The Scandinavian Caledonides as studied by Centrifuged Dynamic Models

By

Hans Ramberg[1]

ABSTRACT.—Experimental model studies of gravitationally unstable systems suggest that the rise of the basal-gneiss culminations in the Scandinavian Caledonides (and other orogenic belts as well) is a buoyancy phenomenon due to adjustment of an unstable stratification of the earth's crust in the geosynclinal region, or an unstable distribution of masses.

As a basis for the discussion a brief review of the main tectonic features of the Scandinavian Caledonides is given with emphasis on the basal complexes in the central parts of the orogenic chain.

A number of dynamic models of various unstably stratified structures run in a centrifuge are described and their structural evolution compared with the basal culminations, the gneiss-filled anticlines and the nappe structures of the Caledonides.

If the remarkable similarity in geometric pattern of models and parts of the mountain chain can be taken to mean an equally close similarity in dynamics—i.e. with reference to driving forces and movements—it follows that the Caledonian deformation is chiefly propelled by the body force of gravity acting on unstable distribution of masses in the geosyncline and its adjacent and/or subjacent neighborhood. In the models the only acting force that produces the complex pattern of rising domes and subsiding synclines, of buckling and stretching, of recumbent folds and creeping nappes is the centrifugal force which is a body force playing exactly the same role in our models as does gravity in the large-scale natural prototypes.

Content

[1] Institute of Mineralogy and Geology, Uppsala.

Acknowledgments

During the experimental work and preparation of the manuscript assistance was received from a number of persons to whom the author's sincere thanks go. Fil.kand. O. Stephansson and Mr. J. C. Lindström did a large part of the laboratory work. Mr. G. Haglund made some of the instruments and rebuilt others. Mrs. A. Kaljusaar and Miss K. Andersson made the drawings while Mrs. A.-M. Ekström typed the manuscript. Mr. G. Andersson and Mr. G. Flodvik did some of the photographic work. Mr. O. Wallner made several replicas of the models.

Preliminaries

Along the east and southeast border of the Scandinavian Caledonides the autochthonous sediments or the overthrust supracrustals of the orogen rest unconformably upon the pre-Cambrian gneissose basement in Sweden and South Norway. Little or no sign of Caledonian metamorphism or plastic deformation, only mylonitization and cataclasis, is visible in the basement here. This stands in marked contrast to the conditions in the more or less isolated regions of basement rocks exposed within the Cambro-Silurian strata of the Caledonides, especially within the coastal regions of Norway. These regions consist chiefly of quartzo-felspathic gneisses whose grade of metamorphism, which at least in part is Caledonian, varies from high epidote-amphibolite facies to granulite facies (charnockites), the latter being encountered in the Vesterålen–Lofoten area (Heier, 1960). Amongst the gneisses are remnants of highly metamorphosed and metasomatosed supracrustals some of which demonstrably are of eo-Cambrian to Silurian age (e.g. sparagmite and Trondheim schists). Anorthosite and eclogite are found especially in the huge Möre-gneiss region. Metadiabases are common in many of these basal complexes.

Field studies by a large number of geologists (see references in Holtedahl *et al.* 1960, pp. 278–284) have established that, though the basal complexes generally are positioned stratigraphically below the Cambro-Silurian strata, the gneisses often show a young, somewhat penetrative relation to the Cambro-Silurian rocks. Augen-gneiss formation, pegmatitization and migmatitization are common in the schists within the gneiss regions. In the neighborhood of the basal gneiss the Cambro-Silurian schists exhibit a regional metamorphism corresponding to that of the gneisses, and the Caledonian structure, which here chiefly is of a plastic nature, is parallel in basal complex and adjacent Cambro-Silurian schists. Often the planar structure is steep and not rarely the positions are inverted in recumbent folds with gneiss cores.

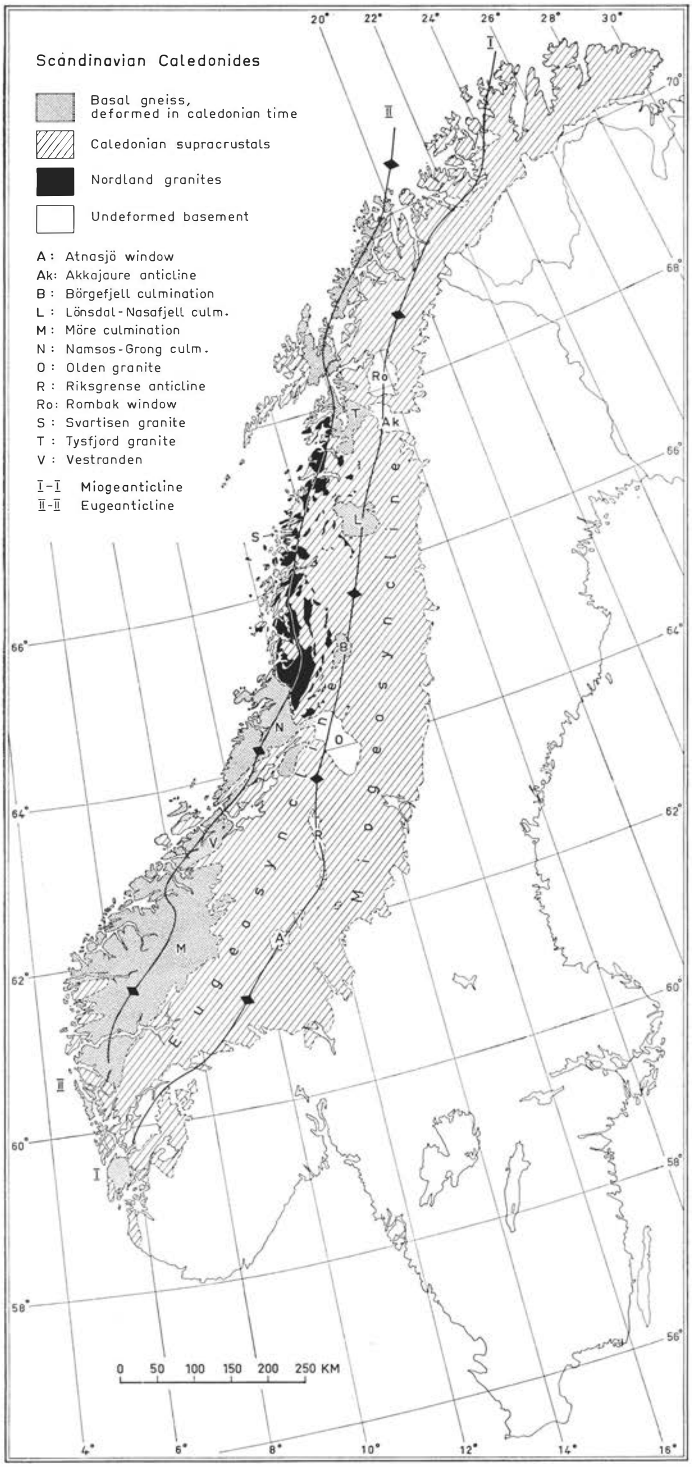
If some of the gneisses are altered schists or lavas of post-Archean age—recrystallized eo-Cambrian sparagmite constitutes for example a considerable part of the Möre- and Vestranden regions—the major portion of the basal complexes appears to be the pre-Cambrian basement exposed to recrystallization, deformation and mobilization during the Caledonian orogenic period. In two separate areas—the Grong–Olden and the Tysfjord–Akkajaure culminations—(Foslie, 1941; Kautsky, 1952) the transition from undisturbed basement to strongly mobilized basal gneiss and granite can indeed be followed continuously in the field from the eastern edge of the Caledonides in Sweden to its central zone in Norway. Isolated basal complexes located close to the eastern edge of the orogen—e.g. the Tömmerås dome (Peacey, 1964) and the Börgefjell- (Foslie and Strand, 1956; Zachrisson, 1964) and Lönsdal culminations—are less affected by Caledonian recrystallization and deformation than the culminations located farther away from the eastern edge, e.g. the Möre–Vestranden–Namsos complexes. The row of basement windows in the sparagmite area east and southeast of the Trondheim synclinorium is only mechanically altered in Caledonian time (Oftedahl, 1952; Asklund in Magnusson *et al.*, 1962, p. 196; Schaar, 1962; Strömberg, 1961).

The basal-gneiss regions are topographic culminations in the basement, culminations which have risen in Caledonian time and only to a lesser degree perhaps reflect a primary unevenness of the geosynclinal floor.

This, in a very schematic way, seems to be the consensus of the field geologists, including the writer, who have worked in the basal complexes and the adjacent metasediments and metalavas.

The general structural pattern of the Caledonides

The main structural trend in the Scandinavian Caledonides is parallel to the eastern edge of the Cambro-Silurian geosyncline. This NNE-SSW trend is marked out by the edges of a number of nappes or thrust sheets and by the rim of Cambrian and eo-Cambrian autochthonous sediments underneath the nappes in South Norway and the northern half of Sweden (Fig. 1). Within certain regions—e.g. the eastern half of the large Trondheim synclinorium and the Finmark area—metamorphic zoning is also roughly parallel to the eastern edge of the mountain chain, the grade of metamorphism increasing generally west- and northwestward away from the edge, and the higher nappes often corresponding to higher metamorphic facies than the lower ones, or the nappes being more metamorphosed than their autochthonous floor. (See Holtedahl


Scandinavian Caledonides
Basal gneiss, deformed in caledonian time
Caledonian supracrustals
Nordland granites
Undeformed basement
A : Atnasjö window
Ak: Akkajaure anticline
B : Börgefjell culmination
L : Lönsdal-Nasafjell culm.
M : Möre culmination
N : Namsos-Grong culm.
O : Olden granite
R : Riksgrense anticline
Ro: Rombak window
S : Svartisen granite
T : Tysfjord granite
V : Vestranden
I-I Miogeanticline
II-II Eugeanticline
Eugeosyncline
Miogeosyncline
0 50 100 150 200 250 KM

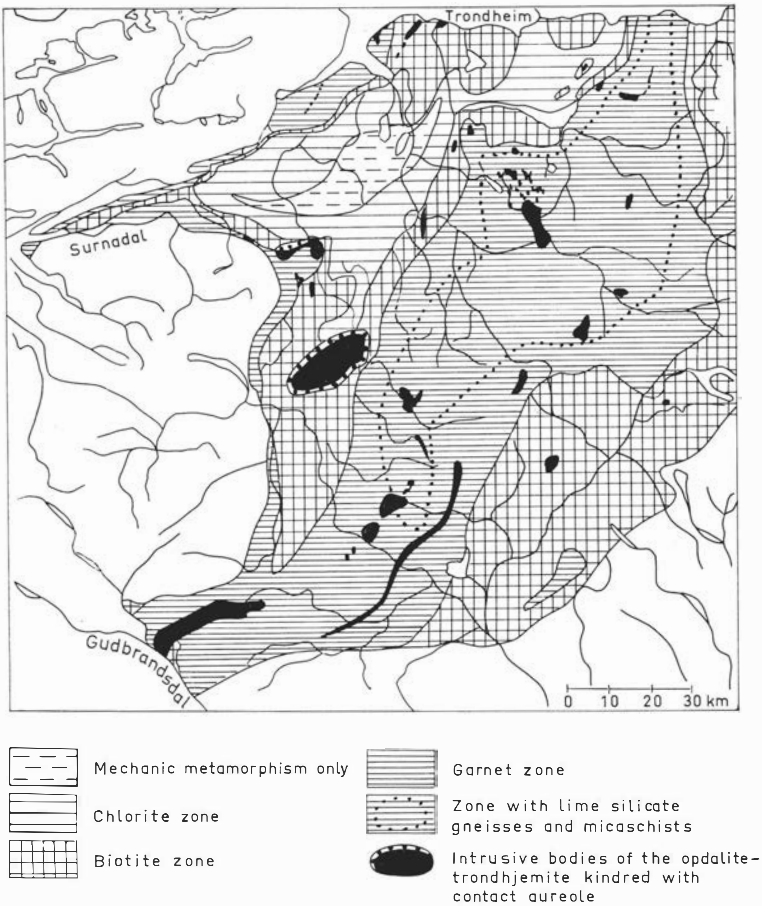


Fig. 2. Metamorphic zones in the Trondheim synclinorium, after GOLDSCHMIDT, 1915.

et al., 1960, pp. 170–278; e.g. KAUTSKY, 1952; MAGNUSSON *et al.*, 1962, pp. 171–223; LUNDEGÅRDH *et al.*, 1964, pp. 177–186).

In the central parts of the orogen, however, and in the neighborhood of the basal-gneiss culminations metamorphic zoning no longer parallels the relatively straight eastern edge of the mountain chain but is chiefly controlled by the more irregular outline of the gneiss complexes as for example shown for the Möre–Vestranden complexes by the map of metamorphic zones in the Trondheim area, Fig. 2, (GOLDSCHMIDT, 1915). Contrary to the conditions at the eastern edge of the orogen the grade of regional metamorphism increases toward the basal gneiss; downward where the position is normal but upward

Fig. 1. Outline of the Scandinavian Caledonides, simplified and modified from geological map of Norway by HOLTEDAHL and DONS, 1960, and geological pre-quaternary map of Sweden by MAGNUSSON *et al.*, 1957. Geanticlinal ridges indicated by author.

where the basement is overfolded, or overthrust, a by no means uncommon structure in the Caledonides.

A general tendency for fold axes to strike along the main Caledonian direction exists, and large-scale "banding" of rocks tends to parallel the same direction—e.g. zones of mica schist and greenstone in the Trondheim area, (e.g. WOLFF, 1964) limestone strata in the Nordland synclinorium, and in the same region many of the granitic or gneissic plutons are elongate in the main Caledonian direction.

However, structural deviations from the main Caledonian trend are common and well known not only locally but over regions of considerable extent. So-called cross folds and cross structures whose axes make an obtuse angle with the main trend have been recorded from many regions in the Caledonides (e.g. VOGT, 1928; LANDMARK, 1951; F. KAUTSKY and TEGENGREN, 1952; G. KAUTSKY, 1952; KVALE, 1948 and 1953; SKJESETH and SÖRENSEN, 1953; STRAND, 1951 and 1964; NICHOLSON and WALTON, 1963; ZACHRISSON, 1964). Structural deviations from the main NNE-SSW trend are especially conspicuous within the neighborhood of the basal-gneiss complexes which generally are wrapped conformably by the Cambro-Silurian schists. Thus, the strike tends to parallel the outline of the basal complexes as is often the case around the Möre–Vestranden–Namsos–Grong chain of culminations as well as the Tömmerås dome (PEACEY, 1964), the Börgefjell- and Lönsdal complexes (FOSLIE and STRAND, 1956) and the Tysfjord "bottom granite" (FOSLIE, 1941). Cross folds and cross lineation are particularly widespread in the thrust sheets along the eastern edge of the orogen.

Since many of the so-called Nordland granites are also conformably surrounded by Cambro-Silurian schists it follows that deviation from the main Caledonian strike is the rule rather than the exception in the schists between the densely spaced Nordland granites. Incidentally some of these granites are probably identical to highly mobilized and rejuvenated basal gneiss as shown for the Svartisen granite by the study of NICHOLSON and WALTON (1963), whose map shows how the E-W trending folds are controlled by the basal granite which occupies the cores of anticlines plunging away from the main granite body. This is a common structural style at the boundary between the caledonized basement and the overlying Cambro-Silurian. It is a typical style in the Namsos complex and in Vestranden, for example, though in the latter region the axes are parallel to the main Caledonian trend. It is also worth noting that the Svartisen granite at another locality is interpreted as a folded sheet overlying metasedimentary rocks (SKJESETH and SÖRENSEN, 1953). Part of the body may therefore constitute a recumbent fold whose upper limb is eroded away. It actually seems that many of the Caledonian granites in North Norway occur as core fillings in recumbent folds or as mushroom-shaped domes because REKSTAD (1929), in his description of the Salta district, gives numerous localities where granite or granite gneiss lie above metamorphic schist.

