500 m to 1 km of pillow lavas are developed, within which the percentage of dikes increases downward until the base is defined as the level at which the lowermost screen of pillow lava is found, indicating that lateral extension strain increases downward from zero at the top to infinite at the base. This shows that the complete pillow complex cannot be accumulated in batches of full thickness, but that the pillow complex, as it accelerates from the ridge axis to the half–spreading rate, must grow from zero thickness at the axial singularity (exact axis of injection at any time), where the top of the sheeted complex must lie just below the surface, to its full thickness some distance from the singularity in a zone of decreasing lateral-extension (dike-injection) rate.

2. Sills become common toward the base of the pillow complex in the Betts Cove Ophiolite, indicating that sill injection accounts for a part of the thickness growth of the lower part of the pillow complex. We suggest that this may be common, but as yet unrecognized, in other ophiolite complexes.

3. Sheeted complex dike widths are commonly in the range 20 cm to 1 m, which, for a half–spreading rate of 1 to 10 cm/yr, indicates an average of one dike emplaced every 1 to 50 yr.

4. Sheeted complex dikes commonly preserve only one chilled margin with a preferred chilling direction, indicating that dikes are split to allow further dike injection and that most injection takes place over a very narrow zone away from which dikes are injected at a rapidly decreasing rate.

5. Sheeted complex dikes are notably and consistently aphyric. The fissures in which the dikes form must either be tapping magma without phenocrysts or tapping fluid segregating from a mush of crystals and interstitial fluid. If the latter is occurring, the composi-

Geological Society of America Bulletin, v. 88, p. 960–968, 3 figs., July 1977, Doc. no. 70707.

tion of the liquid should not be very different from the primary magma if crystallization of the mush crystals has mostly occurred around a cotectic, which is likely in the light of oceanic data (T. Shibata, 1976, verbal commun.) and from the fact that the more leucocratic homogenous gabbros in ophiolite complexes do not, in our experience, show significant zoning of individual mineral grains. It is possible that the crystal mush, where present, may ascend some way up from the bottom of any dike-filled fissure; this may partly account for the absence of sharply defined bases of individual dikes in ophiolite gabbros.

6. There is a strong contrast in grain size between dikes and gabbro, indicating a rapid cooling of gabbro before the emplacement of a particular dike, supporting the idea (Anderson, 1972; Lister, 1972) of hydrothermal convective, rather than conductive, cooling of the upper part of the oceanic crust as it travels from the axial singularity. We conclude that the depth of the transition from sheeted complex to gabbro and, therefore, the thickness of the 100% sheeted dike layer is most probably controlled by the lower limit of hydrothermal circulation at the axial singularity. The numerous anastamosing breccia zones observed within the sheeted complex in some ophiolite complexes also suggest pervasive hydrothermal circulation (Williams and Malpas, 1972).

7. In western Newfoundland ophiolite complexes, the transition from 100% sheeted dike complex to gabbro with less than 5% dikes takes place over a vertical thickness of about 250 m when viewed on a broad scale. On a finer scale, the transition is not homogeneous; at any level within the transition, zones a few tens of metres wide consisting mainly of dikes alternate laterally with similar zones consisting mainly of gabbro. In Mings Bight, one example about 300 m wide consists, near its base, of 95% dikes penetrating about 200 m lower into gabbro than nearby. Each zone of greater dike concentration up to a few hundred metres wide and deep was therefore a temporarily preferred site of dike injection.

8. In the homogeneous gabbro, dikes decrease in percentage downward, although we have not traced any one unambiguously to its roots. Dikes do not gradually increase in grain size to merge with gabbro. This means that the amount of lateral extension decreased from infinite at the base of the sheeted complex to zero at the base of the homogeneous gabbro where the last dike dies out, in turn indicating that individual dikes cut solidified gabbro and are then underplated by gabbro as they move away from the axial singularity, just as dikes and flows in the pillow complex are progressively overplated by lava flows.

9. Rocks of trondhjemitic type are found in the homogeneous gabbro, generally where it is cut by 5% to 20% dolerite dikes. A few occur within the base of the 100% sheeted dike layer. These bodies range from small individual tension gashes and veins through areas of net-vein breccia a few metres to tens of metres across and, exceptionally (Troodos), bodies as much as 1 km in original horizontal width. They are clearly associated with the sheeted-complex–gabbro transition and are part of the ophiolitic magmatic sequence, because many examples cut some dolerite dikes but are cut by others. One example in Troodos is so greatly dissected by dikes that it must have formed very near the axis of injection. Many of these bodies show clear intrusive relationships (net-vein breccias); their mineralogy, texture, and relationships to adjacent rocks, particularly where they crosscut dolerite or contain dolerite or gabbro xenoliths, show that they are not directly hydrothermally altered gabbro, even though such alteration may occur locally.

10. The upper part of the overall gabbro layer, including that within the transition into dike complex, is generally very homogeneous but locally shows diffuse banding a few centimetres or less thick. This homogeneous gabbro is somewhat richer in plagioclase when compared to gabbros of, for example, isolated thick tholeiitic sills and dikes. Plagioclase flotation in a magma chamber is probably not necessary to explain this slight enrichment in plagioclase, particularly because the clinopyroxene content is still large and grains are not interstitial. Also, the presence of plagioclase-rich bands in the underlying cumulate gabbros suggests that plagioclase flotation is not important. The slight concentration of plagioclase may be entirely due to extraction of mafic phases by accumulation from the parent melt. The local diffuse banding may indicate slight changes in water pressure and (or) temperature in the magma chamber.

11. The thickness of homogeneous leucogabbro cut by diabase dikes is a poorly known parameter in well-preserved ophiolite complexes with thick gabbro layers. The upper part with abundant dikes is clearly not less than 300 m thick, but data on the depth to which less abundant dolerite dikes occur is difficult to find. The Mings Bight Complex contains diabase dikes through at least 750 m of gabbro. The Troodos Complex appears to contain diabase dikes nearly as far down as the base of the gabbroic layer; the thickness that this represents depends on the structural interpretation preferred for the central part of the complex. If it is a gentle dome (Wilson, 1959), it is not more than about 1 km; if the western side of the central part is the steep limb of a monocline (our interpretation), it may be as much as 3 km. The Betts Cove Complex has a very thin gabbro layer; where it is clearly cumulate, it contains no dolerite dikes, but where it is homogeneous and not provably cumulate, dikes are found throughout almost all of the 250 m of gabbro. The other complexes do not, at present, provide additional data.

It is important to emphasize that this discussion refers to dolerite and (or) gabbro dikes that do not cut the cumulate or the noncumulate ultramafic rocks. The fact that the dikes of the sheeted complex are "rootless" requires that they were fed from a magma chamber and that the gabbro they cut was plated onto the roof of that chamber. Later events affecting already consolidated oceanic crust are local and may include hot-spot–type magmatism, arc volcano construction in a rear-arc basin, and effects where crust passes a ridge-segment end across a transform-fault–fracture-zone transition. In all cases, dikes would be expected to cut comulate gabbros and ultrámafic rocks and the noncumulate ultramafic rocks. This can be seen in the Mings Bight Ophiolite Complex, where large porphyritic dolerite dikes cut all members of the ophiolite complex and feed large sills within the volcaniclastic-sediment–pillow-lava sequence that lies on the ophiolite complex proper.

