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# INTERLAYER MATERIAL TRANSPORT DURING LAYER-NORMAL SHORTENING. PART II. BOUDINAGE, PINCH-AND-SWELL AND MIGMATITE AT SØNDRE STRØMFJORD AIRPORT, WEST GREENLAND

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## ABSTRACT

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Slow deformation of rock at metamorphic conditions involves an interplay of ductile flow, fracturing and mass-transfer. Boudinage and pinch-and-swell structures from a Precambrian terrane in West Greenland illustrate this point. Macroscopic fractures between separating boudins are expected to be 0.1 mm or less in width. It is in such narrow openings, simultaneously dilating and being filled, that deposition from an oversaturated pore fluid leads to macroscopic differentiation.

The microscopic process thought to underly macroscopic mass-transfer is as follows. Pore fluid occurring throughout the deforming rock has a pressure lower than the normal stresses existing between grains. It will therefore be *oversaturated* compared to a hydrostatic equilibrium situation in which all solid–solid contacts experience normal stresses equal to the pore pressure. Local solution and deposition cause local rearrangements in the pore network and contribute to local ductile flow. Only heterogeneous deposition—in separate, mechanically controlled and dynamically reestablished groupings and alignments of sites—gives rise to macroscopic differentiation products.

The macroscopic theory developed in Part I and the microscopic view presented here are combined in an interpretation of some common features of boudinage and pinch-and-swell structures.

At high metamorphic grade the rocks are melted to some small degree. The principles governing differentiation are not affected by changing the pore fluid from a hydrous phase to melt. Veins in anatectic terranes grow as relatively competent deposits from a highly incompetent melt permeating the deforming rock.

## INTRODUCTION

"Boudinage structure" may be used to describe the situation in which rocks contain *separate* bodies aligned in a plane, provided it can be shown that the structure is formed by deformation of an initially continuous sheet of more or less

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constant thickness. The term entails a kinematic interpretation. Definitions with mechanistic connotations are less useful. "The competent layer is boudinaged" is circular as a field description because competency contrast is inferred from the boudinage structure itself. Historic accounts of the usage of the terms boudinage and boudin are given by Cloos (1947), Rast (1956) and Wilson (1961). Ironically, boudinage is mostly used to describe structures in which the foliation of the country rock is close to the plane of alignment of boudins, whereas Lohest coined the term in 1908 for structures in the Ardennes where foliation is at a high angle to the boudinaged layer (Brühl, 1969).

"Pinch-and-swell" is employed as a purely geometrical term indicating repeated gradual thickness variations in a *continuous* layer, whether this variation is of primary origin or caused by deformation. The definitions may be extended to cover the linear equivalents of the planar structures described here.

Theories for the development of boudinage and of deformation-induced pinch-and-swell analyse the following problem. A layer, or a set of parallel layers, is embedded in a medium of contrasting rheological properties, such that higher differential stress is needed for a certain strain rate in the layer material than that required to attain the same strain rate in the medium. The layer material is called competent and the medium material incompetent. The sequence is shortened normal to layering and extended parallel to it. A first condition in the theories is that the layer-parallel extension rate is the same throughout the medium and the layer initially. On the bulk scale this has to remain valid beyond the initial stages because of the requirements of strain-compatibility (Ramsay and Graham, 1970; Cobbold 1977). That there be no loss or gain in volume for layers or medium is the second condition in current analyses. Whether layer-parallel tensional stress arises such that boudins may be formed was analysed by Ramberg (1955), Strömgård (1973) and Lloyd and Ferguson (1981). Burg and Harris (1982) highlighted the role of shear failure in boudinage. Smith (1975, 1977, 1979) and Fullagar (1980) investigated whether prior to break-up small variations in thickness of the layer can amplify to produce macroscopic pinch-and-swell, and at what intervals for given thickness and assumed rheological contrast. The results of these instructive studies underline the crucial role of rheological properties and the importance of thickness ratios in multilayer models.

Field examples demonstrate that the condition for volume-constancy does not generally apply (Mullenax and Gray, 1984; this paper). It will be argued that transport of material has been essential for many cases of pinch-and-swell and boudinage to develop. The problem of periodic instability posed by these structures has been discussed competently by the authors quoted above. Here, our persistent incompetence is confronted with the hard fact of mass-transfer. The conditions of deposition in some layers during their extension are discussed and the conclusions give rise to an interpretation of boudinage and pinch-and-swell in the light of the differentiation model developed in part I.

The paper ends with a remark about syntectonic mass-transfer during partial melting as this is found to be relevant to many high-grade gneiss terranes. It is concluded that the presence of a small amount of melt, instead of a hydrous phase, does not alter the fundamental mechanism of differentiation.

## FIELD EXAMPLES

### *Setting*

The exposures chosen for this study are situated around the international airport at Søndre Strømfjord, lat. 67°N; long. 50°40′W, in West Greenland. The location is central to the amphibolite facies Ikertôq zone of the early Proterozoic Nagssugtoqidian mobile belt of West Greenland. The "Kangâmiut" swarm of parallel basic dykes (1950 m.y.; Kalsbeek et al., 1978) is found undeformed in Archaean terrane to the south of the area. Within the Ikertôq zone the same dykes have been deformed and metamorphosed to amphibolites during the Nagssugtoqidian orogeny (Escher et al., 1975, 1976; Rapp. Grønl. Geol. Unders., 89, 1979; Van der Molen, 1984). Mass-transfer phenomena are particularly well developed in the area. The greyish granodioritic gneiss contrasts sharply with dark amphibolite and with white or pink pegmatoid in veins and between boudins. Several of Ramberg's (1955, 1956) original observations on boudinage and on pegmatites were made in this area. His descriptions cannot be improved upon, they should be read in conjunction with the present paper.