In view of these relationships between the basal gneiss and the Cambro-Silurian strata of the orogen it may be convenient to consider the latter separately as the Cambro-Silurian (C-S) Caledonides in contrast to both the central (western) basal culminations and the basement in the east which may be regarded as the frame of the C-S Caledonides, the basement in the east and southeast being the external frame, and the central complexes being a kind of internal frame.

Looked upon in this way we find that the structural and metamorphic patterns of the C-S Caledonides in a general way reflect the geometry of the frame; the structural trend in terms of strike of strata and other planar structures, and the metamorphic zoning being parallel to the outline of the frame at least in its vicinity. Some distance away from the frame within the interior of the C-S Caledonides—e.g. in the middle of the Trondheim synclinorium—the structural control of the frame, at least of the exposed part of it, is weakened. The smaller the irregularities of the frame—e.g. the smaller the size of the internal basal gneisses and the smaller the "wavelength" of irregularities of their border—the narrower the zone of the adjacent schists within which the irregularities are reflected. Now the internal frame of the C-S Caledonides is fragmented with short "wavelength" of the irregularities as compared with the eastern external frame. It is therefore reasonable that the boundary of the eastern frame makes itself most felt on the C-S Caledonian structure and controls the main trend.

The role of the basal-gneiss complexes

It is a common and possibly also a reasonable assumption that the Scandinavian Caledonides formed as a result of large-scale crustal compressive forces that acted normal to the geosynclinal trend. It lies close at hand to conclude then that the upheaval of the basal culminations also was a consequence of the regional compression. The basement of the geosyncline could perhaps have buckled in response to the lateral stress provided the wavelength of the culminations is not too large (RAMBERG and STEPHANSSON, 1964), or lateral compressive strain could conceivably have been localized where the basement for one reason or other was especially weak, such that complementary thickening and local elevation of the basement surface would follow.

The writer shall not comment on these possibilities but rather discuss an alternative model for the evolution of the basal culminations, namely that they rose in response to primary vertical buoyant forces, forces which instead of being induced by the lateral stress actually themselves produced much of the horizontal compression as secondary effects.

LANDMARK in his review of cross folds in the Scandinavian Caledonides (1951) considers a similar mechanism as a likely possibility though he does not appeal to buoyancy as the motivating force, but refers to injection of magma as the driving agency (p. 246).

The author's view on the mechanism of the rise of the basal complexes agrees rather closely with ESKOLA's model for the origin of "mantled domes". ESKOLA (1948) expresses explicitly the opinion that the granitic magma rises (in the mantled domes) simply because of its lesser density than the average rock (p. 476).

As will be shown below the author puts strong emphasis on the density difference as producing the vertical force, but he sees no reason to limit the movement to magmatic melts.

The general idea of primary vertical movements in fold-mountain formation is an old one discussed by NAUMANN (1849 and 1858), and developed and modified by HAARMANN (1930), VAN BEMMELEN (1960, and many earlier contributions), RAMBERG (1945) and e.g. BELOUSSOV (1961). The crucial novelty in this article is chiefly the specific application of the idea to the Scandinavian Caledonides, the explicit suggestion of the driving force for the vertical movements and the consequent lateral motions, and especially the experimental support and illumination of the concept.

Experiments with dynamic models in our new tectonic laboratory at Uppsala have convinced the author that the buoyant force of relatively low-density masses is the chief driving agency of the gneissic and granitic culminations. Our experiments also show that bodies as huge as the basal culminations need not be liquid in order to rise. Indeed, the greatest resistance against the rise is offered, not by the strain within the ascending body itself, but by the strain in the rock adjacent to the rising dome, and, since the adjacent schists and gneisses were crystalline during the rise of the culminations, rate of movement is not much increased by assuming a liquid core.

According to experiments the pattern of rise of a substance through another depends much upon the relative viscosity (or relative mobility) of the two substances. When the contrast in viscosity is of the same order as between solid rocks ($\mu = 10^{20} \pm$ poises) and a granitic melt ($\mu = 10^{6} \pm$ poises) the ascent occurred via narrow fissures, or irregular and narrow branching channels as shown in Figs. 3, 4 and 5. Only when the contrast in viscosity is much less such as can be expected between crystalline rocks of unlike composition (e.g. a $\Delta\mu$ of say 10^{2}–10^{3} poises) does the wide, bulky dome-shaped pattern develop in our experiments (e.g. Figs. 13, 54, 68 and 70).

In the following the author, motivated by the results of field study and experimental laboratory work, feels impelled to support the view that the basal culminations play a highly active role in the Caledonian evolution: they are not only obstacles against which the Cambro-Silurian strata have been compressed, neither have they but passively received the Caledonian structure. Instead their own internal structure and the structure in the adjacent C-S Caledonides are for a large part controlled by the buoyant tendency of the sialic basement of the Caledonian geosyncline.

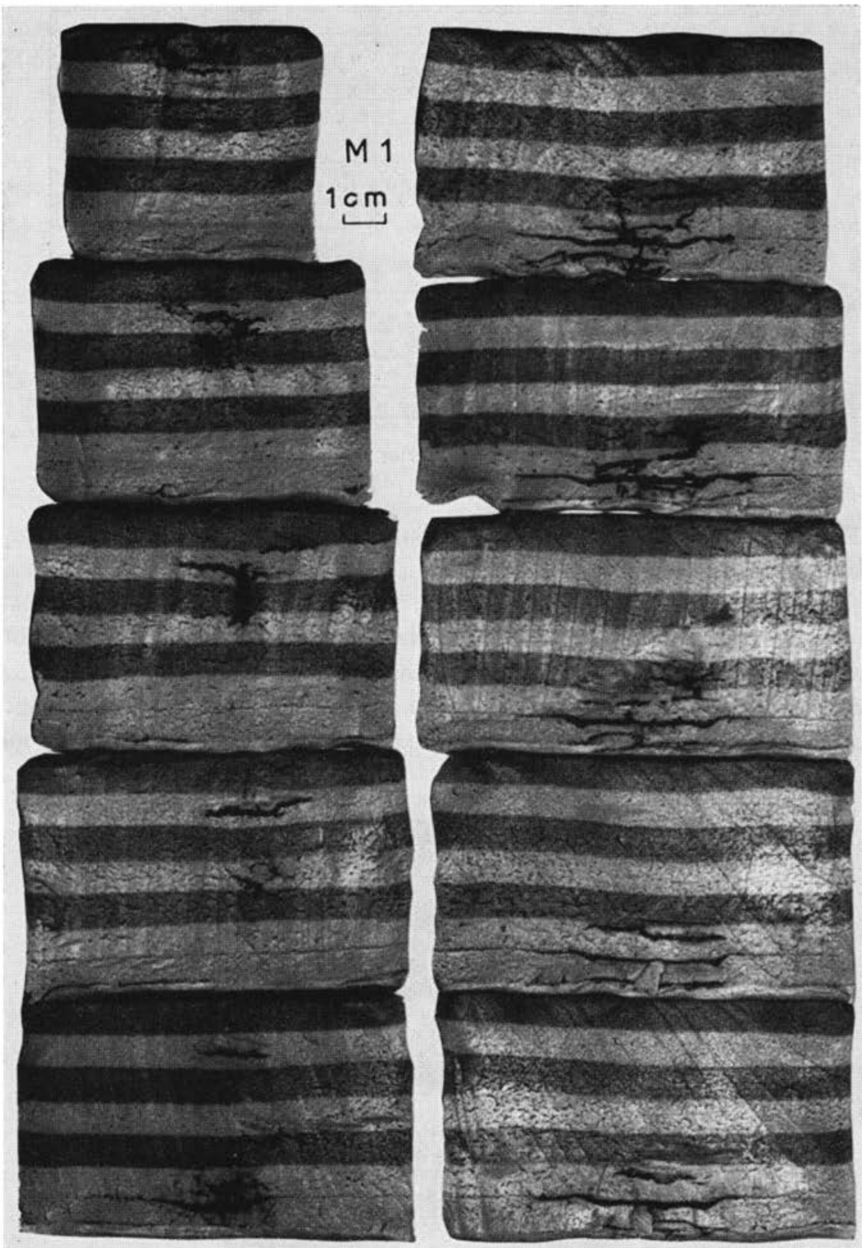


Fig. 3. Sections through a dynamic model (M 1) of a magma-imitation fluid (a $KMnO_4$-solution) rising through the imitation crust (layers of putty with unlike color) as run for 15 minutes at 210*g* in a centrifuge with the centrifugal force pointing downward in the figure. The dark irregular spots within a central vertical zone of the sections show the uneven path of the rising "melt".

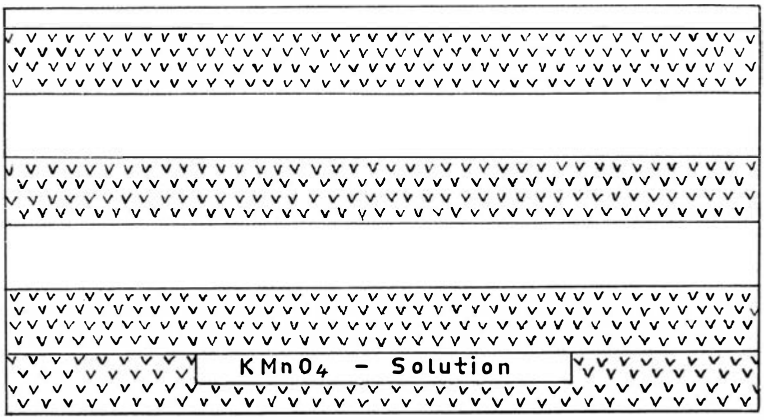


Fig. 4. Section of model M 1 prior to run in the centrifuge.

That the rise of batholiths in fold mountains is responsible for much of the folding has long been recognized, the South-American Andes being the classical region for such structures (e.g. GERTH, 1955, pp. 245–249). In the Alps the central basal massifs of Hercynian ancestry (e.g. the Belledone, the Montblanc –Aiguilles Rouges and the Aar–St. Gotthard massives) have presumably risen because of buoyant forces acting on the relatively light granitic "magma" according to KRAUS (1936 and 1951, I, pp. 68–71). The rise of such huge bodies

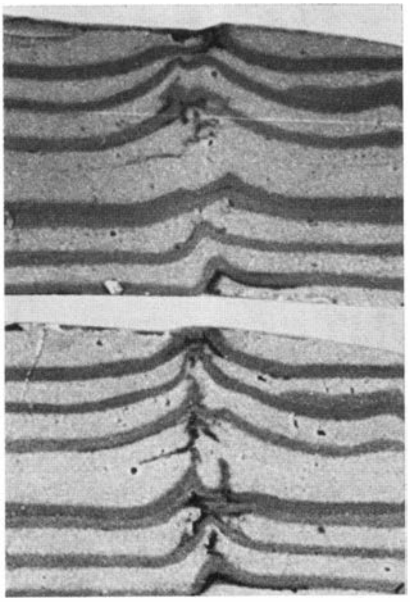


Fig. 5. Cuts through the vent along which a solution of $KMnO_4$ has risen through a layered overburden of putty and modelling clay, run in a centrifuge for less than 1 minute at 1000*g* with the centrifugal force pointing downward in the figure. Height of cuts about 3.5 cm.

could hardly have occurred without structural effect in the surroundings and within their own body.

KAUTSKY (1952, pp. 186–198) points out the great similarity between the basal-gneiss culminations in the Sulitelma–Salojaure–Tysfjord region and the mantled domes in the Appalachians. He shows that much of the enrobing strata are thrust sheets and expresses the view that the upheaval is not due to a local swelling and buoyant rise as a result of palingenesis and granitization in the sense expressed by ESKOLA (1948). KAUTSKY takes the view that the doming simply is a consequence of the crossing of two sets of anticlines, implying buckling and the action of lateral compressive stress, a view which seems to be held by the majority of field geologists who have studied the phenomenon.

Before discussing the experiments on which the author's model is based a brief survey of pertinent field observations is needed.

Field observations

The writer's chief field experience is from the Möre–Vestranden–Namsos regions of basal gneisses for which reason the following account concentrates on these areas. The huge Möre culmination and the smaller Namsos–Grong culmination are connected by a strip of basal gneisses with Cambro-Silurian inliers exposed along the coast of Nord Möre and Tröndelag. This strip was coined Vestranden by TH. KJERULF. In recent years parts of Vestranden have been studied by a number of geologists, e.g. CARSTENS (1955 and 1956), HERNES (1956), STRAND (1953*a* and *b*), RAMBERG (1943) and map in press, based on several summers' field work, Fig. 6.

The structural style of Vestranden is one of tight folds about rather flat and unusually straight axes that parallel the main NE-SW Caledonian trend. Around the outlet of Trondheim fjord and south of this region the axes plunge generally gently toward NE whereas the plunge is toward the SW in the transition between the homoaxial Vestranden structure and the more heterostructural Namsos gneiss, Fig. 7.

Rock types in Vestranden are various quartzo-felspathic gneisses conformably alternating with strongly metamorphosed sparagmite and Trondheim schists, including metabasalt (greenstone) and marble. A characteristic red potash gneissic granite occurs often as anticlinal cores robed by flagstone (sparagmite), mica schists and/or amphibolites which may be traced continuously to demonstrated Cambro-Silurian strata in the Trondheim area. Some of the red granitic gneisses appear to derive from flagstone or a fine grained leptitic rock whose origin is unknown. Granodioritic gneiss, often carrying hornblende, is also positioned as the core of anticlines.

One particularly significant structural feature of Vestranden is the intensive stretching along the fold axes. This stretching is documented by mineral elongation, linear orientation of hornblende and micas, and boudinage or pinch-

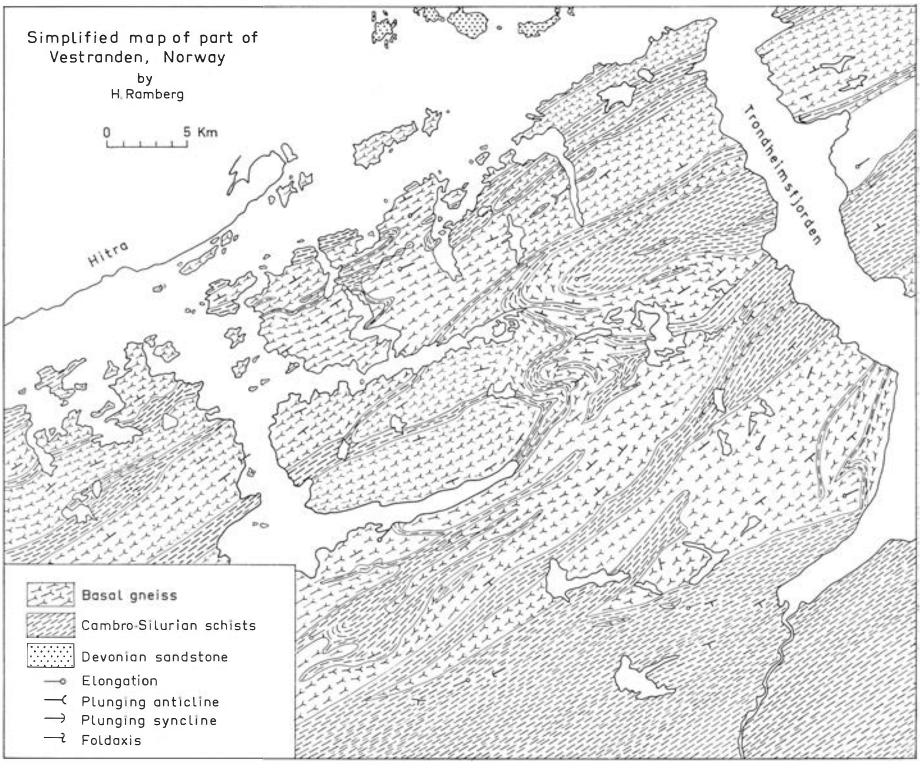


Fig. 6. Simplified map of a part of Vestranden. Published by the permission of the director of the Geological Survey of Norway.

and-swell structure of the more competent rocks such as amphibolites. Tensional cross fractures are common on microscopic-, megascopic- and regional scales.