12. Because of the difficulty of distinguishing plated roof gabbro from rapidly deposited homogeneous cumulate gabbro, the maximum thickness of cumulate gabbro in ophiolite complexes is poorly known. Although there is a transition between the cumulate ultramafic and gabbroic rocks, we use the definition of a significant content of olivine and (or) orthopyroxene to define the cumulate ultramafic layer, leaving clinopyroxene-rich bands unaccompanied on a broad scale by olivine or orthopyroxene-rich bands to be part of the banded cumulate gabbros, because this is an easily recognized field criterion even when the rocks are somewhat altered. Given this definition, not more than 500 m of banded cumulate gabbro has been proved to exist in well-preserved ophiolite complexes, although 2 km or more may exist in the Bay of Islands, Papua, and perhaps the Oman; Betts Cove and Mings Bight contain not more than 100 to 300 m.

13. The most reliably established section of cumulate ultramafic rocks comes from Vourinos (Jackson and others, 1975); even though it is composite and crossed by faults, a minimum thickness of 1 km of ultramafic cumulates is present, given the definitions above. This section show cycles of deposition, which may indicate batch or intermittent supply of magma. Interpretation of the structure of the western margin of the Troodos massif as a steep monocline limb leads to a maximum thickness of 1.5 km for the ultramafic cumulates, including the lower 300 m of dunite. The Bay of Islands Complex probably contains a similar thickness, with the lower part also being cumulate dunite and feldspathic dunite.

14. Although cumulate gabbros and transition-zone rocks and particulary cumulate dunite contain zones affected by penetrative high-temperature deformation, regular undisturbed cumulate layering in the gabbros and ultramafic cumulates demonstrates that ophiolite magma chambers have flat floors. Yet, in the Bay of Islands Complex, the orientation of cumulate layering relative to gross unit layering varies from roughly parallel at the base and top of the cumulate sequence to as much as 90° in the middle of the cumulate sequence.

15. Although detailed cumulate layering cannot be laterally traced for long distances, as it can in many layered mafic plutons, there is no evidence that crosscutting individual plutons can be defined within the gabbro and ultramafic cumulate layers. Therefore, a steady-state mechanism must be inferred for the development of cumulates and homogeneous gabbro.

16. The boundary between the cumulate and deformed noncumulate ultramafic rocks has been shown to be sharp and traceable for 1 km in the Vourinos complex. In the Bay of Islands Complex, this boundary is somewhat obscured because the lower ultramafic cumulates are in part penetratively deformed, a situation that also seems to be the case on Troodos at the base of the dunite. The basal cumulates must therefore be deposited on a noncumulate residuum sufficiently consolidated to preserve the gross cumulate layering but plastic enough to allow their deformation.

17. As the spreading process pulls the lithosphere apart, this harzburgite "basement" has to be formed at the magma supply axis immediately prior to deposition of cumulate material on it; it cannot be much older basement as some have previously suggested.

18. The lack of intrusive crosscutting mafic material in the residual harzburgite and overlying ultramafic cumulates of well-preserved ophiolite complexes show that the extraction of partial melt at the supply axis is extremely efficient and must occur over a very narrow zone closer to the supply axis than the position at which the first cumulates are deposited.

19. The occurrence of residual harzburgite immediately below cumulates and the existence of at lease 7-km thickness of harzburgite in part of the Bay of Islands Complex show that efficient melt extraction and residuum plating occurs from the level of cumulate deposition and that it must continue progressively to at least 7 km below that level.

20. Where magma is released into the chamber at the level of cumulate deposition, it is possible that residual olivine and orthopyroxene crystals may be expelled periodically or continuously some way into the magma before falling back. If this process occurs, (1) cumulate layers of residual crystals may be found in basal cumulates, and (2) there may be (or have been before deformation), in places, a transition between noncumulate, plated harzburgite and cumulate harzburgite composed of residual crystals.

MODEL

Our steady-state model for the evolution of the zone between the ridge axis singularity and a position where the oceanic crust is finally consolidated and accelerated to the half–spreading rate is illustrated in Figure 1. This zone consists of four rheological components; a lid thickening from the axial singularity, a wedge-shaped magma chamber from which gabbro is plated to the base of the lid, a subsiding cumulate zone beneath the magma chamber, and a narrow welt of harzburgitic residuum discharging basaltic magma over a very narrow zone at the axial singularity.

The lid consists, at the axial singularity, of the full thickness of the sheeted complex, the top of which is just below a thin capping of pillow lavas fed by axial dikes and the bottom of which is tapping the magma chamber (13 in Fig. 1). Dike chilling statistics (Moores and Vine, 1971; Kidd and Cann, 1974) indicate that dikes are injected over a narrow axial zone; this is supported by magnetic anomaly data from which Atwater and Mudie (1973) argued a half-width of dike injection from 0.72 to 2 km, whereas Larson and Spiess (1969) suggested 0.28 km. Support for a narrow axial injection zone half-width comes from observations of the inner floor of axial rift valleys; ARCYANA (1975) suggested a maximum of 1.25 km, and Moore and others (1974) indicated a maximum of 0.6 km from surface eruption half-width. The models of Cann (1974) and Moore and others (1974) involve an extremely narrow dike-injection zone, with flows spreading as much as 0.6 km from the axial singularity; this means, assuming isostatic compensation, that the pillow and sheeted complex accelerate very rapidly to the half–spreading rate, that rotation and thinning of flows must occur, that the thickness of the pillow complex is limited by the distance that lavas flow from the axial-injection zone, and that the contact between pillow and sheeted complex will be essentially nontransitional, with few or no dikes penetrating the pillow complex. How-

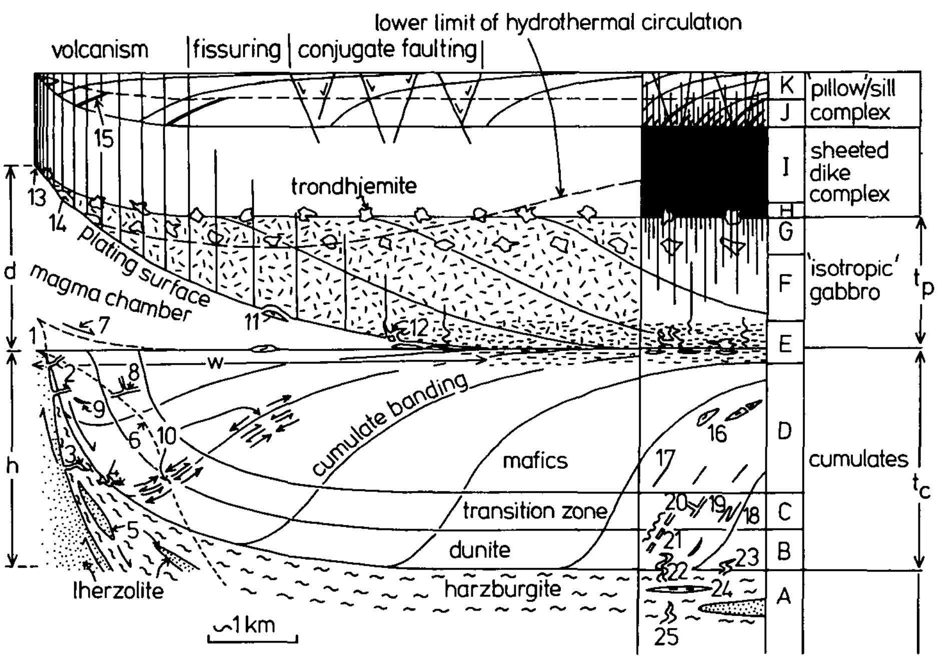

Figure 1. Proposed steady-state model for development of oceanic crust and upper mantle at accreting plate margin. No account is taken of topography, and surface of pillow complex is taken as level datum surface assuming perfect isostatic compensation.