At Søndre Strømfjord Airport, in the area of the U.S. military base, one may visit a large exposure between the dining hall of the airbase and the river directly east of it. Here twelve sub-vertical ENE-striking dykes are found in an almost continuous outcrop of 300 m normal- by 300 m parallel to strike. Wide dykes of 30 m show only incipient boudinage, or a remarkable internal metamorphic differentiation expressed by metre-long and a few centimetre-wide quartz-feldspar lenses parallel to the amphibolite layer surface. Thinner dykes of a few metres only are boudinaged into widely separated lensoid bodies, their separation indicating a minimum layer-parallel extension of 50%. It is argued that the dykes were initially parallel in view of their parallel occurrence in undeformed Archaean terrane to the south, and that they underwent similar extensions in excess of 50% because there is no gradient in boudinage-intensity when going perpendicular to strike from one dyke to another. Thus some dykes and the gneiss between the dykes did achieve their extension by ductile deformation, others elongated by boudinage as well during their deformation history. A description of one of the twelve dykes may serve to illustrate the interplay of rigid body movements, ductile deformation and mass-addition in achieving bulk extension of an amphibolite layer in gneiss.

*Incipient boudinage*

Compared to its present length, the dyke in Fig. 1 has been extended by 20–25% in a non-ductile fashion. This part of elongation may be considered in terms of componental movements normal and parallel to the boudin separation planes. Roughly half of it is taken up by sinistral shear displacement of the boudins with respect to each other resulting in dextral rotations relative to their enveloping surface. The other half of non-ductile extension consists of displacements normal to the boudin separation planes. Bulk ductile deformation of the dyke occurred before, after, and during boudinage as well. A well-developed amphibole foliation and isoclinal folds of thin leucocratic veins bear testimony to a plastic strain history. Foliation and fold limbs run into boudin terminations and they may be picked-up in the next boudin at corresponding levels. Variation in marker horizon spacing from boudin to boudin is evidence of bulk strain post-dating break-up and of spatial variation in it. In dykes at more evolved stages of separation individual boudins have

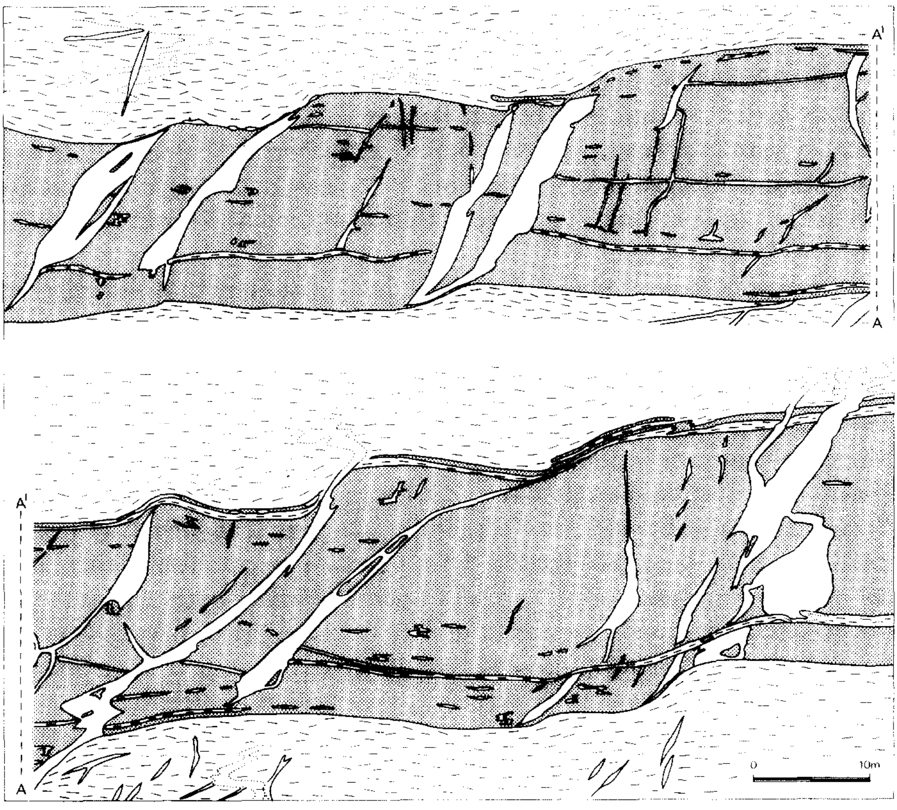


Fig. 1. Boudinaged basic dyke (dark) with pegmatite deposits (white) in gneiss (dashed). The boudins have been displaced in dilatant sinistral shear zones. Exposure between river and dining hall of U.S. airbase.

lost their blocky shape to a more lense-like profile (cf. Wegmann, 1932; Lloyd and Ferguson, 1981).