Toward the Möre complex in the south and the Namsos complex in the north the monotonous homoaxial style of Vestranden merges with the more "bulky", cupolous structure of these two basal culminations. A map by BIRKELAND (1958) shows a number of domal structures in the Namsos culmination, especially its northern part, Fig. 7. Carstens, who has studied the boundary region between the Namsos gneiss and the Trondheim schists states about the situation there (1960, p. 6) that: "Dome structures are a characteristic feature of the basement rocks." Reconnaissance field work by the writer and a study of aerial photos of this very well exposed tract show that the homoaxial Vestranden structure changes to the irregular structure of the Namsos culmination a little north and north east of Åfjorden (see map, Fig. 7).

The recent geologic map of Norway by Holtedahl and Dons shows the strike to turn in wide curves in the eastern half of the Möre complex, indicating a huge eastwardly plunging anticline. Field work by GJELSVIK (1953) also indicates that a huge portion of the eastern half of the Möre complex constitutes an anticlinal or domal structure. On a somewhat smaller scale a complex cupolous

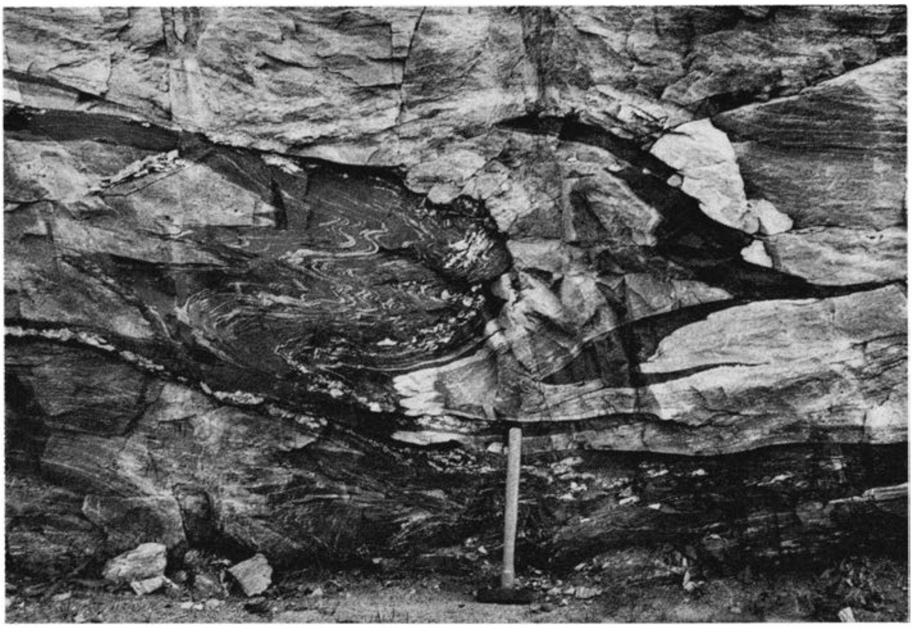


Fig. 8. Medium-scale folds with nearly horizontal axial plane in amphibolite and gneiss, on crest of large basal-gneiss anticline, Ratvåg between Trondheim fjord and Namsos. Phot. H. R. 1964.

nappe pattern, partly with inverted stratification, has developed in the sparagmite-capped basal gneiss southwest of Oppdal (O. HOLTEDAHL, 1938; ROSENQUIST, 1941; H. HOLTEDAHL, 1950). A quotation from HOLMSEN (1960, p. 12) about the basal complex in the Oppdal district is pertinent: "The large-scale structural picture of the culmination area thus exhibits a number of dome-shaped anticlines, partly overturned in diverging directions. The domes are separated by deeply downfolded, elongated and curved synclines, containing the younger rocks and the overturned and inverted parts of anticlines of the basement gneisses."

The nappe structures recorded by MURET (1960) from the Möre gneiss are also of interest, see Fig. 61.

It is noteworthy that at several places evidence exists to the effect that at least some of the large antiforms in the gneiss are not simple buckle folds but must have been produced primarily by a push from below. This is shown by the particular pattern of some second-order folds which occur at the anticlinal region of the major folds. These second-order folds are isoclinal with a more or less horizontal axial plane that is parallel to the layering of the antiform. This situation is for example exhibited in the large antiform in the gneiss a little north of Åfjorden in the boundary region between Vestranden and the Namsos culmination, see Fig. 8.

Toward the east, southeast, south and partly toward the southwest the Möre culmination borders toward the Cambro-Silurian Caledonides. The structure on either side of the boundary is generally conformable and in most places dipping away from the boundary though inversions are found.

The Möre culmination, Vestranden and the Namsos–Grong culmination are bordered by depressions in the basement toward the east, southeast, south and southwest. [In the west the gneiss complexes are cut by the coast line.] Thus, in a general way one may speak of a marginal syncline in the basement, filled with the synclinorium of Trondheim schists toward the east, with the crystalline Jotun nappes above Cambro-Silurian schists in the southeast and filled with the rocks of the Bergen arc in the south–southwest. The main trend of this large depression in the basement is parallel to the outline of the culmination, but it is significant that a number of smaller synclines branch out from the main trough and extend some distance into the Möre culmination and Vestranden, Fig. 7. For a considerable distance the two latter complexes are separated by one of these branches, the Surnadal syncline.

The branch synclines plunge away from the huge Möre culmination. This is true for the fold axes as well as the elongation and lineation where these have been observed.

It is noteworthy that the marginal syncline is separated in three sections by two major cross anticlines, one at Vågå between the Jotun syncline with the Jotun nappes, and the Trondheim synclinorium, another at Grong–Olden separating the Trondheim- from the Nordland synclinoria. The cross anticline in the basement at Vågå has been studied in detail by STRAND (1951 and 1964) who has demonstrated the east-west trend of both fold axes and elongation in this area.

The only certain Cambro-Silurian strata known on the west side of the Möre culmination are the east-west trending synclinal structures overlain by Devonian at Hyllestad–Lihesten, Kvamhesten, Håsteinen and Glopen–Hornelen.

The orientation of the elongation is of great importance for movement analysis. Unfortunately not much information is available on this point except for Vestranden, the Bergsdalen area east of Bergen and the crucial Vågå area where the Trondheim synclinorium meets the Jotun nappes. In these regions elongation and fold axes are generally parallel (unpublished observations by the author in Vestranden, and records by KVALE, 1948, for Bergsdalen and by STRAND, 1951, for the Vågå district). As this is a common feature in deep tectonics one may assume that a similar situation prevails along the boundary of the basal complex. WEGMANN (1959) in a study of the structure along the east boundary of the Möre culmination at Driva in the Oppdal district showed a general eastern plunge of the axes within a rather narrow contact zone. On a visit to the northeast boundary of the Namsos complex the author found a strong lineation plunging about 30° toward the east.

The structure of the Hornelen and Kvamhesten Devonian synclines on the west side of the Möre gneiss is significant. The western part of these synclines shows a conformable structure between basal gneiss and overlying schists whereas in the eastern part thrust planes separate discordantly the Old Red from the gneiss in the east. Writes HOLTEDAHL (1960, p. 293) "We get the impression that the eastern rock masses have moved (glided?) westward above the basement while the western ones have kept their normal contact downward." Such structure may well have developed as the result of a post-Old Red upheaval of the basal gneiss east of the synclines.

The detailed record of lineation in the Bergsdalen nappe area east of Bergen by KVALE (1948 and 1960) shows a rather complicated pattern indicative of local variations of movement in space and time. However, KVALE concluded that the chief direction of movement of the nappes relative to their pre-Cambrian basement was toward the east–southeast (1960, p. 30). In the basal gneiss west of the Bergen-arc syncline a strong lineation, which is parallel to the fold axes (KVALE, 1960) plunges generally toward the east, that is more or less down the dip of the western contact of the inner Bergen arc. This agrees with a rising movement of the basal gneiss west of the arc system relative to the latter.

The structural pattern that indicates a rising movement of the basal complexes relative to the surrounding Cambro-Silurian Caledonides should be seen in relation to the contrasted degree of regional metamorphism of the two systems. The metamorphism in the Möre culmination is generally higher than that in the schists in the Trondheim synclinorium or the schists underneath the crystalline nappes in the Jotunheimen area. It is especially interesting to note the presence of eclogites in the Möre gneiss as studied by ESKOLA (1921). These basic bodies must once have been buried below several tens of kilometers of rock in order to have acquired their special mineralogical paragenesis (see the discussion on eclogites by YODER *et al.*, 1962, Fig. 43, p. 498). Possibly a layer of overburden more than 30 km thick has been removed by erosion in parts of the Möre culmination. During this slow elevation and consequent release of pressure the more chemically mobile quartzo-felspathic minerals would recrystallize and acquire the characteristics of lower grade of metamorphism (mainly amphibolite facies though granulite facies also occurs) whereas the chemically sluggish eclogitic assemblage would remain essentially unchanged in the internal parts of the bodies (RAMBERG, 1963).

A model of the evolution of the Möre–Vestrand–Namsos culminations

Since it appears to be common agreement among field geologists familiar with the Scandinavian Caledonides that the basal complexes have been elevated relative to the surrounding regions of Cambro-Silurian rocks in Caledonian time a realistic mechanism for this vertical movement must be sought. It is too much to hope for a mechanism that explains all geological details of the regions, but the following requirements must be fulfilled, viz: (1) that the essential field facts are explained, (2) that no crucial observations are demonstrably contrary to the model, and (3) that the model is physically realistic.

Records from other orogenic regions—e.g. the South American Andes (GERTH, 1955), the Rocky Mountains (JONES, 1963 and EARDLY, 1963), the American–Canadian coast range, the Appalachians (BILLINGS, 1945) and the Alps (KRAUS, 1936 and 1951) show that upheaval of quartzo-felspathic masses, both as rather homogeneous batholiths, as mantled domes and in the form of gneissic complexes, is an integral part of orogenetic evolution.

The problem of the central culminations is therefore important for mountain-making processes in general.

The physical principles behind the suggested model are simple: as a consequence of alkalifelspar and quartz being less dense than other common rock-forming minerals there exist forces that tend to place granitoid material in the uppermost part of the earth's crust. These forces are of chemical nature as well as mechanical. The effect of the gravitational field on the chemical forces has been discussed in several papers by the author (RAMBERG, 1944*a* and *b*, 1945, 1948). Though these forces and their associated processes are of consequence for the evolution of the basal gneisses and the metasomatism of the Cambro-Silurian strata we shall in the present discussion deal solely with the mechanical aspect of the problem.

It follows from elementary mechanics that a layer of less density placed below an overburden of higher density represents an unstable arrangement in the field of gravity. There exists a tendency for the two layers to exchange place by one mechanism or other. That the unstability of such an arrangement leads to salt-dome rise is since long an accepted fact well supported by model experiments and detailed field studies, but when it comes to vertical movement of solid silicate rocks and whole portions of sial buoyancy has only occasionally been called upon as the driving agency (e.g. GROUT, 1945, ESKOLA, 1948 and KRAUS, 1951 discussed the buoyancy of granitic melts; RAMBERG, 1963 studied experimentally the buoyant ascent of crystalline bodies).

Before discussing the buoyant force in connection with the upheaval of the basal culminations in the Caledonides we shall comment briefly on the pure physics of the phenomenon in terms of experiments and theory.

Some introductory experiments

To study by simple means the behavior of an unstable density stratification in the field of gravity one may place a layer of viscous oil in a flat-bottom container and fill up with water, syrup or another suitable fluid which is somewhat more dense than the oil and does not mix chemically with the latter, Fig. 9. To prevent disturbing the even-thick oil layer while filling over the "overburden" fluid it is found practical to increase the viscosity by refrigerating the oil layer.

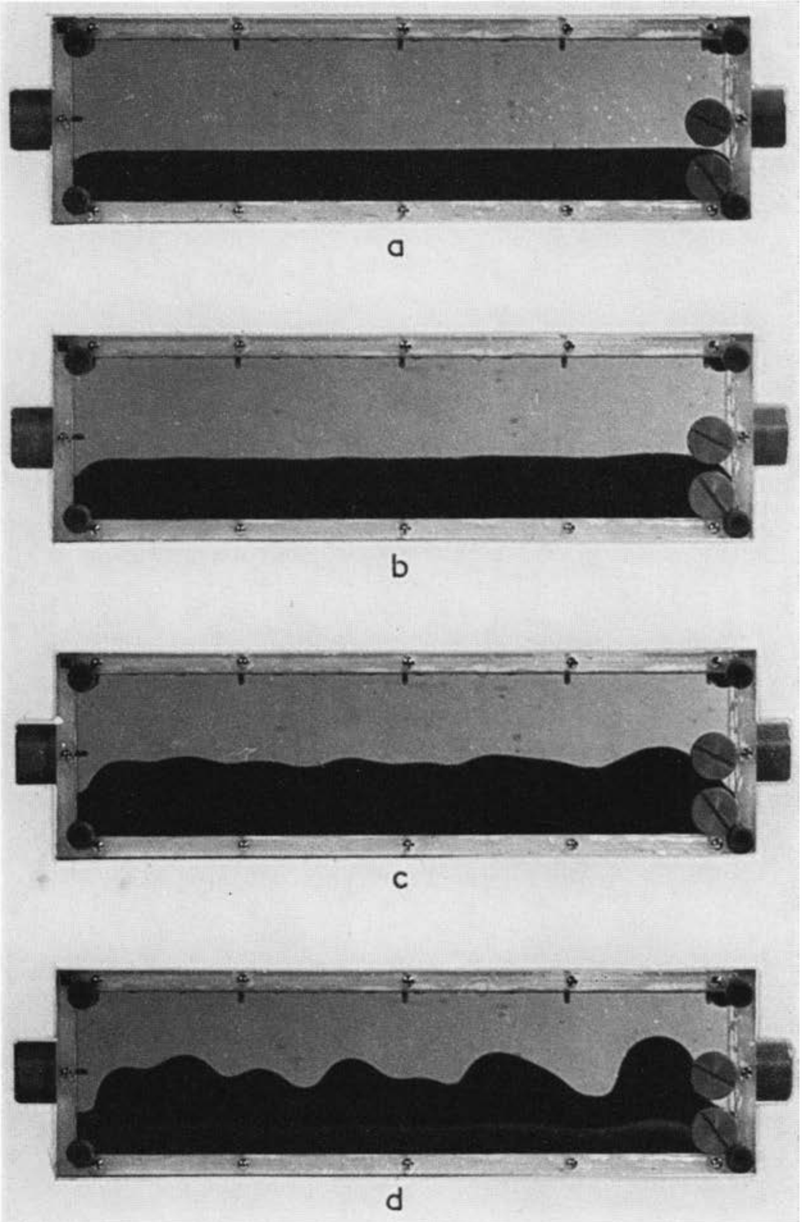


Fig. 9 *a*

A more convenient arrangement consists of a rectangular completely tight box of e.g. plexiglass filled entirely with the oil layer and the suitable overburden fluid.[1] Normally the oil floats on the top and the unstable layering is simply

[1] A closed plexiglass rectangular box for dome experiments is also used by Dr. S. B. SPIJER to whom thanks are due for interesting discussions.