ever, because perfect one-way chilling does not occur in the sheeted complex, because dikes increase in percentage downward in the pillow complex, and because the evidence suggests that individual flows do not travel more than 0.5 km from their source because systematic lateral chemical and petrologic changes occur in the FAMOUS area emissive zone (Ballard and others, 1975), we prefer dike injection and flow accumulation rate modes of exponential form where the zone of eruption roughly concides with the zone of dike injection (Fig. 2B), represented in the FAMOUS area (ARCYANA, 1975) by the axial mound.

Integration of the dike injection rate gives half-spreading velocity, and integration of flow accumulation rate gives pillow complex thickness (Fig. 2B) (Deffeyes, 1970; Anderson and Noltimier, 1973). Assuming perfect isostatic subsidence of the thickening pillow complex, there are three ways in which the subsidence may be accommodated. The simplest is by vertical faulting (Fig. 2D) either (1) on faults with the same slip rate with an exponentially increased spacing, or (2) on evenly spaced faults with a decreasing slip rate, or (3) by continuous slip at a decreasing rate on faults formed very

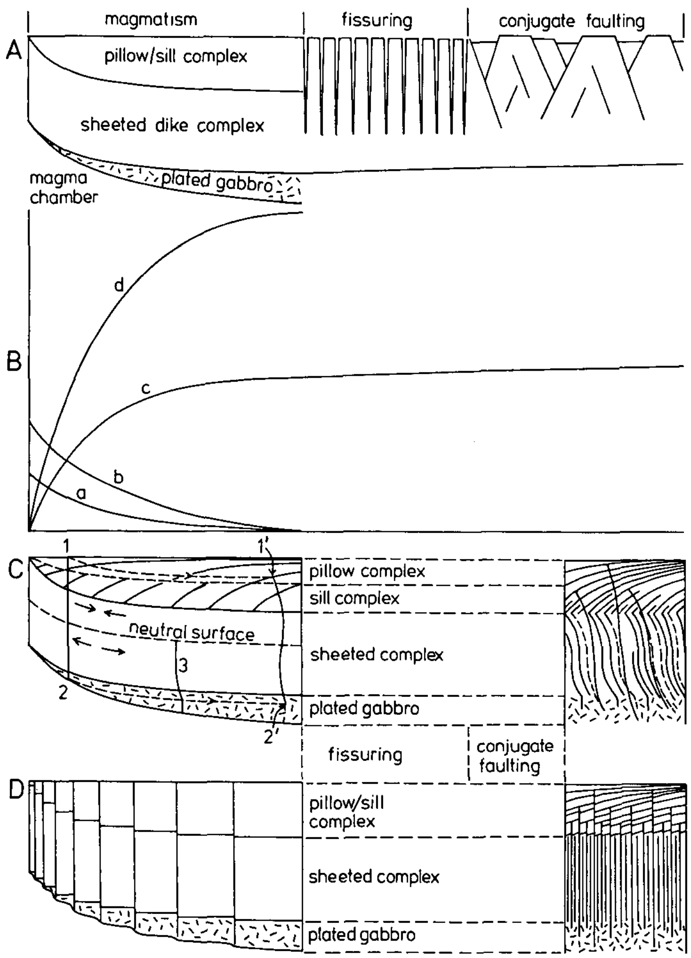

Figure 2. Models for accretion and thickening of magma chamber lid comprising pillow-sill complex, sheeted dike complex, and plated gabbro. A, section showing proposed thickening of lid by addition of pillow-sill complex and plated gabbro (followed by fissuring and conjugate faulting observed by ARCYANA, 1975, in FAMOUS area). B, a = dike injection (lateral extensional strain) rate; b = basalt accumulation rate; c = lateral velocity of lid relative to axial singularity (half-spreading rate — integrated dike injection rate); d = thickness of pillow-sill complex (integrated dike basalt accumulation rate). C, continuum acceleration of lid through thickening continuum bend fold in isostatic equilibrium. D, continuum acceleration of lid with isostatic equilibrium maintained by gravity faulting.

near the axial singularity. Conditions 1 and 2 would involve the development of underplated faults in the homogeneous gabbro and overplated faults in the pillow complex. Condition 3 would involve faults that penetrate the full thickness of the lid but have a decreasing vertical displacement upward and downward from the sheeted complex. Dikes injected away from the axial singularity are likely to follow these faults and show a spectrum of relationships, from intruding fault breccias to being deformed themselves. A more complex possibility, illustrated in Figure 2C, is a nonfaulted model whereby the sheeted complex accelerates through a thickening bend fold. Again, dikes will be progressively underplated and overplated and decrease in percentage away from the sheeted complex, but the bend fold will have a neutral surface of maximum shearing strain, above which compressional strains and below which extensional strains develop. This will lead to dikes developing sigmoidal forms (For example, dike 1-2 progressively displaced to 1′-2′, Fig. 2C), the final degree of sigmoidal formation depending upon the proximity of initial injection to the axial singularity. Parts of dikes that have undergone pronounced rotation have not been detected in ophiolite complexes but, as they will be difficult to detect, this is not at present a criterion for rejecting this hypothesis. Sub–neutral-surface extensional strain may enhance the injection of dikes that do not rise above the neutral surface (3 in Fig. 2C).

A further feature of both fault and bend fold modes may be the injection of sills (Fig. 2C; 15 in Fig. 1) low in the pillow complex, as suggested for self-perpetuating sheet swarms in Iceland (Walker, 1975). We regard the bend-fold hypothesis as unlikely for normal ridge segments because the rocks involved are in regional tension and under physical conditions that will lead to brittle, not ductile, deformation. If the lid is subject to fissuring and then conjugate faulting (Fig. 2A), as in the inner rift valley (gja and fracture modes; ARCYANA, 1975) the pillow complex, in particular, will be cut by a complex sequence of fissure and fault breccias, with the basal part of layer 1 sediments showing showing complex facies variations from pelagic sediment to talus breccias. The occasional off-axial mound volcanism recorded by ARCYANA (1975) enhances the view that the sides of the tapering magma chamber extend beyond the zone of axial dike injection and surface eruption and that further minor acceleration is accomplished by strain largely unaccompanied by magmatism beyond the main magmatic zone.