More localised ductile straining occurred during break-up as well. Against the dominant NE-striking pegmatites the strong amphibole foliation of boudins is deflected, slightly northwards against eastern terminations of boudins and slightly southwards against western terminations. The deflections comprise a change in strike of 10–20 degrees over zones of at the most 20 cm wide. A clear angular relation between the foliation and the amphibolite–pegmatite contact is everywhere maintained except at very pointed northeast and southwest corners of some boudins. The structure is best described as a sinistral shear zone with pegmatite in its center. The shear zones are related to the local incipient stage of boudinage because deflections are not notably better developed against wide pegmatites than against narrow ones. Shear zones without pegmatite have not been found. Only part of the sinistral displacement between boudins can be ascribed to localised shear within amphibolite. Clearly, sinistral movement of boudins relative to each other continued as they were being separated and while pegmatite grew in between.

Mass-transport occurred within and to the extending dyke. Transfer on the local, centimetre- to decimetre-scale is inferred from small foliation-parallel pods of plagioclase and quartz that are surrounded by dark rims of almost pure amphibole-rock. Narrow centimetre-wide crosscutting veins in amphibolite, including those in narrow sinistral shear zones, have the same composition as the pods as long as the vein does not intersect with gneiss. K-feldspar becomes a major constituent of the pegmatoid wherever veins contact the gneissic medium on either side of the dyke, or the thin gneiss layer within the dyke. This is the case for all larger pegmatite bodies. Equant decimetre- to metre-size crystals of K-feldspar occur in the widest pegmatite bodies, along with equally coarse plagioclase and quartz. Biotite is a minor constituent of pegmatite; it shows preference for the contacts against amphibolite, where it forms a conspicuous coat of pure mica. Gradients in gneiss- or amphibolite composition towards major pegmatite bodies cannot be recognised. In the case of Fig. 1, gneiss next to a larger pegmatite body is alike to gneiss more than 10 m away. Apparently there are different mass-transfer "cirquits" in the deforming rock sequence. The one within the dyke operated on a small scale. A much larger cirquit of different composition was "tapped" by veins connecting with gneiss.

Rigid body movements, bulk- and local ductile deformation and mass-transfer have been described separately. There can be little doubt that they operated simultaneously during much of the strain history, each contributing to the total extension of the dyke. The relative importance of these factors varied from dyke to dyke and possibly from moment to moment.

### *Other aspects of boudinage*

The space between boudins is almost totally filled by pegmatite in Fig. 1. Particularly in more advanced stages of boudinage the medium, gneiss, has moved into the gaps between separating boudins. Complete lack of pegmatite between boudins has not been found, but the amount of deposited material can be very small (Figs. 2A, 3A and B).

The spacing of crosscutting pegmatites is highly variable, from less than half the thickness of an amphibolite layer to more than ten times the present thickness, and such variation may be found in the same boudinaged layer. The often postulated regularity of the aspect ratio between the intermediate and the short axes in boudins does occur as well (Figs. 2 and 4). Fig. 1 illustrates that sharp distinctions between intrafolial boudinage, within the dyke, and boudinage dividing the dyke cannot always be made. It is felt that only fairly large deformations under relatively constant conditions can lead to regular patterns. Variation of the simplest of conditions with time or location may result in finite patterns of great irregularity (Mandelbroth, 1982).

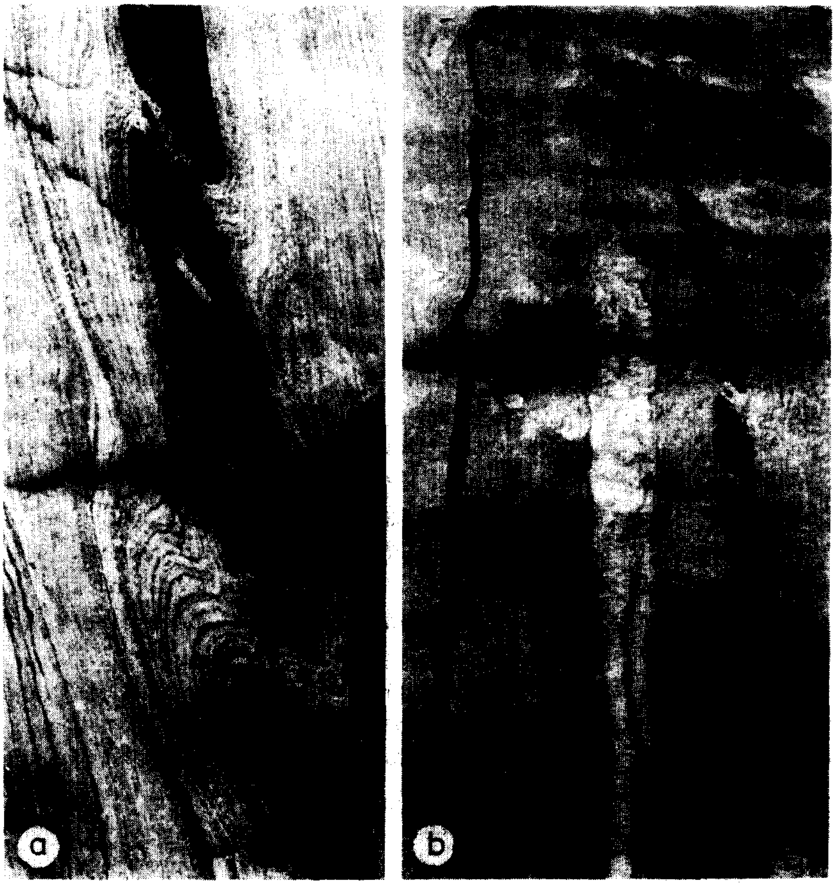


Fig. 2. a. Amphibolite boudins with gneissic medium in between. The folds die-out with distance from the boudin-terminations. b. Thin, isoclinally folded layer of amphibolite; low spacing of veins in one limb, no boudinage in the other. Exposure near the Inland Ice.