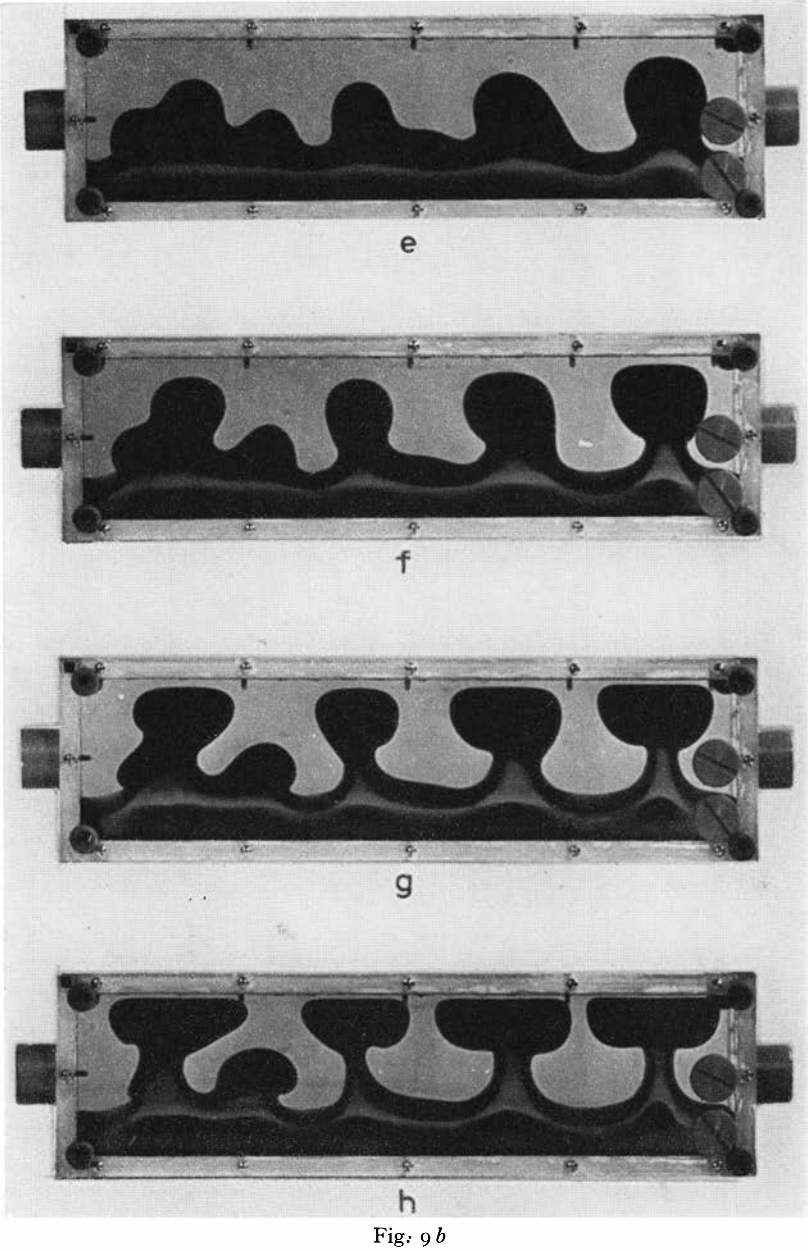
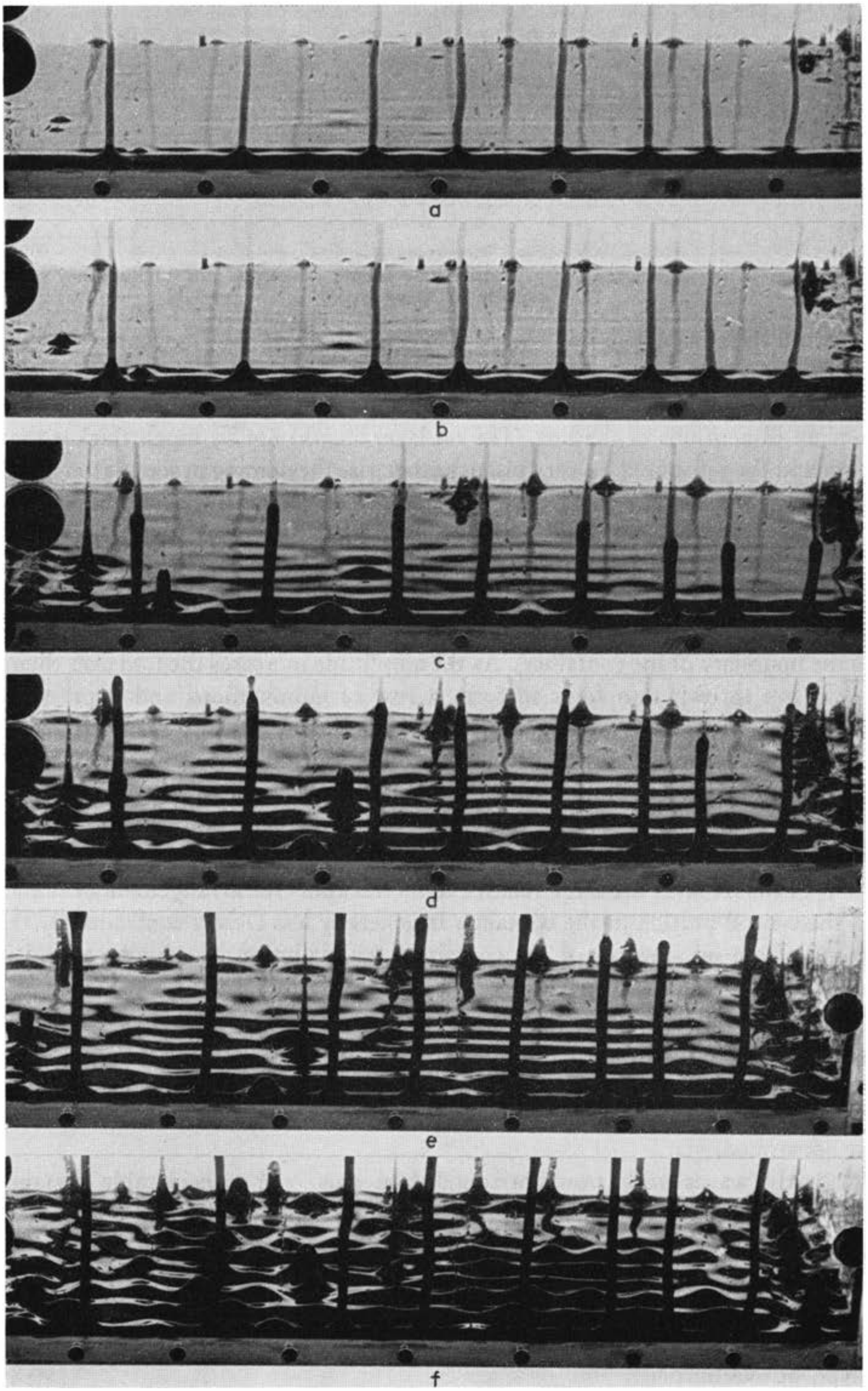


Fig. 9 *b*

Fig. 9. Successive stages of dome-evolution of oil in syrup overburden. Note even spacing between dornes.

Fig. 10. Successive stages of the development of folds and domes, due to buoyancy of a layer of viscous oil (black) with density 1 g/cm³ overlain by a salt solution with density 1.2 g/cm³.


a
b
c
d
e
f

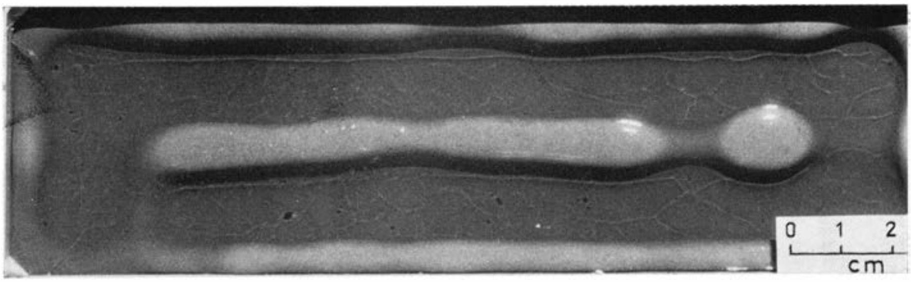


Fig. 11. Anticline of silicone putty, density 1.14 g/cm³, developed due to buoyancy when overlain by heavy syrup.

achieved by turning the box upside down which starts the buoyant rise of the oil. After some time the stable layering is established again, and the box is ready for a new turn and so on ad infinitum. About 30 cm × 20 cm × 10 cm are convenient dimensions of the box. The oil layer should not be much thicker than 1 cm and the oil should be very viscous otherwise the doming process is too rapid.

Though the surface of the oil layer may be completely even for a short time after the "overburden" fluid is poured over (or the box has been overturned) one soon notes that the oil surface starts to develop waves whose amplitude gradually grows. These waves, which of course neither oscillate nor move laterally, at first have nearly horizontal axes (whose strike by the way is parallel to the boundary of the container). As the amplitude increases the fold axes themselves are thrown into folds to form a row of culminations and depressions along the anticlines. With increasing speed the domes rise into the overburden. If the overburden is thick relative to the oil layer, and thus also relative to the cross section of the domes whose lateral dimension is proportional to the thickness of the source layer, it is seen that the upper parts of the domes become dispatched and rise to the surface as separate bodies. If both horizontal dimensions of the oil layer are large relative to its thickness the arrangement of domes in the central portion of the container is generally less clearly controlled by the shape of the oil sheet or of the container. An evolution of the above kind is shown in Fig. 10.

Because of the homogeneity of the media and the simplicity of the initial layering this test fails to show many of the structural details of interest to tectonics such as developed in the more realistic models described below, but on the other hand some of the essential features of dome evolution are clearly demonstrated, viz:

(1) firstly waves with near-horizontal fold axis over considerable distance develop at the unstable interface;

(2) the fold axes tend to parallel the shape of the container (or the boundary of the source sheet, or of the overburden, see p. 45);

(3) the anticlines develop gradually into rows of domes which then consequently become arranged parallel to the outline of the container (or source layer, or overburden);

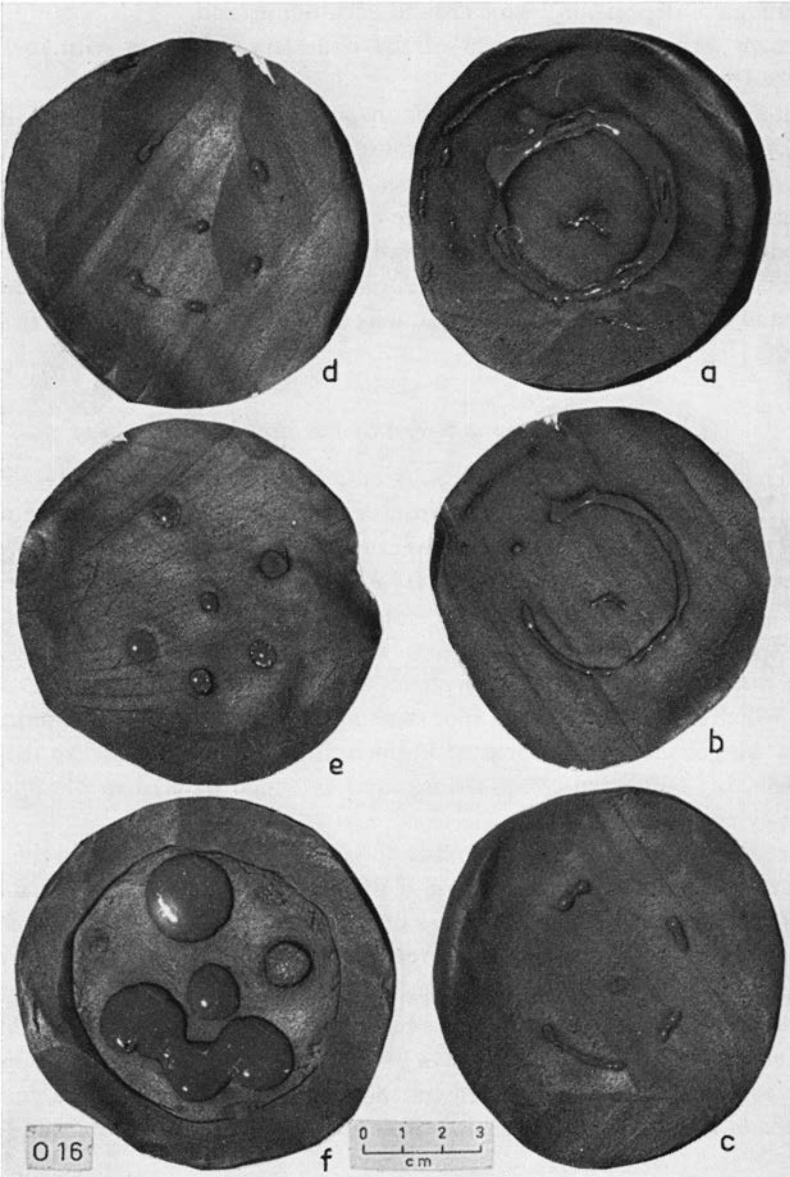


Fig. 12. Horizontal sections at different levels through model of buoyant silicone putty, $\varrho = 1.14$ g/cm^3, which has risen from a source layer through an overburden of painter's putty, $\varrho = 1.81$ g/cm^3, run in centrifuge. *f* shows free surface, *a* shows deepest section just above the source layer and cutting through the early-formed ridge whose shape is controlled by the circular outline of the model. Note the splitting-up of the ridge into separate domes at higher levels (*c*, *d*, *e*).

(4) a marginal depression forms around each dome, and
(5) the pattern-controlling effect of the boundary decreases with increasing distance from the boundary.

This pattern-controlling effect of the shape of the container (or of the buoyant sheet or overburden) is perhaps even more striking in the test shown in Fig. 11 where in a narrow rectangular container only one single anticline with almost perfectly horizontal axis grew along the centerline of the container (beside, the buoyant material also rose along the wall).

In a circular container an anticlinal culmination with closed fold axis develops parallel to the circumference, and gradually a number of domes evolve from that anticline (Fig. 12).

The principle flow pattern of the doming movement

When a dome or an anticlinal culmination grows in an unstably stratified system the rising material is taken from a region in the unstable layer around the root of the dome or anticline. This zone may be called the source zone, and the unstable layer the source layer. The flow lines[1] in the source zone lie in radial planes about a circular dome and in planes normal to a rising anticline.

That is, around a circular or somewhat oblong dome the motion in the source zone is in the form of convergent flow. An element of fluid (e.g. a small cube) which moves toward the root of the dome is continuously strained as it moves. The element is compressed in horizontal direction normal to the radial flow line (i.e. tangential compression), and extended parallel to the flow lines (radial extension).

Since the source layer also shrinks in the vertical dimension within the source zone in consequence of development of the marginal syncline, the radial extension is greater than what compensates only for the tangential compression in the theory of fluid mechanics of incompressible fluids.

Because of drag along the upper and lower boundaries of the source layer a horizontal component of shearing has to be added to the strain described above. But a small element of fluid midways between the upper and lower boundaries of the source layer is not affected by shear in the flow direction, a condition which is of interest for the question of *a*- and *b* lineation in tectonics and petrofabric.