The width of the magma chamber can be estimated from cooling calculations giving estimates for the rate of deposition of the cumulates. Hess (1960) estimated rates of accumulation in the Stillwater layered mafic pluton from conductive cooling calculations. As it is likely that convective removal of heat by water from the oceanic lithosphere near the ridge axis exceeds, by an order of magnitude, his values for conductive heat removed, these estimates can be taken as minimum values for accumulation rates. He obtained accumulation rates of 5 cm/yr for ultramafic cumulates (slow) to 15 cm/yr for gabbro cumulates (fast), with 10 cm/yr as an average value. At these rates, he estimated the Stillwater Complex (8 km thick) to have solidified in not more than 10^5 yr. Applying this estimate to ridges spreading at 1 and 10 cm/yr half-rate, the magma chamber will extend 1 km and 10 km, respectively, from the ridge axis; this is in reasonable agreement with widths estimated on the basis of surface magmatism and topography.

Hydrothermal circulation in the lid (Fig. 1) is probably responsible for the convective cooling of the crust indicated by heat-flow studies (Anderson, 1972; Lister, 1972) and probably defines the thickness of the sheeted complex and, to a large extent, the rate at which gabbro is plated to the magma chamber roof. Hydrothermal circulation is also likely responsible for the extensive sulfide-zeolite-epidote-chlorite-calcite-quartz mineralization and (or) alteration of sheeted and pillow complexes (Gass and Smewing, 1973; Scott and others, 1974) and for the hydrous phases and alteration in the upper parts of the homogeneous gabbros. Mineralization of fractures will cause their progressive healing away from the axial singularity (Schreiber and Fox, 1976), which in turn is likely to cause the lower limit of hydrothermal circulation to progressively rise and may account (Fox and Delong, 1976) for the thinning of layer 2A away from ridge axes (Houtz and Ewing, 1964).

The maximum thickness (t_p) of plated gabbro is controlled by the axial depth (d) of the magma chamber. The dip(β) of successive freezing surfaces is controlled by the depth/width ratio of the magma chamber (Figs. 1, 3). The occasional wispy banding observed in the homogeneous upper ophiolite gabbros may be a 'plating band'. The shape of the magma chamber and attitude of the plating surface, possibly controlled by hydrothermal circulation and perhaps by spreading rate, control the shape and, therefore, the strength of the lid. If the lid is considered as a beam or cantilever with a fulcrum at the bottom edge of the magma chamber, it is supported partly by its own tensile strength and partly by underlying magma. The support by the magma is due to the flotation effect; as the magma must be somewhat less dense than the solid lid derived from it, this cannot completely support the lid. Overpressure in liquid magma is not a realistic mechanism for supporting the lid in the situation in which dike widths indicate extrusion and consequently pressure release every 1 to 50 yr. Therefore, for a lid generated above a magma chamber with a large depth to width (d/w) ratio, the lid is more likely to be capable of self-support and less likely to be faulted. Also, as illustrated in Figure 3 A and (A), homogeneous gabbro plated on a steeply inclined surface is less likely to be subsequently injected by dikes because the thick part of the lid is far less likely to break than the thin axial part; this will give rise to a rapid decrease in dike percentage below the sheeted complex and a very narrow axial injection zone. By contrast, a lid above a magma chamber with a small depth to width (d/w) ratio is less likely to be self-supporting, more likely to be faulted, may possess a wider axial injection zone, and may result in dike percentage in the homogeneous gabbro decreasing more gradually downward (Fig. 3, B and (B), C and (C).

It has been proposed that the magma chamber consists mostly or entirely of "mush," a self-supporting framework of crystals with 30% or less residual interstitial liquid (Sleep, 1975). This, if true, would provide an excellent method to support the lid. Extra support by some means is probably necessary because the magnitude of the bending stresses that can be calculated for the lid underlain by magma are very large and greatly exceed the predicted strength of the rocks. However, the existence of a significant thickness of cumulates requires the zone over which they formed to be magma; they would not exist if it were a self-supporting mush. Our model (Fig. 1) allows mush to exist beyond the plating and accumulation surface boundaries, but not inside them.

A possible consequence of slight rotation and subsidence of the lid is plastic deformation at the bottom edge of the magma chamber. An isolated zone of deformation in otherwise fairly homogeneous gabbro, therefore, might be an expression of the plated to cumulate boundary. The Betts Cove Ophiolite Complex shows a feature that probably resulted from this process; the base of the gabbro layer that contains dikes shows a folding of the dikes over a thickness of about 100 m from an attitude perpendicular and higher up to parallel and lower down, with the cumulate banding immediately below. Such folding of dikes may be common where lids have a large width/thickness ratio. It is possible that blocks may stope from the plating surface (11,12 in Fig. 1) and sink to become incorporated in the cumulates below, but we have not observed evidence (for example, gabbro xenoliths containing diabase dike segments) of this process in the ophiolite complexes with which we are familiar.

The origin of the trondhjemites, which are mainly restricted to the upper part of the gabbro and the lowest parts of the sheeted complex, is problematic, apart from the observation that they must form within the zone where active dike injection is proceeding. If crystallization is taking place on a cotectic, as is likely, the volume of liquid that gets to trondhjemitic compositions is extremely small. Furthermore, there is a problem in finding a real process to segregate this small volume of liquid from where it must form in residual pore space of the plated gabbro. Small areas of net-vein breccia can probably be accounted for by faulting and disturbance of the lid, thus remobilizing any trondhjemitic residual fluid, but larger bodies, such as those in Troodos, are difficult to explain. This difficulty is, however, general to all models so far proposed.

The oceanic crust must be finally consolidated by welding of the gabbro plating downward on the magma chamber roof to the cumulate gabbro below. The magma chamber near this point is likely to be thin and crystallization rapid, because it is the farthest point from the mantle magma supply axis. It is relatively difficult to identify rapidly deposited homogeneous gabbro cumulates on their own and, therefore, it is difficult to distinguish them from gabbro plated on the adjacent roof. The base of the plated roof gabbro, therefore, may be extremely hard to identify unequivocally in many ophiolite complexes.

If the shape of the lid and underlying magma chamber is dominantly controlled by the extent of hydrothermal circulation, wider axial injection zones are likely to characterize parts of ridges where circulation is more restricted and conductive cooling is more important. The wider axial injection zone (~50 km) in Iceland (Daignieres and others, 1975) may result in this way above a magma chamber with a small d/w ratio, because pervasive hydrothermal circulation is only very locally developed as the axis is above sea level.