Dyke-cutting pegmatites in the area generally show the range in strikes seen in Fig. 1. Rotation of boudins relative to their plane of alignment varies in the area. Boudins with high aspect ratios are not rotated although some look as though they may have become realigned after initial shear displacements at their ends (Fig. 3A). At their terminations tight folds have been found looking much like those produced experimentally by Ramberg (1955) and theoretically by Lloyd and Ferguson (1981) (Figs. 3A, B). Fold packets associated with rotated boudins die out with distance into the gneiss (Figs. 2A, 3A). The rotation of boudins with respect to their plane of alignment has been studied by Ghosh and Ramberg (1976), in theory and in two-dimensional models. The ratio of pure shear and simple shear components in the strain history, the aspect ratio of boudins and the orientation of their axes relative to the direction of simple shear all influence rotation. Ghosh and Ramberg demonstrate that for combinations of these factors boudins may rotate faster or slower than their plane of alignment. The dykes of the Søndre Strømfjord Airport area clearly underwent non-coaxial strain histories. The relevant measurable quantities, aspect ratio and relative rotation have not been mapped systematically but the excellent exposures provide good opportunity for a follow-up in the field.

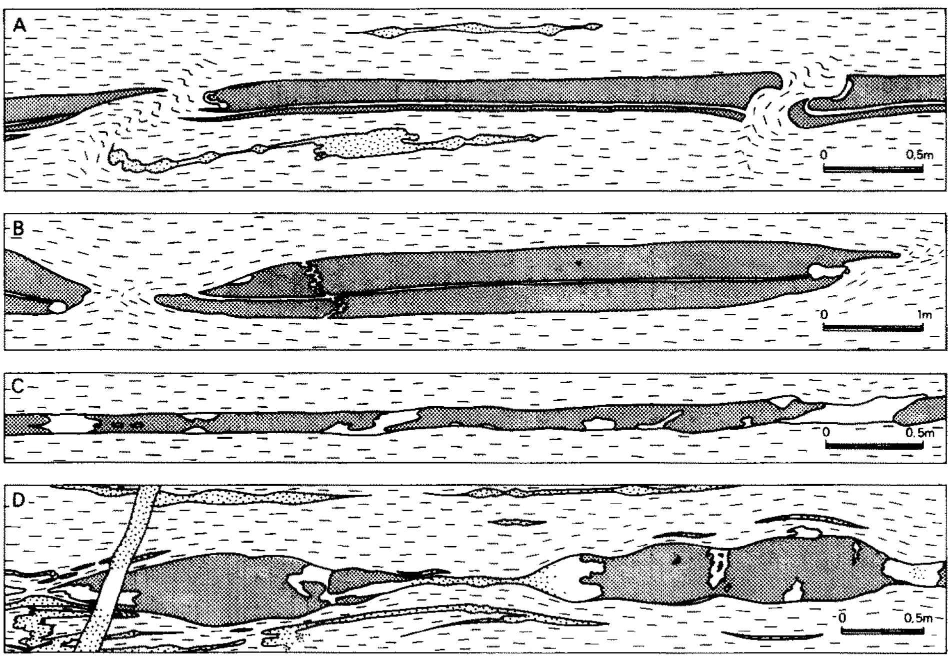


Fig. 3. A–D. Different forms of boudinage. Amphibolite (dark), plagioclase and quartz deposits (white), K-feldspar bearing pegmatite (stippled), gneiss (dashed). B. Folded vein indicative of initial boudinage followed by ductile flattening. D. Composite deposits between lense-shaped boudins. The K-feldspar bearing portions have grown later than K-feldspar-free pegmatoid on either side. Note host control on the composition of the late crosscutting vein at left.

Minor boudinaged layers of amphibolite are often coated by pure quartz and plagioclase pegmatoid on the long surfaces of boudins as well as in the spaces between them (Figs. 4A, B). Black fragments then seem to float in white layers against a background of grey biotite gneiss. These layers are remarkably planar, even when the separation of individual boudins amounts to ten times their present thickness (Figs. 4A–C). There is a tendency for proportionally more pegmatite and/or medium along the length of thin boudinaged layers than along the length of thicker ones. This is seen in a single exposure in Fig. 4B; the same phenomenon, but on a much larger scale, was described earlier for the twelve dykes having variable components of non-ductile extension.