The strain is not observable in the uniform materials used in the tests above, but in more realistic experiments described below passive as well as active markers, usually in the form of layers, show the details of the flow pattern. [A passive marker has the same rheological characteristics as the enclosing material. Such a marker records flow undisturbed by the presence of the marker. An active marker is in our tests generally more competent than the enclosing

[1] The term "flow lines" in this article is synonymous with the term "particle path" often used in fluid dynamic literature. "Flow lines" and "stream lines" are not identical except in steady flow, see e.g. COLE, 1962 p. 16.

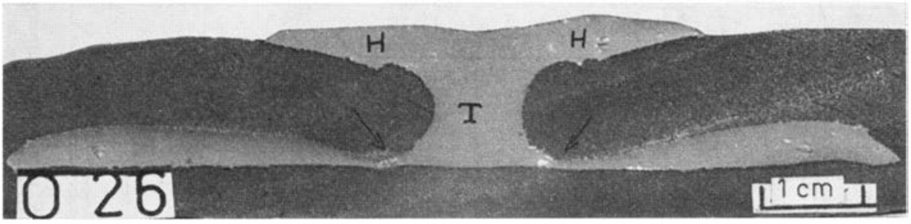


Fig. 13. Cross section of dome of silicone putty (light grey) pierced through overburden of painter's putty (dark grey) with well developed marginal sink (arrows), trunk (T) and hat (H).

material, thus, the marker signals compressive strain by developing buckles and extensive strain by developing tensile fractures, boudins etc.]

The flow pattern described above can be visualized by the special types of deformation of an active marker in the form of a relatively thin sheet of competent material embedded in the source layer. Such a marker sheet, if not too stiff or too thick, develops buckles with radial fold axes in the zone of convergent flow in the source zone, the fold axes bending into a vertical direction as the flow passes the root zone and enters the trunk portion of the dome. The radial extensive component of the flow produces often tension fractures or necked-down zones in the competent sheet in direction normal to the flow lines.

Upon approaching the free surface the domes spread laterally leading to fragmentation of the competent enclosed sheet and lateral separation of the fragments.

A geologically significant feature of the dome evolution is the development of a marginal syncline above the source zone because of thinning of the source layer and sagging of the overburden. However, details of the marginal syncline is best studied in models of plastic or viscoelastic materials. Figs. 13 and 31 give for example excellent impressions of the structure of a marginal syncline.

Experiments with more realistic materials and structures

The movements and deformations generated by the buoyant rise of a part of an unstable basement through its overburden belong to the type of tectonics called gravity tectonics because the propelling force is the pull of gravity. This kind of tectonic processes are best studied by centrifuged models in which the

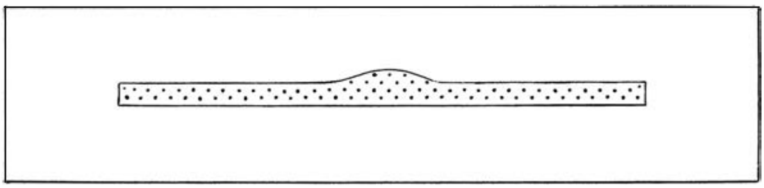


Fig. 14. Model O 26 before run in centrifuge. Dotted: silicone putty; white: painters putty. Bulge on source layer made in order to determine the site of dome formation.

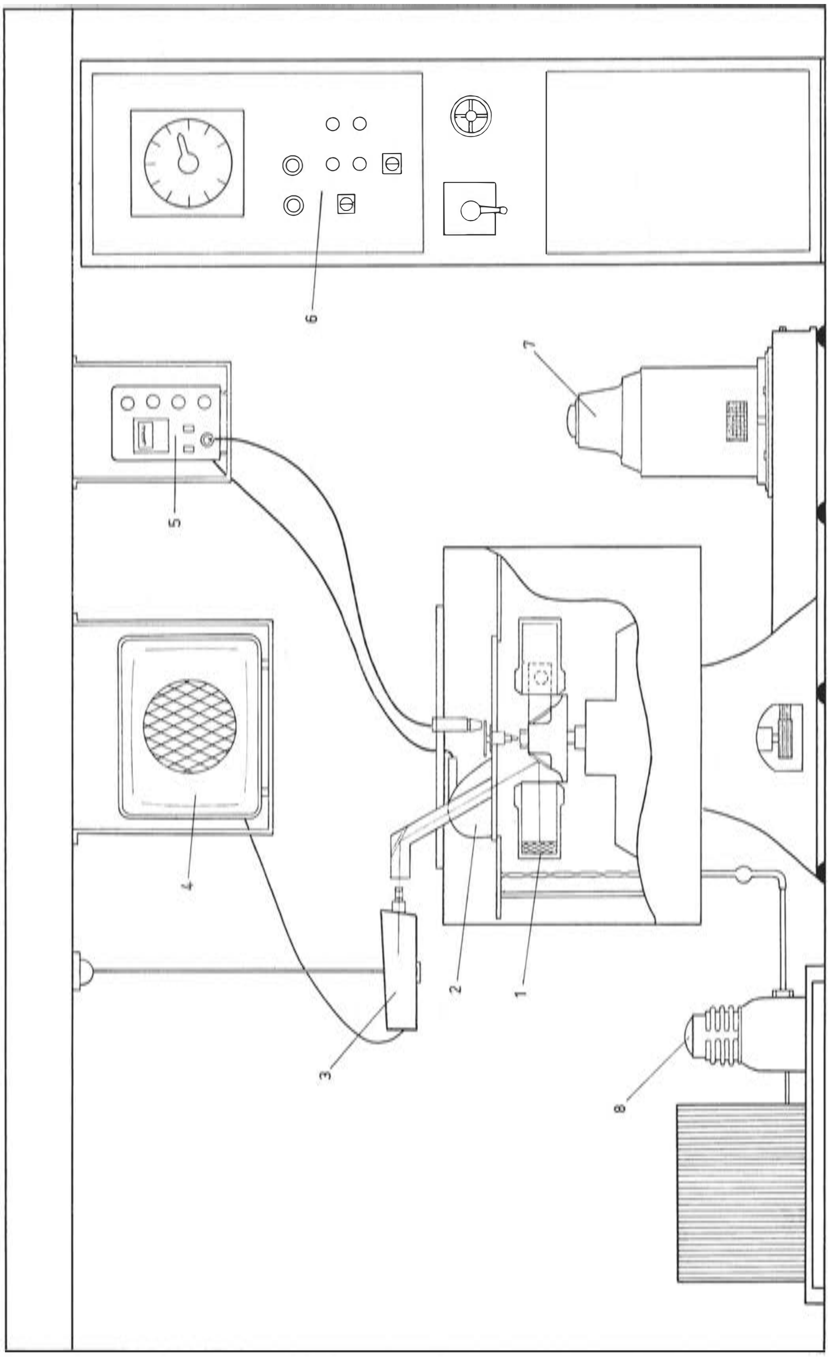


Fig. 15. Centrifuge arrangement. 1: model in centrifuge cup, 2: stroboscopic light reflector, 3: TV camera, 4: TV receiver, 5: stroboscope, 6: temperature and speed control cabinet, 7: motor, 8: refrigerator unit.

gravitational pull is replaced by a strong centripetal acceleration which permits one to use materials with considerable strength and high viscosity without violating the scale requirements.

At present there is no other practical technique by which the detailed folding, fracturing, boudin-formation and other typical structures may be studied as integral parts of gravity tectonic. Another possibility, namely to replace the body force of gravity by a magnetic force (RAMBERG, 1963, p. 4) does not give enough room for variations in the models. The great advantages of the centrifuge method for tectonic model studies have been discussed at several occasions by the writer and coworker (RAMBERG, 1963; 1964*a* and *b;* 1965; 1966, and with STEPHANSSON 1965) and need not be repeated here. At this place it may be pertinent to mention that a manuscript to a book on gravity tectonics in theory and experiment is in preparation.

We shall in the following describe a number of centrifuged models which we think are pertinent to the Caledonian problem at hand.

Shape and dimensions of domes

In their interesting paper on salt-dome models PARKER *et al.* (1955) found that the diameter of their domes was approximately the same as the thickness of the source layer. If this be unconditionally true it would be fatal to the suggestion that the basal culminations in the Caledonides are domal protuberances rising from an unstable substratum. The Möre culmination in West Norway is at least 150 km across its shortest diameter. This must be much more than the thickness of a gneiss-granitic substratum of the Caledonian syncline. Even the Tömmerås dome, which is one of the smallest basal upheavals, is more than 20 km across, a figure which would not be unrealistic for the thickness of continental sial, but probably rather high for sial in a geosyncline.

Now, our model experiments show that the relation between dome cross section and source-layer thickness (called D/h in the following) depends on a number of parameters such as:

(1) The particular stage of evolution: the more mature a dome the less the D/h

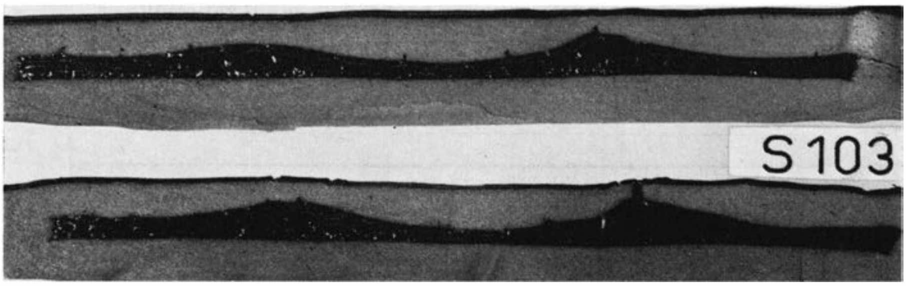


Fig. 16. Cross sections through model showing early stage of dome formation of black silicone putty, $\varrho = 1.34$ g/cm^3, in painter's putty, $\varrho = 1.87$ g/cm^3. Run in centrifuge for 30 minutes at 2000g.

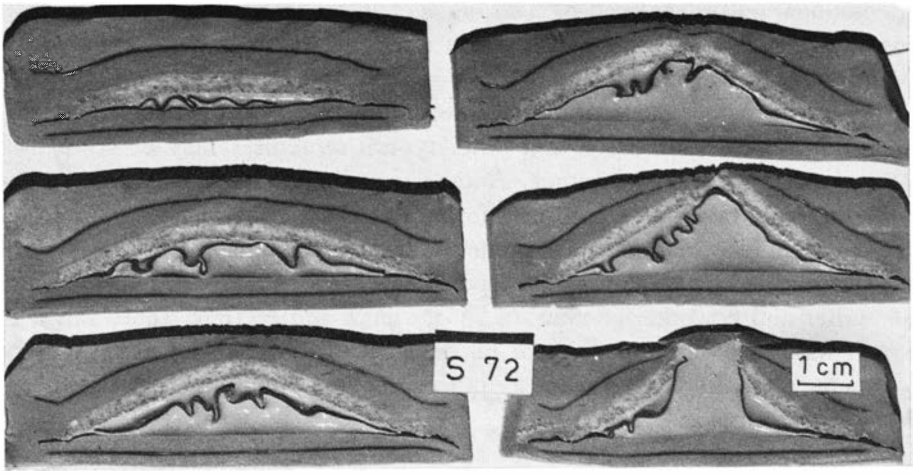


Fig. 17. Sections through centrifuged model of silicone putty, painter's putty, modelling clay and wax powder arranged as shown in Fig. 18 before run. Note the very broad shape of the dome, the buckling of the competent dark sheet in the doming body, and the boudinage of similar sheet in overburden. Run for 10 minutes at 1300*g*.

value. Moreover, at an advanced stage the domes are mushroom-shaped such that the D/h value varies greatly with the position of the cross section measured. In the initial stage the bulge on the source layer is wide relative to h, see Fig. 16.

In their early stage domes may be shaped like cylindrical folds with horizontal axes (Fig. 10), the axes being parallel to the boundary of the source sheet or to some linear or steep planar structure in the overburden. If source sheet or overburden varies in thickness such that the thickness has a gradient that is uniform over a certain region, then the axis of the early anticlinal domes tends to form normal to the thickness gradient. This is well demonstrated in our experiments.

For such anticlinal domes the dimension parallel to the axis bears no relation to the source-layer thickness.

(2) The strength or stiffness of the overburden. A strong overburden will bend only in wide, gentle curves or perhaps break and be pushed up as large flakes by the buoyant substratum (Figs. 17 to 22). Under such conditions the culmination can be much wider than the thickness of the original source layer.

(3) The mobility or viscosity of the material below the source layer is also significant. If this material is rigid the dome diameter is relatively small, if the

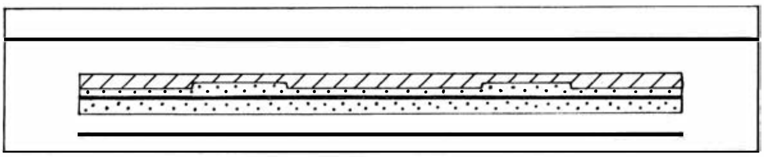


Fig. 18. Section through model S 72 before run. Dotted: silicone putty, $\varrho = 1.14$ g/cm^3; white: painter's putty, $\varrho = 1.87$ g/cm^3; black: modelling-clay sheets; inclined hatching: powdered wax soaked in oil, $\varrho = 0.9$ g/cm^3.

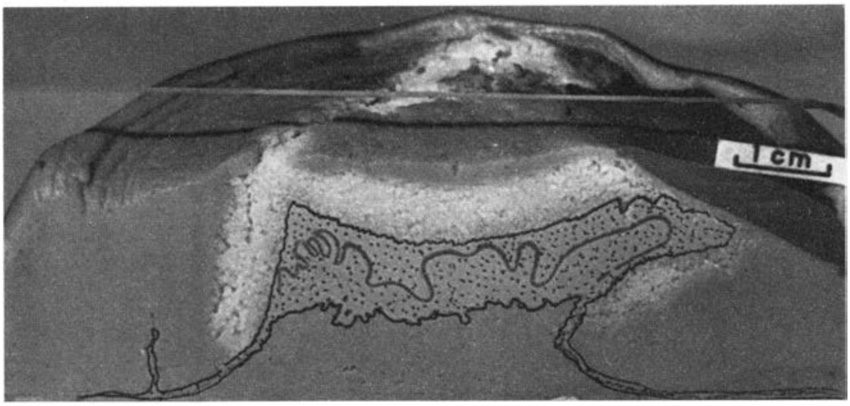


Fig. 19. Section through centrifuged model similar to S 72 but with somewhat thicker layer of powdered wax between source layer and overburden. Dotted: silicone-putty "dome" (note buckled layer of modelling clay); white: oil-soaked wax; uniform grey: painter's putty.

basal material is mobile the drag along the bottom surface of the source layer becomes negligible, material flows easily toward the domes and their diameter becomes therefore large relative to the thickness of the source layer. This may be demonstrated by letting the source layer float on a heavy liquid such as mercury, Figs. 23 and 24.
(4) Competent sheets embedded in the source layer tend to make the initial domal structure very broad relative to the thickness of the source layer, Figs. 25, 26.

In principle, therefore, our tests (e.g. Figs. 17, 19 and 25) show that the possible variation of the relation between D and h in dome models does not make the Caledonian basal culminations impossible as buoyancy phenomena. It is furthermore likely that the huge complexes such as e.g. the Möre gneiss consist of a number of minor buoyant bodies that have coalesced with remnants of the overlying supracrustals squeezed between. This seems to be the case in the Namsos complex as mapped by BIRKELAND (1958), and found by unpublished recent field study between Trondheim fjord and Namsos by the present author.