Assuming a steady-state model, there are two ways of building a cumulate sequence. Cann's (1974) and Greenbaum's (1972) solution involves an axial zone of basalt discharge at the base of a wedge-shaped magma chamber from which the cumulate surface rises to the edge of the magma chamber. In such a model, the thickness of cumulates is governed solely by the difference in elevation of the cumulate surface between axis and edge of the magma chamber, the angle between cumulate banding and gross lithologic

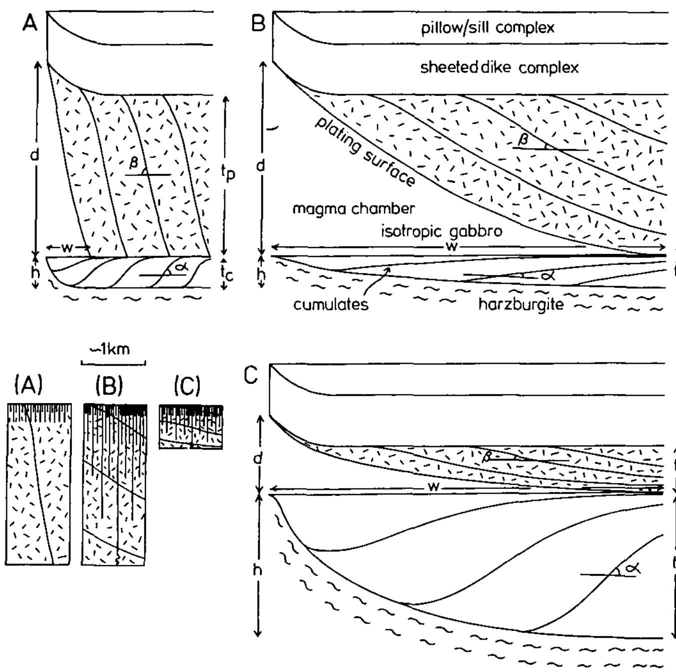

Figure 3. A, B, C, schematic illustration of dependence of thickness of plated gabbro (tp) and cumulates (tc) on magma chamber depth (d), height of axial lherzolite welt (h), and width of magma chamber (w). Sections (A), (B), and (C) show proposed dike distribution in plated gabbros of A, B, and C, respectively.

layering is the angle of rest of the cumulate surface, thickness of cumulates is the product of magma chamber width and the tangent of angle of cumulate rest, and, to account for vertical petrologic variation from dunite to gabbro, the magma chamber must be laterally zoned. For a magma chamber with a half-width of 5 km and cumulate angle of rest of 5°, the thickness of cumulates would be 0.44 km. In the Bay of Islands Complex, there are at least 2.5 km of cumulates; with a tolerable cumulate rest angle of 5°, the magma chamber would have a half-width of 28 km; for likely half-widths of ~5 km, the cumulate angle of rest would be an impossible 26.5°.

Sleep's (1975) model obviates these geometric difficulties by invoking a flat magma chamber floor and steady-state differential subsidence of that floor. We also adopt this geometry in our model (Fig. 1), whose corollaries are as follows. Basaltic magma must be discharged to feed the magma chamber from a very narrow axial zone to account for the sharp cumulate-harzburgite surface and the general lack of mafic dikes in the cumulates. This implies a tapering rising diapiric welt of partially melted lherzolite from which residual harzburgite is plated under the base of the subsiding cumulates. The thickness of cumulates (t_c) measured perpendicular to gross lithologic layering is controlled by the subsidence trajectory of the base of the cumulates away from the axial diapir and the height (h) of the axial diapiric welt (Fig. 3). However, a maximum value of t_c is placed by the basal width (w) of the magma chamber (Fig. 3), unless gross expansion of the volume of the cumulate pile is allowed after deposition, which is unlikely. Differential subsidence is a necessity of the model. The subsidence rate, as shown in Figure 1, decreases from a maximum near the axial welt to zero at the edge of the magma chamber; if a different shape is assumed for the welt, it might increase to a maximum and then decrease to a minimum, but we consider this possibility less likely. The subsidence causes a progressive bending deformation and rotation of cumulate banding, so that final banding orientation in the consolidated mature cumulate sequence describes a sigmoidal form relative to gross lithologic layering (Figs. 1, 3); the degree of rotation depends on the cumulate subsidence trajectory, which is controlled by the shape and therefore the rate of subsidence of the axial welt harzburgites. We believe that the variation of cumulate banding orientation in the Bay of Islands Complex (Smith, 1958; Williams and Malpas, 1972) resulted from such rotation. It is important to note that the overall cumulate section exposed in the Bay of Islands Complex is at right angles to the general orientation of dikes in the sheeted complex and therefore to the ridge axis that generated it, as in Figure 1, and can therefore be expected to show this phenomenon if our model is correct. Other ophiolite complexes may have sections of cumulates overall parallel to general dike orientation; in this case, false inferences about the dip and therefore the thickness of the gross layering may be made by extrapolation from the dip of individual layers seen in outcrop. In our model, as in those of Cann (1974) and Greenbaum (1972), individual horizons deposited at the same time rise from the base to the top of the cumulates, thus changing composition from dunite to gabbro. This means that the magma chamber, or at least cumulate deposition, must be laterally zoned accordingly.

Further possible consequences of this mode of cumulate development are as follows. An increased pulse of axial diapiric rise would produce an axial ridge on the magma chamber floor with outward slopes (1 in Fig. 1). This would produce slumping and outward flow of cumulates, giving rise to slump- and flow-type folds (7 in Fig. 1) and perhaps a flow lineation parallel with the spreading direction; if the increased pulse were persistent, a new higher level cumulate floor would be established, leading to thicker cumulates and thinner plated gabbro. Conversely, a sinking of the welt would lead to the same effects but in the opposite sense. A variety of deformation structures are seen in zones in the banded cumulate ultramafics and gabbros. The commonest is a gneissic foliation parallel to cumulate banding, varieties also including a strong lineation. This fabric is not a result of the weak grain alignments produced by cumulate deposition; cumulate grains are deformed and polygonized, and examples, where this fabric is slightly oblique to banding or axial surface to flow-type folds of banding, demonstrate that the deformation postdates accumulation. This deformation can be shown to be a ridge axis deformation from folds locally crosscut by and from xenoliths of foliated gabbro contained in undeformed gabbro. Other deformations seen include a banded gabbro–anorthosite dike in the cumulates of the Mings Bight Complex that probably originated by bulk remobilization of a layer with residual interstitial liquid through overlying consolidated cumulate layers, a process also inferred by Hopson (1975) in the Point Sal Complex. Large-scale boudinage of cumulate layers is well displayed in the Betts Cove Complex, where boudins of websterite and wehrlite layers can be traced within cumulate gabbro.

Underplating of harzburgite may occasionally lead to the entrapment of lherzolite lenses in various stages of partial melting, depending upon the depth at which they are entrapped (5 in Fig. 1), varying from primitive lherzolite or ariegite (Church, 1972), to gabbro lenses (24 in Fig. 1). These appear to be very rare and the extraction of melt from residuum to be extremely efficient. A variety of small localized rootless dikes, mainly websterite and dunite but occasionally gabbro, occur in the harzburgites and lower cumulates of the Bay of Islands Complex and may have originated from the partially melted lherzolite–paleted harzburgite interface, later becoming underplated by harzburgite (1, 3 in Fig. 1).