It is not possible to establish exactly to what extent material for pegmatite has been derived from the basic layers themselves, which then must have had a composition different from the present boudins, and to what extent material removed from surrounding gneiss contributed. Fig. 3D illustrates both the problem

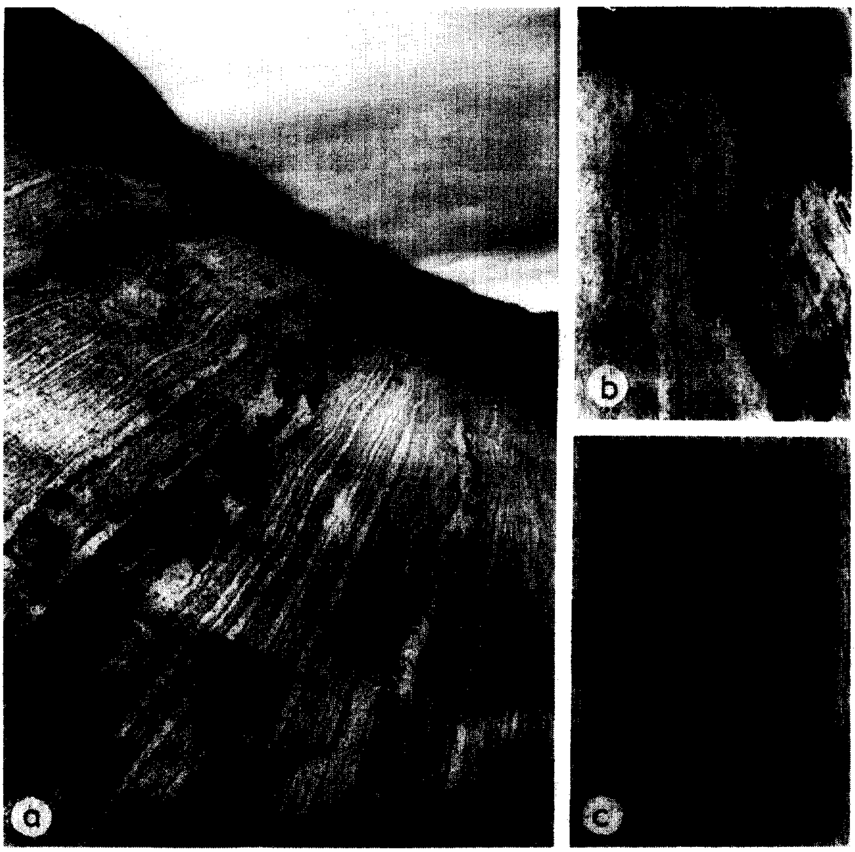


Fig. 4. A–C. Thin boudinaged layers of amphibolite in differentiated gneiss.

and the unlikelihood that all pegmatoid material results from redistribution within the extending amphibolite layer. The domains without K-feldspar may have obtained some material from the amphibolite nearby. However, the central K-feldspar bearing portion, with pinch-and-swell structure and with dark biotite selvages against gneiss, cannot have been derived solely from amphibolite. Lateral growth of the layer resulted in compositional zoning between separating boudins. The K-feldspar bearing portion was added later than parts on either side.

*Pinch-and-swell veins in gneiss*

Veins in the gneisses of the area are found straight, sigmoidal and folded, crosscutting and parallel to foliation, constant in thickness or with pinch-and-swell. The wide range in shapes and sizes indicates that veins were formed at all times during deformation and that growth of an individual vein could take place in a short interval compared to the total period of deformation. Straight veins at a high angle to foliation developed at the very end of straining (Fig. 3D). Folded veins grew earlier and were deformed without much further growth after being established. However, pegmatites concordant with the foliation continued their growth with further straining. This is the case whether they originally developed in- or rotated towards that orientation. Pegmatite volumes between amphibolite boudins grew by addition of material to an extending layer as was shown in previous sections. There is no reason to assume the growth history of otherwise identical parallel veins to be different (Fig. 3A, D).

A process can be envisaged in which planar pegmatites form when the supply rate of material keeps up with the rate of volume addition required for layer-parallel extension at constant thickness. Pinches develop when the supply rate is less than that; they are then thought of as material added more recently than the swells on either side. The most narrow portions in one part of the layer are not necessarily younger than less narrow pinches in another. This view highlights an aspect of pinch-and-swell formation which is left out by theories dealing with the amplification of periodic instabilities during layer extension at constant volume. In practice the lateral growth, ductile straining and non-ductile rigid body movements will go hand in hand as in the case of an extending amphibolite layer. Ramberg (1955, p. 520) wrote "The only difference is that, in the case of pinch-and-swell quartz or pegmatite veins, the boudins themselves consist of quartz or quartz and feldspar". In this light the origin of swells may seem more problematic than that of pinches: was the entire vein at some stage at least as thick as the thickest swell? This is not so. During any progressive deformation initially straight veins may go through a process of shortening followed by elongation to give rise to swells at the location of former fold-hinges (Fig. 5A–C). That this sequence operated in the Søndre Strømfjord area could be inferred but not demonstrated because of complete recrystallisation. An unusually well preserved example from similar Proterozoic amphibolite facies gneisses