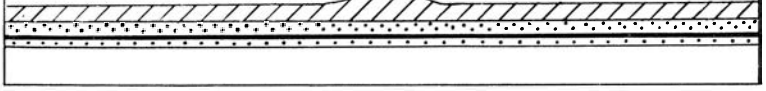


Fig. 20. Model as shown in Fig. 19 prior to run in centrifuge. Dotted with black line: source layer of silicone putty with layer of modelling clay; inclined hatching: oil-wax mixture; white: painter's putty.

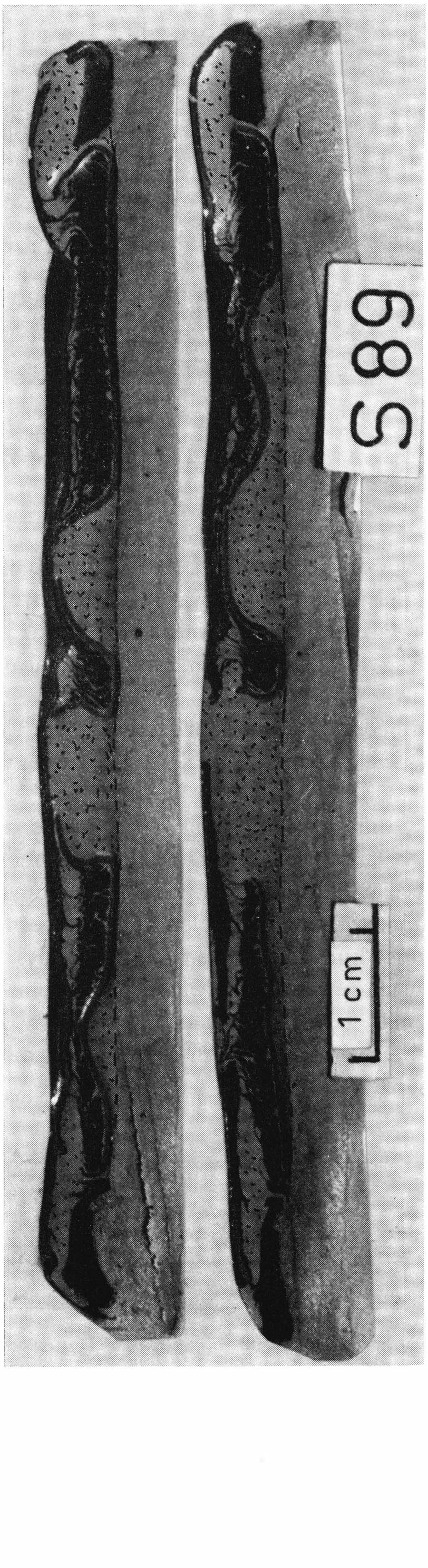


Fig. 21. Sections through centrifuged model of domes of silicone putty (dotted) with density 1.14 g/cm³ formed in overburden of heavy silicone putty (dark), density 1.34 g/cm³, with stiff sheet of modelling clay in overburden and in source layer. Light grey uniform substance on bottom is heavy painter's putty. Run for 20 minutes at 2000g.

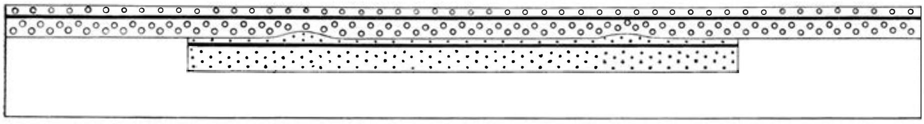


Fig. 22. Section through model S 89 before run in centrifuge. Dotted: red silicone putty, $\varrho = 1.14$ g/cm^3; small circles: silicone putty with magnetite powder, $\varrho = 1.34$ g/cm^3; black: modelling clay sheets; white: painter's putty, $\varrho = 1.87$ g/cm^3.

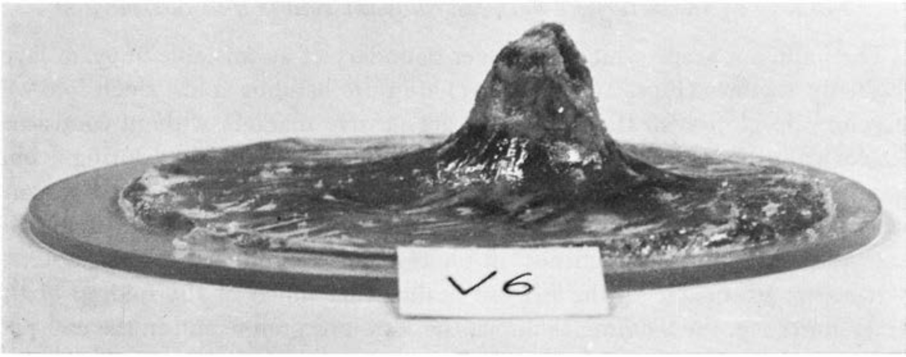


Fig. 23. Stitching-wax dome grown in overburden of putty, run in centrifuge. Overburden removed. Diameter of bottom plate 10 cm.

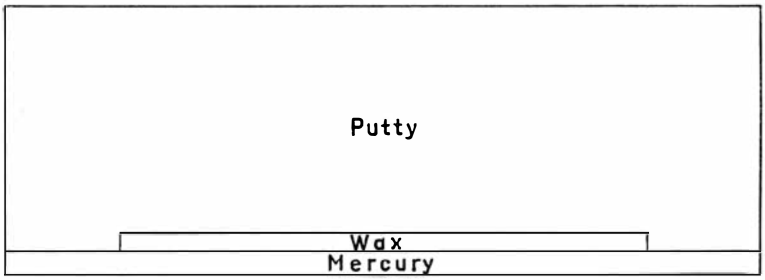


Fig. 24. Section through model V 6 before run.

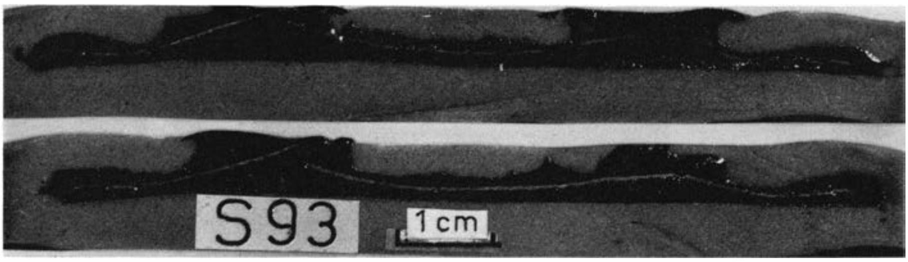


Fig. 25. Cross sections through model with domes of black silicone putty (density 1.38 g/cm^3) with thin sheet of rather stiff modelling clay. Overburden and substratum consist of painter's putty with density 1.87 g/cm^3. Run in centrifuge for 15 minutes at 1300*g*.

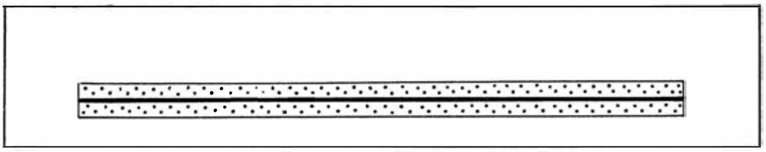


Fig. 26. Model S 93 prior to run in centrifuge. Dotted with black line: source layer of silicone putty with layer of rather stiff modelling clay; white: painter's putty.

Folding of the interface between buoyant source and overburden

The undulant shape which the upper boundary of an unstable buoyant layer gradually assumes (Figs. 21, 27 and 31) are pure bending folds. Such fold will therefore develop even if the layering are passive markers without contrasted rheological properties, Figs. 27 and 28. These folds readily form during evolution of layered models with unstable density stratification of wax, bouncing putty (or silicone putty), modelling clay, painter's putty and similar materials. Examples are found in practically all photographs reproduced in this paper.

It seems reasonable to the present author that much of the folding of the strata above e.g. the Tömmerås dome, the Tysfjord granite and in the east part of the Möre- and Namsos gneiss complexes, is produced by the same mechanism as in the models, i.e. by the active rise of the anticlines and the culminations.

More or less well developed marginal synclines adjacent to domes are produced in most models. The shape of these depressions may vary depending

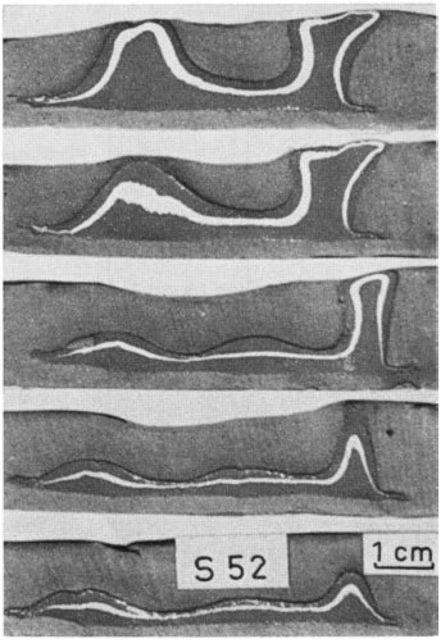


Fig. 27. Cross sections through centrifuged model of domes of silicone putty (grey, with smooth texture) with layer of white silicone putty developed in overburden of painter's putty.

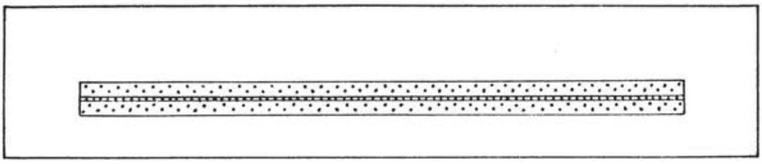


Fig. 28. Cross section through model S 52 prior to run in centrifuge. Dotted: red silicone putty, $\varrho = 1.14$ g/cm³; vertical lines: white silicone putty, $\varrho = 1.14$ g/cm³; white: painter's putty, $\varrho = 1.87$ g/cm³.

upon the rheological properties of the materials, but unless the overburden is too stiff for correct scale conditions (such that it fails to sag down in a potential syncline), marginal synclines are necessary complements to the domes. (See also PARKER *et al.*, 1955 and RAMBERG, 1963.)

The writer finds it natural to conceive of such Caledonian structures as the "folding trough" underneath the Jotun nappes south of the Möre culmination and the Trondheim synclinorium east of this culmination as rim synclines produced complementary to the ascent of the Möre bulge. Likewise the Bergen arc is a synclinal structure that possibly may be regarded as the marginal syncline to a basal culmination west of the arcs. In North Norway parts of the Nordland schists rest in marginal synclines possibly sagged down to compensate for the rising Nordland granites whose composite volume is quite considerable, see map. Fig. 1.

Since the flow in the source layer—i.e. in the sialic basement—below the rim syncline is more or less normal to the syncline one should expect to find elongation normal to the synclinal axis at these deep levels which unfortunately never are exposed. But even at the exposed contact between the culmination and the rim syncline the movement in the source and the drag at the contact should point in a general way toward the central part of the culmination. In this connection note e.g. the cross structures and cross axes recorded by WEGMANN (1959) at the eastern boundary of the Möre culmination. See also p. 14.

Buckle folding connected with the doming process

The convergent flow in the source layer mentioned above (p. 22) causes competent sheets enclosed in the source layer to buckle about axes pointing radially out from the root of a dome. Analysis of the flow lines shows that the fold axes bend around in the root zone of the dome to become vertical in the trunk of the dome. A large number of our models produced excellent example of this kind of buckle folds as shown in Figs. 29, 40, 44, and 45. As a matter of fact it is impossible to prevent such folds to form if sheets somewhat more competent (greater strength, higher viscosity) than the source layer are embedded in the latter. If the embedded sheets are too stiff, however, they do not buckle but instead may even hinder the evolution of domes when the buoyant force is not strong enough to break or bend the stiff inliers, see e.g. Fig. 25.

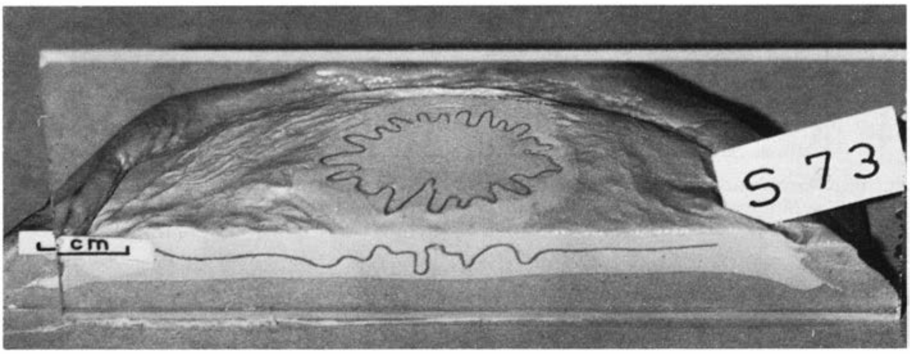


Fig. 29. Model S 73 after run in centrifuge. The model is cut in half, the putty overburden is removed to show the source layer with the buckled sheet of modelling clay, and the dome is cut horizontally about 7 mm below the original surface.

During the growth of domes flow also takes place in the overburden and in the substratum below the source layer. If the overburden and/or the substratum consist of material not much stiffer than the doming layer then the flow lines in a zone adjacent to the source layer are parallel to the flow lines in the latter layer. Consequently competent sheets enclosed in the overburden close to the source stratum or to the dome tend to buckle about fold axes parallel to those of the buckles within the source layer and the dome.

This is of obvious tectonic significance: note the numerous records to the effect that the fold pattern in a gneiss culmination is similar and parallel to the fold pattern in the adjacent schist mantle (e.g. for the Tömmerås dome, see PEACEY, 1964).

A general evolutionary pattern of growing domes is their tendency to spread to a funnel shape as they rise toward the free surface of the overburden. The reason for this is the lower confining pressure in the upper parts of the overburden. After piercing the surface nothing prevents the lateral motion, and the domes spread to flat horizontal cakes. This movement produces bending folds that often are recumbent and isoclinal, see Figs. 31 and 33. But the spreading of the domes also generates a horizontal component of compressive stress which may produce buckle folds of surface strata in front of the spreading dome. Such folds are often associated with, and parallel with, surface folds produced by sagging of the overburden into the marginal syncline. This structural feature is

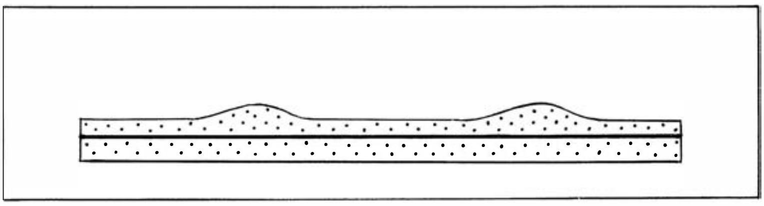


Fig. 30. Cross section of model S 73 before run in centrifuge. Dotted: red silicone putty with thin sheet of modelling clay; white: painter's putty.