The complex, sometimes polyphase, high-temperature deformations observed in the harzburgites and those in zones in the lower cumulates of western Newfoundland and other ophiolite complexes are of essentially identical character, and we do not believe that a separate explanation for each is justified. We suggest that they mainly originate in a zone of high shearing strain induced between the rising axial diapir and the subsiding cumulates. Plated harzburgite, cumulates, and the rootless dikes would move progressively through a zone of high shearing strain in which deformation, filter pressing, and local dike injection overlap in a complex way. If differential cumulate subsidence is strong, bending deformations of cumulate layering may be important (10 in Fig. 1) giving rise to local intrados-extrados relations, which may account for the fact that in the Bay of Islands Complex, some layers show boudinage and others open buckle folds (21 in Fig. 1). Oscillation of the level of the top of the axial welt may add to these effects and will probably produce similar structures, by slumping processes, whose effects will be localized mainly at the base of the cumulates. Suggestions that the deformation structures in the ultramafic rocks are produced by shearing on the lithosphere low-velocity zone are, we suggest, inappropriate for the harzburgites, which must be plated near the ridge axis in a zone where relative transport of material is dominantly in a vertical sense. This dominance of vertical transport at the axis, combined with the observations that the deformation fabrics are always essentially parallel to the gross lithologic layering and that the deformation fabrics affect some of the cumulates, argues in favor of our model for the development of the cumulate and underlying rocks.

There are some problems that cannot be resolved from data available now. These are as follows: the relative thicknesses of plated gabbro, cumulate gabbro, and cumulate ultramafics; changes in these as a function of magma chamber width; the origin of large trondhjemite bodies; and correlations between spreading rate, magma chamber width, and thicknesses of the various units.

The amount and composition of magma supplied for a given increment of spreading on normal ridge segments appears to be fairly constant, because the total thickness of crust in normal oceanic situations is fairly consistently 6 to 7 km and because compositional variations of erupted magmas are small and insignificant compared with the variation at any one site. Taking the overall composition and amount of mafic material to be essentially constant, then the amount of ultramafic cumulates at the base of any section of oceanic crust can be changed only in one of two ways. One is to allow the gabbro to be overall more (or less) mafic; the other is to change significantly the combined thickness of the pillows and dikes. Differences in the overall composition of the gabbros in different ophiolite complexes will probably be very difficult to document unless they are gross — for example, if most is nearly anorthosite. In general, an overall less mafic gabbro will accompany a larger thickness of ultramafic cumulates and vice versa. A significantly smaller total thickness of pillows and dikes allows a larger thickness of plutonic rocks, leading to some increase in the thickness of both the ultramafic cumulates and the total (cumulate and plated) gabbro. There are data suggesting that layer 2 (dikes and pillows, broadly speaking) is thinner for fast-spreading ridges (about 1 km instead of 2 km). This should result in a somewhat thicker ultramafic cumulate layer, although the increase, for this reason, might be at most 500 m, which is probably not enough to explain completely the thick high-velocity layer 3b observed in the Pacific but apparently not observed in crust formed at slow-spreading ridges. Given a fixed total cumulate thickness and a fixed thickness of dikes and pillows, different widths of magma chamber may lead to different thicknesses of ultramafic cumulates if the relative temperature gradients in the chambers are different. Different *relative* subsidence curves for the axial welt under magma chambers of different widths will also produce different thicknesses of ultramafic cumulates. Whether such differences exist and how they correlate with magma chamber width are unknown and not yet predictable, but they might be used to account for variations in thickness of the high-velocity lower crustal layer 3b.

The final thickness of plated gabbro will be controlled by the amount of time the lid is open to cooling by hydrothermal convection; this is a function of magma chamber width and spreading rate. A correlation between these factors is unknown at present, and the likely total thickness of plated gabbro for a given chamber width is therefore not yet predictable. Hydrothermal circulation is likely to become less effective as a cooling mechanism as the thickness of the lid increases away from the axis, because existing cracks in rocks are essentially closed at 1.5 kb, equivalent to a depth about 1 km below the base of the sheeted dikes, because new fractures are likely to be more widely spaced as the lid thickens, and because old fractures will tend to be closed by mineralization as they travel out from their point of formation. The shape chosen for the gabbro plating surface is convex downward for these reasons, even though a convex-upward (gothic arch) shape would give a mechanically stronger lid. Dynamically generated tensional fractures are not restricted to depths above a level corresponding to a pressure of about 1.5 kb and will propagate down to where plastic or viscous behavior closes them; this depth, at any point, is likely to be very close to or at the upper boundary of the magma chamber, particularly since the fractures will be used by circulating water. Therefore, the ultimate thickness of plated gabbro will also depend on the width of active tensional fissuring compared to the width of the magma chamber, which is also unknown. It appears that the difference between observed and theoretical heat flow near the axes of fast- and slow-spreading ridges is very similar; assuming the magma chamber time widths are similar, there should be only minor differences in final plated gabbro thickness.

The model, as illustrated in Figure 1, has been deliberately drawn with approximately equal thicknesses of plated and cumulate gabbro. Their combined thickness is shown in the same proportion to the combined thickness of dikes and pillow lava as oceanic layer 3 (broadly speaking, gabbro) to oceanic layer 2 (broadly speaking, dikes and pillows). This has been done because the thicknesses and any differences in thickness of plated and cumulate gabbro in different ophiolite complexes are so poorly known. The model is easily modified to accommodate any necessary changes in the relative thicknesses of plated and cumulate gabbro when these are defined from study of thick, well-preserved ophiolite complexes.

If the final thickness of the plated gabbro is very small (a few hundred metres), which cannot be ruled out from present data, the magma chamber will have a very small depth and a small depth/width ratio, and the cumulates will be extremely thick. Temperature zonation of the magma chamber will be most easily achieved in this situation. Also, the possibility exists, in this case, for quite large variations of magma chamber width with time, due to variations in efficiency of hydrothermally controlled cooling acting on a relatively small volume of magma and consequent variations in the rate of crystallization at and near the edges of the chamber.

Although small volumes of trondhjemite can be made in the interstitial liquid in nearly solid plated gabbro, large bodies, especially those made near the axial singularity, are problematic. The largest bodies of trondhjemitic rocks known in ophiolite complexes, particularly those of Troodos, are sufficiently large to imply that the volume of their parent magma was as large as the magma chamber, and therefore that it froze solid to produce them. Such interruptions to reasonably steady-state spreading might occur through ridge axis jumping by moderate amounts (a few kilometres) and are not a concern in our steady-state model.