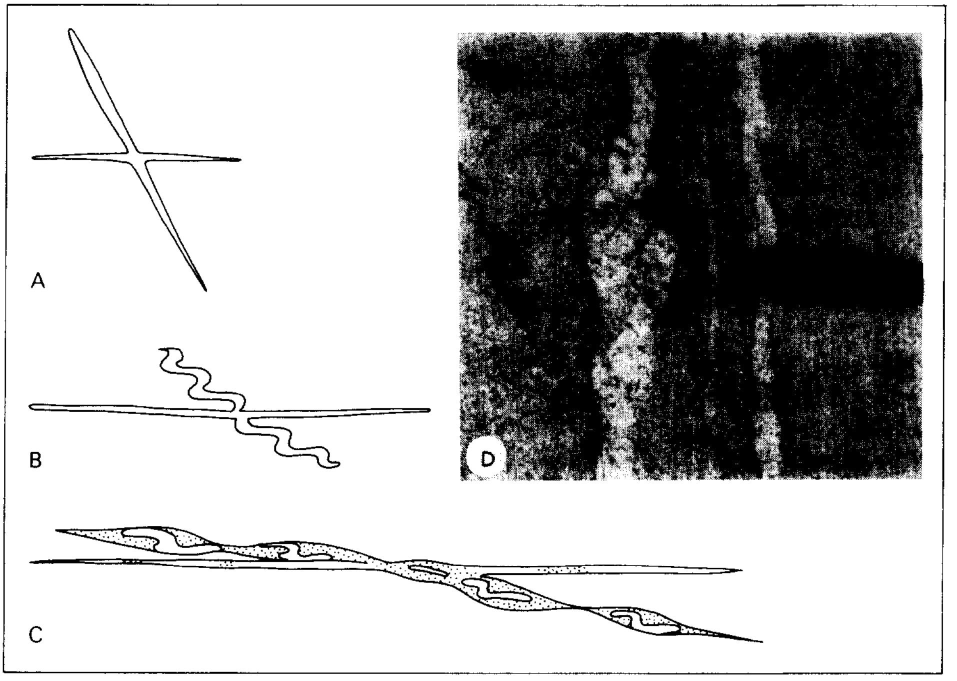


Fig. 5. A–C. The formation of pinch-and-swell veins during progressive deformation. Initial shortening leading to folds is followed by extension, mass addition and recrystallisation resulting in swells at the sites of former fold hinges. D. A swell *in statu nascendi*, from Orijärvi, southwest Finland.

in the Orijärvi area of southwest Finland is shown in Fig. 5D.

The composition of pegmatite veins and bodies in the area is as variable as their shape. In some continuous veins mineralogy depends on the rock type transsected (Fig. 3D), in others there is no apparent host control. Biotite-rich margins between gneiss and pegmatite may be marked or all evidence of material removal from adjacent gneiss can be absent. Pinches are sometimes different in composition (quartz) from swells (quartz, plagioclase and K-feldspar). This mineralogical variation again indicates that veins formed and continued to form under variable conditions during the deformation history. Changes in composition need not only reflect changes in pressure or temperature or the amount and type of fluid carrying dissolved species. Strain rate as well is a petrological factor determining the growth and composition of syntectonic veins. In this respect it also is worth noting that even at one instant differently oriented veins experience different rates of length change.

## SUMMARY AND DISCUSSION

The central point of the field descriptions is clear: certain layers in a deforming rock-mass gain volume and much of the new material is added in the direction of elongation of such layers. Boudinage and pinch-and-swell are interpreted accordingly. In the more competent material ductile straining alone cannot provide the total rate of length change imposed on the sequence as a whole. The layer breaks and, while the fragments continue to be deformed in a ductile manner, new material is deposited in the opening spaces between them. If new material can no longer be added at a sufficiently high rate, the system chooses from two options: opening gaps are filled with unmodified medium material, or the layer material is deformed ductilely at a higher rate. In the latter case boudinage does not evolve, and thin layer-crosscutting veins, grown during an incipient stage of boudinage become folded with continued layer-normal shortening (Fig. 3B). Abortive attempts at boudinage are by no means uncommon.

The physico-chemical constraints determining the instantaneous contribution of various possible mechanisms to deformation are discussed below.

*The mechanical situation allowing fracture opening*

We note, as many have done before us, that net tensional force across a contact deep in the earth can only be achieved at very high differential stresses, of the order of the overburden pressure, unless a pore fluid at pressure is present. Pore pressure reduces the normal force with which surfaces are pressed together to carry the lithostatic load. The moderately low differential stresses thought to apply to slow ductile deformations at depth do cause surfaces to separate. In a layered system being flattened, net tensional force arises in the most competent layer as soon as pore pressure becomes equal to the least normal stress in the direction parallel to layering. Any gradual increase in the strain rate imposed on the system could have that effect; an increase of the pore pressure is another possible cause.

Large numbers of aligned grain contacts and grains have to fail to produce the macroscopic disruption involved in boudinage. The strength of individual grain boundaries may be negligible both in tension and in shear but the interpenetrating geometry of the network of grain boundaries gives rise to a small finite strength of rock. It could be compared to strength of a jigsaw puzzle in which the saw cuts between pieces have zero strength also. The slowness of geologic deformation favours non-abrupt failure during boudinage. Some links along a potential macroscopic failure surface will open during increasing differential stress in the competent layer, many links will open when stress is high, and the longest lasting links break when the stress difference has dropped to post peak values. From then on the strength of the rock will be less, perhaps comparable to that of a weak grain-boundary, until recrystallisation and other adjustments reestablish the interpenetrating grain-boundary network. The boudins between fractures continue their ductile flow under influence of the difference between the highest compressive stress and the pore pressure.

An order of magnitude calculation may bring out the fracture widths to be expected. Let 1 MPa, or less, be a reasonable value for the long-term tensional strength of several metres of rock length. The minimum compressive stress, $\sigma_{xx}$, can now drop at the most 1 MPa below the pore pressure before surfaces with pore fluid in them commence to separate. During the slow opening of a macroscopic fracture the associated *increase* of $\sigma_{xx}$ within the boudins would be no more than 1 MPa to become equal to pore pressure. For reasonable Youngs Moduli between 10 and 100 GPa the resulting *elastic shortening* will be $10^{-5}$ to $10^{-4}$ at most. The spacing of macroscopic fractures between boudins varies from a few centimetres to several metres depending on the thickness of the layers. Taking 1 m to represent spacing in our calculation it is concluded that fracture widths are unlikely to exceed 10 $\mu$ to 0.1 mm. This width is at least on order of magnitude smaller than the average grain-size of the rock.