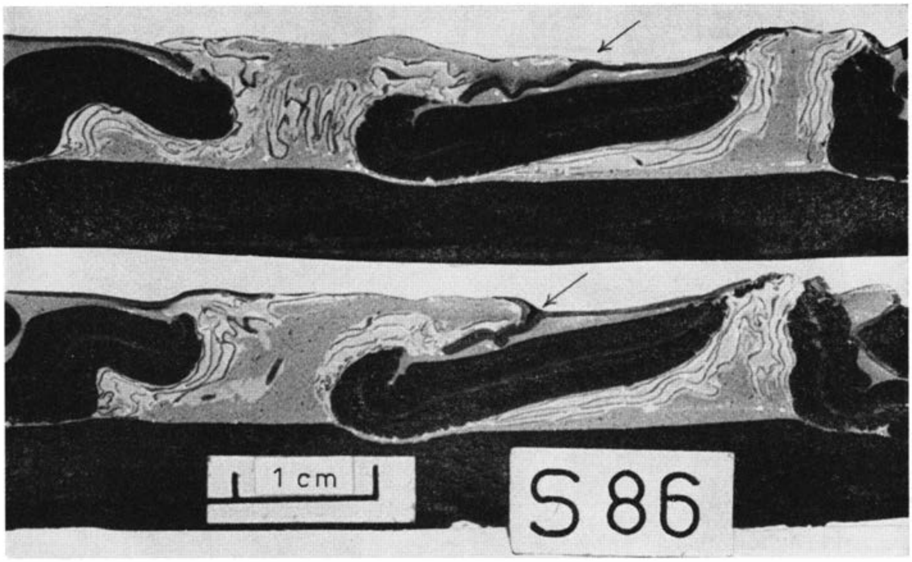


Fig. 31. Cross sections through model of silicone domes with thin sheets of modelling clay and overburden of painter's putty (black) after run in centrifuge for 15 minutes at 1300*g*. Note recumbent folds at arrows.

commonly developed in our centrifuged models, particularly in the region between domes which form close together, Figs. 33, 34, 35 and 62, p. 58.

It is worth noting that the axes of these surficial folds are normal to the axes of the folds produced within the source layer or deep in the overburden close to the source. The latter fold axes point toward the dome, the former are parallel to its circumference.

In conclusion we arrive at the following picture of the folds produced in the course of evolution of domal structures:

(1) The rise of the culminations and the subsidence of the adjacent marginal syncline constitute first-order bending folds of layers in the overburden and in the source stratum. The axial trend is parallel to the circumference of the dome.

(2) The convergent flow in the source stratum produces buckling folds in enclosed competent layers. The fold axes point toward the root of the dome and bend to become parallel to the axis of the trunk portion of the dome. Due to viscous drag second-order folds of layers within a contact zone in a soft overburden may be conformable to those in the adjacent part of the doming body.

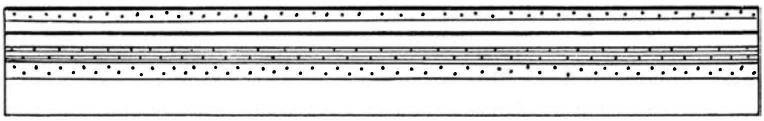


Fig. 32. Cross section of model S 86 before run in centrifuge. White: painter's putty; black: sheets of modelling clay; dotted with lines: silicone putty with thin sheets of modelling clay; dotted: silicone putty.

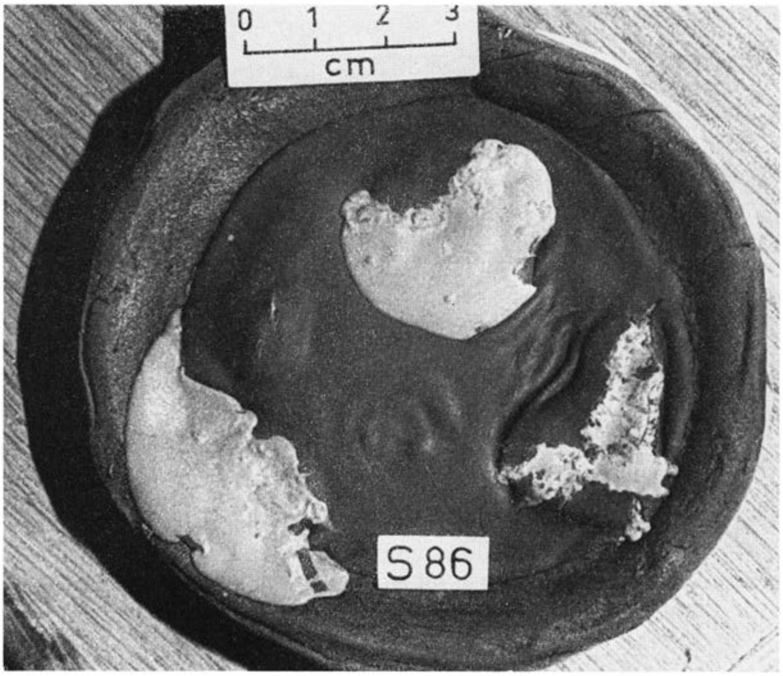


Fig. 33. Model S 86 after run in centrifuge as seen from above. Note pierced domes, tearing of sheet of modelling clay above domes and buckling in front of spreading domes.

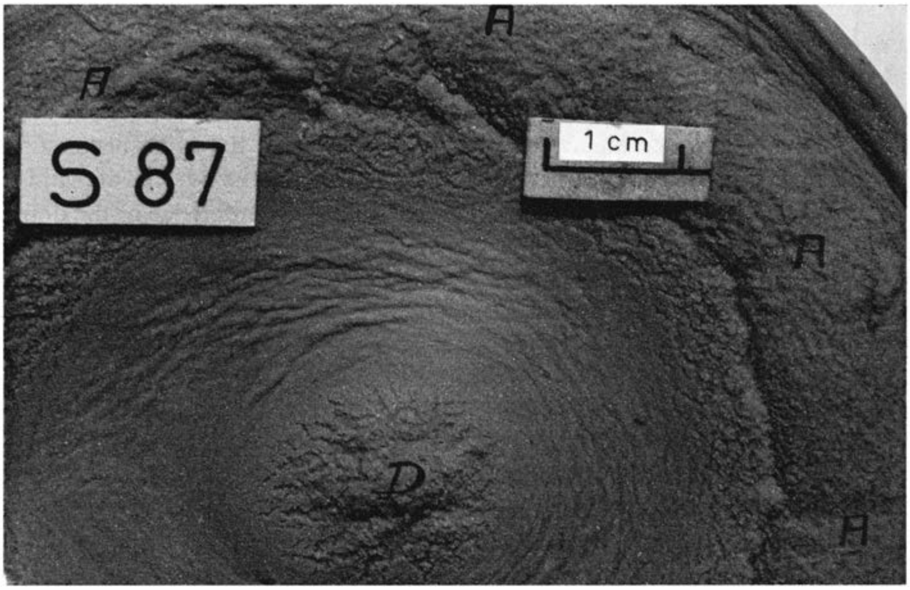


Fig. 34. Buckles of surface of putty overburden in syncline between a central dome (D) and anticline (A) along the edge of the circular model.

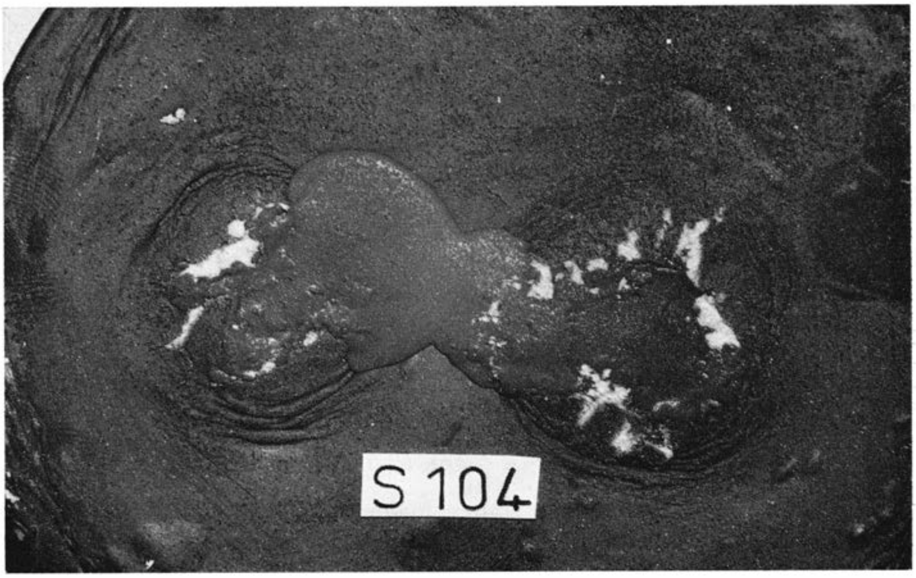


Fig. 35. Buckles and tensional tears in surface sheet of modelling clay above two domes of silicone putty run in centrifuge. Distance between centers of domes about 4 cm.

(3) The spreading of the upper part of a dome, especially of its piercing "hat", tends to produce buckling folds in surficial strata. Such folds have an axial trend parallel to the front of the laterally expanding lobes.

(4) Sagging of the overburden into the marginal syncline and sliding from the roof may also give rise to folds of the kind noted under point (3).

(5) Lateral spreading of the domes produces recumbent folds and Pennine-type nappe structures.

The Möre culmination and its huge rim syncline in the basement below the Trondheim schists and the Jotun nappes may be compared with the first-order bending folds of point (1) above.

The smaller-scale folds of the schists in the Trondheim synclinorium may be regarded partly as buckling folds formed during sagging in the marginal syncline (point 4 above). In part these folds may be due to the spreading of the Möre culmination and its pushing aside the surroundings during the rise.

Some of the folds along the eastern contact of the culmination with eastwardly plunging axes (WEGMANN, 1959), and the branch synclines of Cambro-Silurian (Devonian) strata that extend into the Möre complex from the west are possibly buckles produced by the convergent undercurrent of source material toward the huge basal culmination.

The klippen with inverted regional metamorphism on Hardanger vidda, the Jotun nappes and some of the nappes along the eastern margin of the Caledonides may well be eroded remnants of huge recumbent folds of type 5 above.

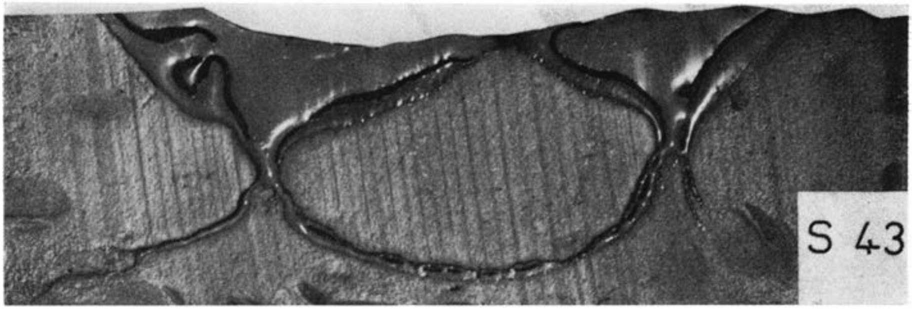


Fig. 36. Model of double-dome of silicone putty with boudins of modelling clay in overburden of painter's putty. Late stage of evolution.

Stretching, with consequent formation of pinch-and-swell and boudinage structures in connection with dome growth

In the convergent flow of the source material toward the root of a dome extension is parallel to the flow lines. Therefore elongations of mineral grains and other structural units must parallel the flow line in this particular geometry of flow (but not in all kinds of flow in tectonics; divergent flow—in which the flow lines spread apart—results in elongation normal to the flow lines), and embedded competent sheets or bodies are apt to fracture in tension normal to elongation, or to develop necked-down zones normal to this direction. In other words boudins and pinch-and-swell structure may develop under the proper relation between geometric dimensions and rheological properties. Such fragmentation of originally continuous relatively competent sheets embedded in dome-forming material was often produced in our experiments, both in the marginal syncline where the linear extension along the flow lines is at a maximum, as well as in the spreading top part of the domes where planar extension is at a maximum. Figs. 36 and 69 show examples from centrifuged models.

The boudinage- and pinch-and-swell structures so common in Vestranden studied by the writer (p. 11) is chiefly due to linear stretching parallel to the almost horizontal fold axes. We shall see below that this linear structure is probably developed in the source layer stretched between two super culminations connected by the strip of Vestranden gneisses.

When the doming material spreads on the free surface of the overburden the spreading mushroom is effected by planar extensive strain in the horizontal

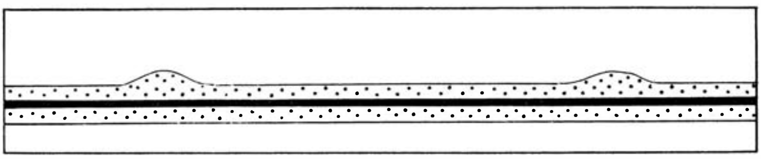


Fig. 37. Model S 43 before run. Dotted: silicone putty; black: modelling clay; white: painter's putty.

plane—i.e. extension occurs in all directions in the horizontal plane as demonstrated by the behavior of enclosed markers. In case the spreading takes place below a thin surface cover which is not pierced because of e.g. low density or excess strength, the cover becomes exposed to a system of planar tensile stresses that tend to break up the cover along tension fractures oriented in several directions. The lateral spreading leads to inversion of layers with recumbent folds and nappe structure, Figs. 54–58.

Interference of flow around two domes; evolution of Vestranden

Flow in source layer and domes

When two active domes protruding from the same layer are close enough for their source zone to overlap then the tangential compression and the radial stretching become accentuated in the region between the domes, Figs. 38 and 39.

This accentuation of the strain in the source layer in the region between two domes gives rise to especially intensive folding around radial axes in this region, and a correspondingly large axial stretching as visualized by the behavior of embedded competent sheets in our models as shown in Figs. 40 to 46.

The relationship between movement of a particle—e.g. tectonic transport of a small portion of a rock—and strain of the same particle shows some interesting

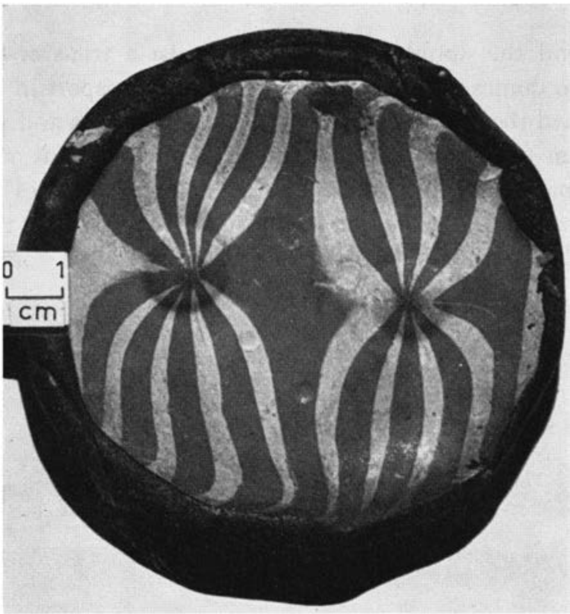


Fig. 38. Deformation of source layer around two domes. The figure shows the under side of a source layer of silicone putty consisting before run of straight strips of alternating light and dark colors. To prevent drag along the bottom the source layer was floating on mercury during run in centrifuge, and the deformation pattern shows how the material was sucked toward the roots of the two domes resulting in an intensive stretching in the region between the domes.