There remains the problem of whether the magma chamber under fast-spreading ridge crests is wider or narrower than that under slow-spreading rifted crests, and also whether the time widths are significantly different from one to the other. Conductive heat flow calculations obviously lead to a wider chamber under fast-spreading axes, but with the same time width as the narrower chamber resulting under slow-spreading axes. The more obviously fractured nature of slow-spreading ridges might be taken to imply that the hydrothermal convection will be more active in those ridges and will therefore tend to make the chamber even narrower. However, heat-flow studies suggest that convective heat removal is equally important at and near the axes of both fast- and slow-spreading axes, so the widths of chambers under both will be reduced from their conductive values; this still implies a wider chamber under fast-spreading ridges, but it is not predictable, from present data, whether one has a larger time width than the other. In contrast to this analysis, which strongly suggests a wider chamber under faster spreading axes, the existence of rifted axes to slow-spreading ridges as distinct from nonrifted axes for fast-spreading centers suggests, if this phenomenon relates in any way to the strength of the lid, that the lid is weaker on slow-spreading ridges. This implies that the magma chamber will be wider under slow-spreading axes than under fast-spreading axes, because the strength of the lid depends on its thickness and length of overhang, given the same densities of rocks and magma in each case. Because the thermal argument is more convincing and because other reasons have been proposed for the existence of rifted axes, it is probable that the phenomenon of rifted crests to slow-spreading ridges has little connection with the strength of the lid to magma chamber. However, the stresses that can be estimated to be acting on the overhanging lid of a wide magma chamber are enormous and not to be neglected, so we suggest that the magma chamber must be very narrow under both fast- and slow-spreading ridges. This suggestion is reasonably plausible from consideration of the estimates of cumulate deposition rates made by Hess (1960). Five kilometres of gabbro and ultramafic cumulates can easily be accumulated in 10^5 yr; this corresponds to a magma chamber half-width of 1 and 10 km, respectively, for spreading rates of 1 and 10 cm/yr. For a magma chamber time width of 10^4 yr, which is still plausible, the actual widths for the same two spreading rates are 0.1 and 1 km, respectively. Although this is probably somewhat beyond the lower limit for the time width of the chamber, it need not be very much wider even for a fast-spreading ridge.

CONCLUSIONS

The steady-state model for plate accretion described above has been developed mainly from petrologic and structural constraints observed in western Newfoundland ophiolite complexes. The model predicts a number of structural consequences that, in the context of the overall geometric consequences of the model, may be tested by detailed fine-scale mapping of the strucural characteristics of ophiolite complexes.

The more important components of our suggested model are the following.

1. Pillow lavas progressively accumulate, in any one section, from essentially zero thickness at the axis to full thickness 1 or 2 km from the axis. The rate of accumulation may be found by studies of the change in percentage of dikes at different levels in the pillow lava unit.

2. The thickness of 100% sheeted dikes (typically 1 km) is controlled by the depth of hydrothermal circulation at the axial singularity.

3. At least 300 m of gabbro is plated, starting at the axial singularity, onto the roof of sheeted dikes in a complementary, progressive manner to the lavas above. As this gabbro is plated, it must be rapidly cooled by the hydrothermal circulation and, at least near the axial singularity, cut by tensional cracks that are filled by dikes. The total thickness of plated gabbro is unknown and is likely to be difficult to define; however, the maximum depth to which diabase dikes that do not cut the cumulates penetrate the gabbro gives a minimum thickness.

4. The magma chamber is zoned, with dominantly olivine cumulates being deposited near the magma supply axis, grading out to dominantly clinopyroxene-rich cumulates that in turn pass into clinopyroxene-plagioclase cumulates, which continue to the edges of the magma chamber. Magma-chamber zonation is strongly suggested by the spatial variation of the composition of erupted magmas (Ballard and others, 1975). The fact that all well-preserved ophiolite complexes consistently show the vertical sequence of gross layering required by this zonation indicates that the thermally controlled system and the environment producing it are more or less steady-state and that *rapid* convective overturn is in some way inhibited. The relative lack of current-bedded and wash-out structures in the banded cumulates supports this view. If the supply of new magma to the chamber is roughly balanced, on a short time scale (100 yr), by eruptive removal and the volume created by spreading, the new hot magma should rise directly to the apex of the chamber and should not grossly disturb the somewhat cooler magma nearer the sides. The fact that the zonation exists is a strong argument for a fairly continuous supply of new magma; this is the only reasonable mechanism to keep the central zone in the magma chamber hotter than the sides when the whole roof is being cooled, mainly by hydrothermal circulation.

5. The modal compositions and relative volumes of the plated and cumulate rocks show that the phases involved in crystal fractionation are dominantly clinopyroxene and plagioclase; olivine is dominant near the axis but rapidly becomes negligible away from the axis and is, overall, a subordinate phase. Orthopyroxene is always a very minor phase; opaque minerals are conspicuously absent from all the rocks, except for chromite in the dunite. Because fresh magma is being supplied fairly continuously to the chamber, fractionation models must allow for some mixing between this and fractionating magma.

6. A flat-floored magma chamber is required because of the large thickness of cumulate rocks and the need to form them within a reasonable distance from the axis. To do this with an essentially steady-state process requires that there be a very narrow axial welt consisting of a mush of residual crystals and melt at its center and plated residual harzburgite thickening downward and sideways at its sides. The central part of the welt will be about 70%:30% residuum:melt at the depth where harzburgite plating starts and will progressively change upward to near 100% melt at the magma chamber floor. The rapid subsidence of any point on the welt is, we suggest, controlled by isostatic response to harzburgite plating and cumulate deposition and to the balance between melt extraction and harzburgite plating in a zone that gradually widens with depth to a width comparable to that of the magma chamber; thermal changes are suggested to be unimportant as a control.

The precise shape of the axial welt, defined by the cumulate ultramafic–plated (noncumulate) harzburgite boundary, is at present uncontrolled. Studies of the attitude of cumulate banding to overall gross lithologic layering will test this part of the model and give data on the shape of the welt. The shape we have chosen gives a maximum amount of room to accommodate the cumulate pile and the zonation required in the magma chamber.

7. The magma chamber, above the section in which cumulates are forming, must be filled with magma (liquid + crystals), not mush (framework of crystals with interstitial liquid). Mush may exist only below the cumulate deposition surface and above or outside the roof gabbro plating surface.

8. The penetrative deformation seen to affect the harzburgite and zones in the cumulate ultramafics and lower gabbro is, we suggest, not due to horizontal shear by lithosphere plate movement on the asthenosphere, and particularly not to horizontal shear between asthenosphere diverging sideways under the ridge axis more quickly than the lithosphere, as has been implied in several previous treatments. It is evident that material transport in the mantle at shallow depths near the axis under the magma chamber must be dominantly vertical to compensate for the gap that would otherwise develop between the two diverging plate edges. Therefore, we suggest that the penetrative deformation is due mainly to a vertical shear stress between the rising mush of residual crystals–basalt melt and the subsiding plated harzburgite–lower cumulates; some deformation in the cumulates may result from slump or flow processes caused by a rise or fall in the level of the top of the axial welt and to bending stresses in the differentially subsiding cumulate pile. Although the shear stress is vertical, the resulting structures end up near horizontal because of the geometry of the subsiding welt and cumulate pile, and near parallel with cumulate and other banding because merely a small shear strain will rotate these markers into near parallelism with foliation and lineation. Episodic rise or fall in the level of the axial welt requires episodically reduced or accelerated shear strain below; this mechanism may be responsible for zones of more strongly and more weakly deformed harzburgites.

ACKNOWLEDGMENTS

This work was supported by grants from the National Science Foundation.

REFERENCES CITED

Anderson, R. N., 1972, Petrologic significance of low heat flow on the flanks of slow-spreading midocean ridges: Geol. Soc. America Bull., v. 83, p. 2947–2956.

Anderson, R. N., and Noltimier, H. C., 1973, A model for the horst and graben structure of midocean ridge crests based upon spreading velocity and basalt delivery to the oceanic crust: Royal Astron. Soc. Geophys. Jour., v. 19, p. 137–147.