Boudin separation is caused by *extensional shear failure* rather than ordinary tensional failure in some of the cases described. Etheridge (1983) has shown that the possible magnitudes of the differential stress and of the difference between $\sigma_{xx}$ and pore pressure are only slightly increased by this change in character of the macroscopic fracture. Our order of magnitude calculation applies to both cases.

*A microscopic view of local deposition*

It seems unlikely that the observed heterogeneous deposition of material in narrow macroscopic fractures can be attributed to local, perhaps periodically re-established pressure drops of the pore fluid in and around the opening fractures. This is difficult for the following reasons. Pore pressure is needed to open the fracture in the first place, as discussed above. High viscosity and high incompressibility of the pore fluid in combination with low initial porosity and small permeability of the rock in the vicinity of the fracture would favour local pressure drops. But this would lead to removal of components from near the fracture as well. Some of these factors may well play a role in small-scale diffusion but we note that, in many of the cases described, the adduced material has been removed homogeneously from fairly large volumes of rock. An alternative interpretation is suggested.

In a "tight rock" (Fletcher, 1982) fluid is omnipresent in tubular pores at grain edges and in grain-boundary films but it is limited in quantity. There is no net flux of fluid through the system, at least there need not be. Short-term, short-range readjustments of the pore network may occur by slip or plastic deformation of grains or by migration of their boundaries during recrystallisation, etc. In the long run and on the macroscopic scale there is no coordinated movement of pore fluid (i.e. of porosity) from one area of the system to another. New pores are created as others disappear. Macroscopic mass-transfer is considered to take place gradually and over long periods in such a steady state interconnected pore network.

The relevant pressure of the pore fluid is the same throughout the layered system and less than —or equal to— the least compressive stress in the strongest layer (previous section). Pressure differences arising in the fluid phase during deformation will be randomly distributed and non-directional; they are not important for large scale mass-transfer.

Films and stringers of fluid are everywhere in contact with crystals that are pressed together with normal stresses exceeding pore pressure. *Relative to a hypothetical hydrostatic state in which all normal stresses between solids are equal to the pore pressure the fluid will be oversaturated in soluble components* (cf. Paterson, 1973; Robin, 1978; Green, 1980). Or, stated differently, there is an underpressure of the pore fluid relative to the normal stresses existing in the solid framework of deforming crystalline rock. This effect is a simple consequence of the differential stress causing deformation of the system. Importantly, relative underpressure of the fluid and oversaturation of soluble components will persist as long as the system is kept deforming.

A heterogeneous deposition from this pore fluid is not to be expected if normal stresses between grain boundaries remain compressive in all directions throughout the layered stack. There may well be a modest transfer of matter towards the competent layer but this leads to homogeneous deposition as there is no particular grouping of sites available for new crystals to nucleate and grow. The length-scale and the rate of mass-transfer increase once a macroscopic fracture develops as layer-parallel stress becomes equal to pore pressure. Now there is an alignment of sites where nucleation and growth can proceed. New material will be deposited as long as the narrow fracture is maintained; it does not have to open any further than initially. The rate of vein growth equals the rate of boudin separation.

The growth of heterogeneities is thus considered to be simply governed by a geometric and kinematic constraint: deposition occurs where surfaces move apart. There is no relation to pressure differences within the fluid. Note, however, that the localisation and orientation of macroscopic fractures are determined dynamically by the imposed rate of deformation and the flow and strength characteristics of the solids.

In the absence of localised deposition the concentration gradients of solutes in the pore fluid will be minimised. Depending on the length-scale pertaining and on the time available this may be achieved to degrees varying from complete mixing of components derived from different minerals and different layers to an almost site-bound composition of the pore fluid solution. When permanent sinks are established in the form of dilating and simultaneously filling fractures the concentration gradients will be continuously maintained, allowing mass-transfer to proceed and enabling the growth of macroscopic differentiation products.

### *Application to field examples*

On a macroscopic scale mass-transfer during deformation may be considered to be governed by differences in mean stress arising from rheological contrasts between layers (Part I). At the microscopic level transport depends on diffusion through pore fluid from sites where crystals are pressed together to sites were crystals are in free contact with pore fluid. The mechanically controlled alignment of the latter along fractures in layers of more competent rheology links the macroscopic and microscopic interpretations.

Several field observations are readily interpreted in terms of the macroscopic model. Mass transfer is found to be strongest, and boudins are found to be most separated, in extended thinner layers of amphibolite. This is in accordance with the model relating layer-parallel stress and the transfer of matter to the relative magnitudes of the characteristic diffusion length and the layer thickness (fig. 5 in Part I). Where deposition has occurred in the direction perpendicular to very thin layers as well as parallel to it (Fig. 4) this coating is seen as evidence of the discussed "stable layer-thickness" for differentiation to which the deforming system evolves. Apparently the thin amphibolite fragments have acted as nuclei to the differentiation packets growing in the gneiss (compare fig. 9 in Part I).

Insight can also be gained from switching back and forth between the macroscopic and microscopic views of mass transfer. The parameters "mobility" and "characteristic diffusion length" used in Part I have to be seen as *effective quantities* only. They are macroscopic expressions of many contributing factors such as the diffusivities of the various components dissolved in the pore fluid, the absence or presence of deposition sites, the degree of interconnection and tortuosity of the pore network, etc. Field examples in which effects of mass-transfer are concentrated at the contacts between contrasting materials (e.g., biotite-rich margins on pegmatite veins) are in accordance with the macroscopic model: the rate of mass-transfer should be highest in the vicinity of the contact in simple situations (figs. 5–8, Part I). However, it is common to find layers in an incipient stage of boudinage with lenses of deposited material in their centres and in more advanced stages the deposits are found to be wider there than at the margins (Figs. 1 and 6a). This may seem puzzling at first but it is recalled that the highest differential stress and the lowest layer-parallel stress are to be expected in the centres of layers when a diffusive exchange with

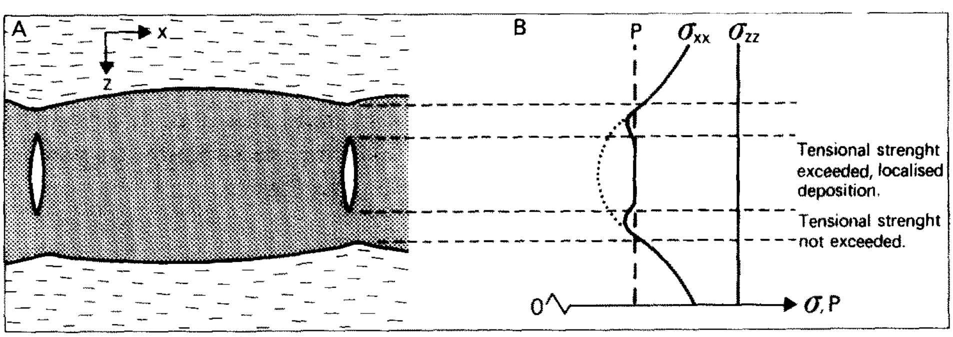


Fig. 6. A. Lensoid deposits in the centre of a competent layer. B. Dynamic interpretation.

the surrounding medium is possible (figs. 4–8, Part I). Tensional fractures form first, more often and stay longer within the layer. Thus layers are effectively divided into margins with relatively low mobilities and correspondingly low diffusion lengths and a central band in which these parameters are much increased (Fig. 6B). The result is that mass-transfer contributes more to extension within the layer than at the contacts with the medium. Part of the material deposited within such lenses may actually be derived from the layer material itself as argued previously.

*The nature of the pore fluid*

So far, in the microscopic model and in the applications presented, it has been tacitly assumed that pore fluid is hydrous. Many of the deposits found did grow from a hydrous phase with dissolved components. Lenses and veinlets of almost pure plagioclase in amphibolite and sizeable quantities of quartz in pinch-and-swell veins or in the cores of zoned pegmatites between boudins cannot have grown from melt (Ramberg, 1956). The temperatures which would be required for melts of such compositions are inconsistent with the grade of regional metamorphism. The upper amphibolite facies conditions and the more or less granitic composition of many other veins suggest, however, that partial melting of gneiss may have played a role in the formation of differentiation products. This conclusion applies to high-grade gneisses and migmatites in general (Vernon, 1976; Robin, 1979).

Veins formed in a partially melted system must have been predominantly solid during deformation. The shapes of syntectonic veins, pinch-and-swell structure and ptygmatic folds, indicate a higher competency for pegmatite than for surrounding gneiss. The viscosity of any melt phase, however, is negligible compared to the effective viscosity of the solid grains between which melt occurs (Van der Molen and Paterson, 1979). The contrast is so enormous as to make the difference between a small volume percentage of viscous melt and a small volume percentage of even less viscous hydrous fluid irrelevant for the mechanical effect of pore pressure during slow geologic deformation. This effect causes the localisation of deposition as argued above.

The geometric similarity of differentiation products in low-grade and high-grade rocks is striking to those familiar with both. Infillings between boudins, in tension gashes, and in crosscutting veins may be fibrous more often in low-grade rocks (Durney and Ramsay, 1973), but there is no suggestion of a fundamental change in the mechanism of mass transfer. At higher grades recrystallisation produces more equant crystals in the deposits to which material is continually added, whether from a melt or from a high-temperature fluid. The large recrystallised grain size is responsible for the high competency of pegmatite relative to the much finer grained surrounding gneiss.

The thermodynamic description of eutectic or cotectic melting and freezing may be rather different from that of solution and dissolution in a hydrous phase. A consideration of these factors is beyond the scope of this paper. In migmatites differentiation is thought to depend on a steady-state interconnected network from sites of anatectic melting to deformation-controlled sites of freezing and recrystallisation in dilating macroscopic fractures. After a rise in temperature, or after an increase in the water content, the melt fraction increases and melt may be arranged differently throughout the system to form a new steady-state distribution. Differentiation itself does not depend on changes in pressure, temperature or chemical composition, it can proceed at constant melt fraction as long as deformation continues. Several of the problems noted by Johannes and Gupta (1982) and Vernon (1976) in their discussions of migmatisation are readily solved if the dynamic process of syntectonic mass-transfer is taken into account.

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