Arcyana, 1975, Transform fault and rift valley from bathyscath and diving saucer: Science, v. 190, p. 108–116.

Atwater, T., and Mudie, J. D., 1973, Detailed near-bottom geophysical study of the Gorda Rise: Jour. Geophys. Research, v. 78, p. 8665–8686.

Ballard, R. D., Bryan, W. B., Heirtzler, J. R., Keller, G., Moore, J. G., and van Andel, T. J., 1975, Manned submersible observations in the FA-

MOUS area: Mid-Atlantic Ridge: Science, v. 190, p. 103–107.
Cann, J. R., 1974, A model for oceanic crustal structure developed: Royal Aston. Soc. Geophys. Jour., v. 39, p. 169–187.
Church, W. R., 1972, Ophiolite: Its definition, origin as oceanic crust, and mode of emplacement in orogenic belts, with special reference to the Appalachians: Canada Dept. Energy, Mines and Resources, Earth Physics Br. Pub. 42, p. 71–85.
Church, W. R., and Riccio, L., 1974, The sheeted dike layer of the Betts Cove Ophiolite Complex does not represent spreading: Discussion: Canadian Jour. Earth Sci., v. 11, p. 1499–1502.
Colemen, R. G., 1971, Plate tectonic emplacement of upper mantle peridotites along continental edges: Jour. Geophys. Research, v. 76, p. 1212–1222.
Daignieres, M., Courtillot, V., Bayer, R., and Tapponier, P., 1975, A model for the evolution of the axial zone of mid-ocean ridges as suggested by Icelandic tectonics: Earth and Planetary Sci. Letters, v. 26, p. 222–232.
Deffeyes, K. S., 1970, The axial valley: A steady state feature of the terrain, *in* Megatectonics of continents and oceans: Rutgers, N.J., Rutgers Univ. Press, p. 194–222.
Dewey, J. F., and Bird, J. M., 1971, Origin and emplacement of the ophiolite suite: Appalachian ophiolites in Newfoundland: Jour. Geophys. Research, v. 76, p. 3179–3206.
Dewey, J. F., Fox, P. J., and Kidd, W.S.F., 1973, Structure and generation of oceanic crust and mantle: Geol. Soc. America Abs. with Programs, v. 5, p. 597–598.
Fox, P. J., and DeLong, S. E., 1976, Healing of oceanic layer 2A: Geophysical and geological consequences: EOS (Am. Geophys. Union Trans.), v. 57, p. 329.
Gass, I. G., and Smewing, J. D., 1973, Intrusion, extrusion and metamorphism at constructive margins: Evidence from the Troodos Massif, Cyprus: Nature, v. 242, p. 26–29.
Greenbaum, D., 1972, Magmatic processes at ocean ridges, evidence from the Troodos Massif, Cyprus: Nature, v. 238, p. 18–21.
Hess, H. H., 1960, Stillwater Igneous Complex, Montana: A quantitative mineralogical study: Geol. Soc. America Mem. 80, 230 p.
Hopson, C. A., 1975, Features of igneous cumulates in ophiolites formed at spreading ocean ridges [abs.]: EOS (Am. Geophys. Union Trans.), v. 56, p. 1079.
Houtz, R. E., and Ewing, J. I., 1964, Sedimentary velocities of the western North Atlantic margin: Seismol. Soc. America Bull., v. 54, 867–895.
Jackson, E. D., Green, H. W., II, and Moores, E. M., 1975, The Vourinos Ophiolite, Greece: Cyclic units of lineated cumulates overlying harzburgite tectonite: Geol. Soc. America Bull., v. 86, p. 390–398.
Kidd, R.G.W., and Cann, J. R., 1974, Chilling statistics indicate an ocean-floor spreading origin for the Troodos Complex, Cyprus: Earth and Planetary Sci. Letters, v. 24, p. 151–155.
Larson, R. L., and Spiess, F. N., 1969, East Pacific Rise crest: A near-bottom geophysical profile: Science, v. 163, p. 68–71.
Lister, C.R.B., 1972, On the thermal balance of a midocean ridge: Royal Astron. Soc. Geophys. Jour., v. 26, p. 515–537.
Moore, J. G., Fleming, H. S., and Phillips, J. D., 1974, Preliminary model for extrusion and rifting at the axis of the Mid-Atlantic Ridge, 36°48′ North: Geology, v. 2, p. 437–440.
Moores, E. M., 1969, Petrology and structure of the Vourinos Ophiolite Complex of northern Greece: Geol. Soc. America Spec. Paper 118, 74 p.
Moores, E. M., and Jackson, E. D., 1974, Ophiolites and oceanic crust: Nature, v. 250, p. 136–138.
Moores, E. M., and Vine, J. F., 1971, The Troodos Massif, Cyprus, and other ophiolites as oceanic crust: Evaluation and implications: Royal Soc. London Philos. Trans., ser. A., v. 268, p. 443–466.
Norman, R. E., and Strong, D. F., 1975, The geology and geochemistry of ophiolitic rocks exposed at Mings Bight, Newfoundland: Canadian Jour. Earth Sci., v. 12, p. 777–797.
Penrose field conference participants, 1972, Ophiolites: Geotimes, v. 17, no. 12, p. 24–25.
Schreiber, E., and Fox, P. J., 1976, Compressional wave velocities and mineralogy of fresh basalts from the FAMOUS area and the Oceanographer Fracure Zone and the texture of layer 2A of the oceanic crust: Jour. Geophys. Research v. 81, p. 4071–4076.
Scott, R. B., Rona, P. A., McGregor, B. A., and Scott, M. R., 1974, The TAG hydrothermal field: Nature, v. 251, p. 301–302.
Sleep, N. H., 1975, Formation of oceanic crust: Some thermal constraints: Jour. Geophys. Research, v. 80, p. 4037 4042.
Smith, C. H., 1958, Bay of Islands Igneous Complex, western Newfoundland: Canada Geol. Survey Mem. 290, 132 p.
Upadhuay, H. D., Dewey, J. F., and Neale, E.R.W., 1971, The Betts Cove Ophiolite Complex, Newfoundland: Appalachian oceanic crust and mantle: Canada Geol. Assoc. Proc., v. 24, p. 27–34.
Walker, G.P.L., 1975, Intrusive sheet swarms and the identity of Crustal Layer 3 in Iceland: Geol. Soc. London Jour., v. 131, p. 143–161.
Williams, H., 1973, Bay of Islands Map-Area, Newfoundland: Canada Geol. Survey Paper 72-34, 7 p.
Williams, H., and Malpas, J., 1972, Sheeted dikes and brecciated dike rocks within transported igneous complexes, Bay of Islands, western Newfoundland: Canadian Jour. Earth Sci., v. 9, p. 1216–1229.
Wilson, R.A.M., 1959, The geology of the Xeros-Troodos area: Cyprus Geol. Survey Mem. 1, 135 p.

MANUSCRIPT RECEIVED BY THE SOCIETY MAY 3, 1976
REVISED MANUSCRIPT RECEIVED JULY 19, 1976
MANUSCRIPT ACCEPTED SEPTEMBER 24, 1976