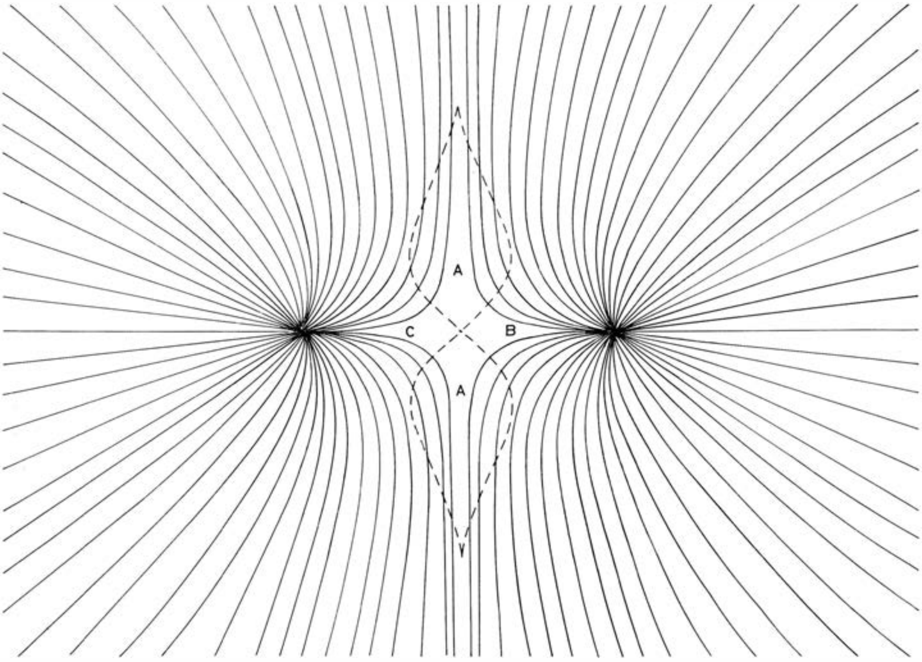


Fig. 39. Flow lines (particle paths) for flow in source layer toward two domes close enough for their flows to interfere, see text.

variations around the double-dome structure. In a transverse zone midway between the two domes the movement or tectonic transport in the source layer is directed toward the connecting line between the domes and about normal to that line (region *A*, Fig. 39). The flow is here divergent with consequent stretching normal to the flow lines. The result is *b*-tectonites with elongation and fold axis parallel to one another but normal to the tectonic transport. In the regions *B* and *C*, which still are in the region between the domes but closer to

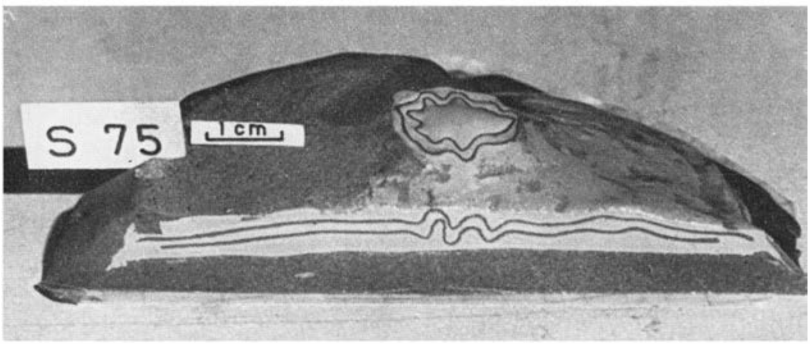


Fig. 40. Model S 75 cut in half after run for 2 minutes at 700*g* in centrifuge. The photo, which shows one of the two domes formed, shows buckles of the embedded sheets of modelling clay in the region between the domes and in the trunk of the dome which has been cut along a horizontal plane. Part of the overburden has been removed.

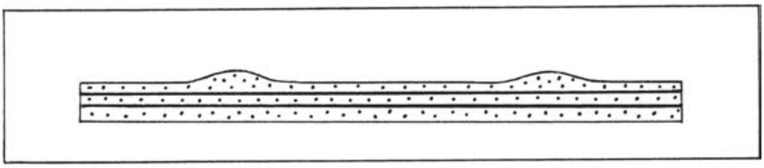


Fig. 41. Cross section of model S 75 before runi n centrifuge. Dotted: silicone putty, $\varrho = 1.14$ g/cm^3; white: painter's putty, $\varrho = 1.87$ g/cm^3; black: thin sheets of modelling clay.

one or the other, the movement is directed toward the nearest dome and the flow is strongly convergent. Fold axis and elongation (at any one given point) are also now parallel but both coincide with the direction of flow or tectonic transport in the buoyant source layer. The writer suggests that Vestranden has achieved its tight homoaxial folding and intensive axial stretching just because it is located in the region between two super domes, viz. the Möre- and the Namsos culminations, each of which probably is a composite culmination.

Though the folding in Vestranden could possibly be explained by an assumed regional Caledonian compression in NW–SE direction, the strong stretching parallel to the almost horizontal fold axes is hard to reconcile with such a model. The difficulties of explaining the commonly encountered lineation and elongation in deepseated crystalline schists parallel to the main trend of orogenic chains has long been recognized. It is hard to see how the entire mountain chain has been lengthened in the horizontal direction, yet such a result would follow if the local elongations are integrated along the whole chain. Some explanations have been offered in geologic literature, for example to the effect (1): that the mineral lineation, when parallel to the fold axis (so-called *b*-lineation), does not mean a geometric extension of the whole rock body, or (2): that the horizontal lengthening is taken up by the arced structure showed by some orogenic belts.

Without going into the argument here the author agrees completely with those workers who claim that preferred mineral lineation in one direction also means local lengthening of the rock in the same direction, quite independent of

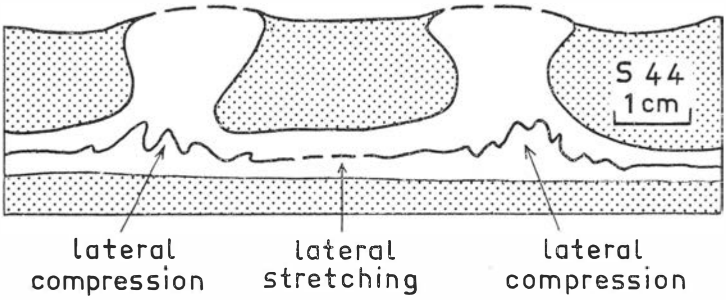


Fig. 42. Cross sections through centrifuged model of double domes of silicone putty (white) with embedded sheet of modelling clay (black) in overburden of painter's putty (dotted). Note intensive buckling of the sheet of modelling clay in the root zone of the domes, and the stretching and fracturing of the sheet in the region between the domes.

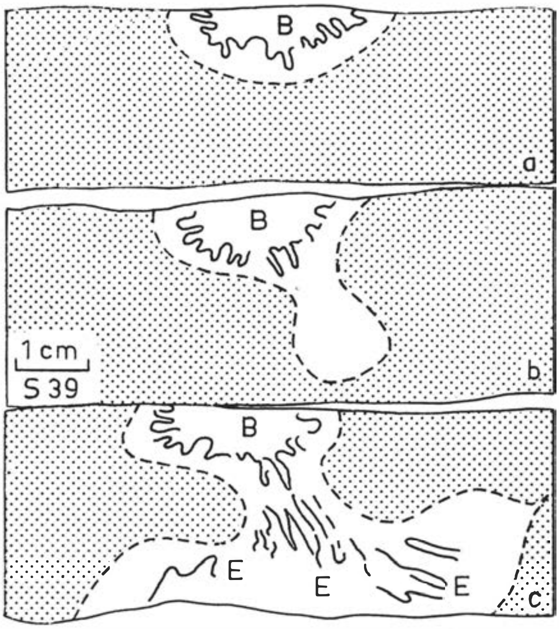


Fig. 43. Horizontal sections at different levels of a centrifuged model with two domes of silicone putty (white) with embedded thin sheet of modelling clay, in overburden of painter's putty (dotted). The sequence *a*→*c* shows increasing depth from a few mm below the surface (*a*) to the root zone with exposure of the ridge between the domes. The one dome (*B*) is cut in half by a vertical section (in order to study the profile), the other dome (*E*) has a wide, irregular shape in horizontal section. Note buckling with steep axes in section *a* whereas the fold axes are nearly horizontal in the ridge connecting the two domal culminations in section *c*.

the angular relation between the elongation and the flow direction (in convergent flow the elongation is apt to parallel the flow line, in divergent flow elongation makes an obtuse angle with the flow direction). In other words lineation is parallel to the long axis in the strain ellipsoid. Hence possibility (1) is out of the question.

With reference to possibility (2) one notes that the folds in Vestranden are unusually straight, striking ENE–WSW for a stretch of more than 200 km; bending in the horizontal plane does therefore not explain the horizontal stretching.

The writer believes that the model proposed above is realistic. It holds that the deepseated horizontal elongation and lineation represent a true stretching caused by nearly horizontal convergent flow, from a laterally distributed buoyant quartzo-felspathic source, toward domal culminations who are close enough for their source zones to overlap. In the Caledonides the Vestranden represents one example on deepseated stretching between domal structures. If several basal culminations line up along an orogenic chain, as they often do, it is easy to imagine that elongation parallel to the main trend be a common feature in deep sections of mountain chains, see e.g. Fig. 71, p. 68.

Deformation in overburden between domes

It is worth noting that the deformation in the upper part of the overburden between domes is quite opposite to the deformation and flow in the source layer and in the lower strata of the overburden. Partly because of the spreading tendency of the upper portion of domes (Fig. 35), partly because of the tendency of the overburden to sag down in the marginal syncline surface folds have generally concentric axes around domes. As the top strata of the overburden are pushed in front of the expanding lobes of the dome, or are sliding down into the marginal syncline, the concentric fold axes may become lengthened and *b*-type elongation could develop. Thus, the combined fold–elongation arrangement is reversed relative to the pattern in the source layer. This result from our model experiments is of considerable tectonic consequence and should be kept in mind when the structure of the basal culminations and their surroundings is studied. It is believed by the author that many of the Nordland granites may represent the spreading and overturned portion of domes protruding from a central swell of the sialic basement of the Caledonides. One may therefore expect to find concentric fold axes in the strata around some of these plutonic bodies. Unfortunately not enough detailed structural field work has as yet been done for a test of this possibility.

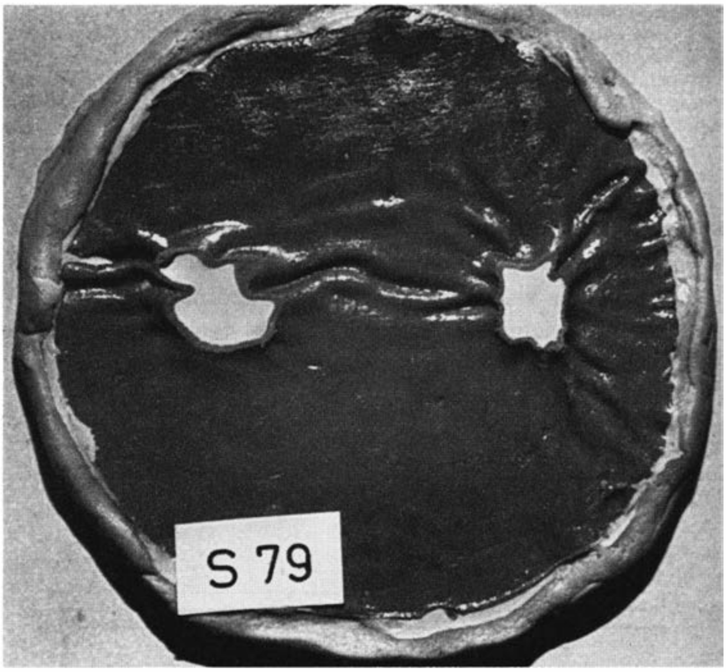


Fig. 44. Centrifuged model of two domes of silicone putty (white) with thin top sheet of modelling clay (dark grey) rising through an overburden of painter's putty (removed in photo). Note buckling of modelling-clay sheet around and between domes.

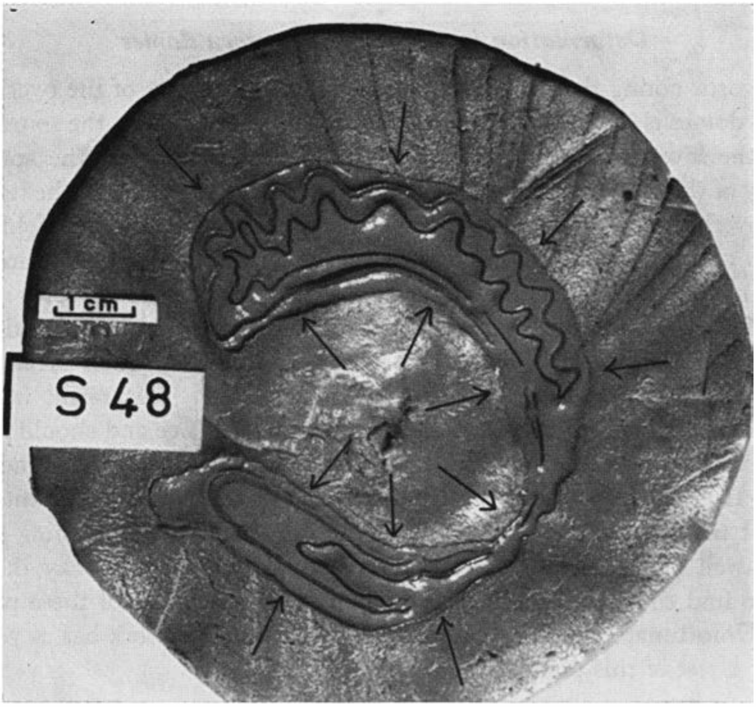


Fig. 45. Horizontal cut through circular ridge of silicone putty with two embedded sheets of modelling clay risen through overburden of painter's putty. Run at 750*g* for 4 minutes. During the rise of the buoyant ridge flow in the source layer was centripetal toward the ridge from the edge of the model, and centrifugal from the center of the model toward the ridge. That is the former flow was convergent, the latter divergent (arrows). In consequence stretching (boudins!) of the modelling-clay sheets occurred in the concave part of the ridge and compression (buckling) occurred in the convex part of the circular ridge.

As noted above (p. 36) radial stretching also occurs in connection with the piercement of domes. The spreading dome itself is being stretched laterally, a process which is made visible by for example boudinage structure and tension cracks in competent inclusions, and by vertical tension fractures in several directions in the cover, Fig. 48. Downslope sliding of sheets riding on top of the dome may also give rise to radial stretching and elongation depending upon the

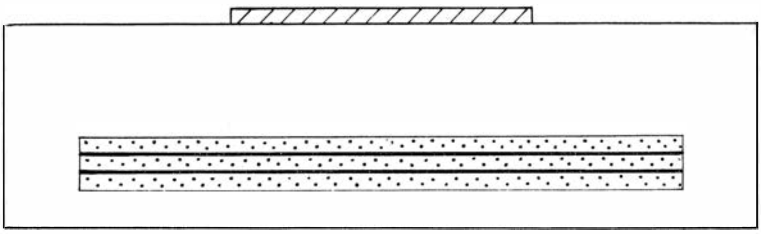


Fig. 46. Model S 48 before run. Dotted: source layer of silicone with sheets of modelling clay embedded; white: painter's putty; inclined hatching: top layer of modelling clay.

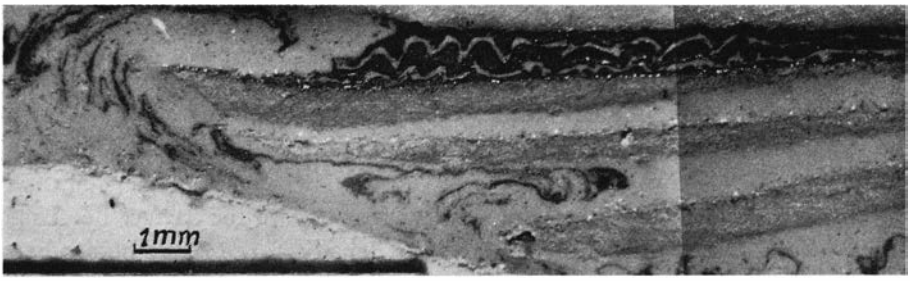


Fig. 47. Buckling of "sedimentary" strata in front of a spreading dome in model S 120. Light uniform grey with dark "sliren": silicone putty; darker grey with granular texture: overburden of painter's putty; black with buckled sheets of white material: dark silicone putty with sheets of modelling clay. The model is similar to S 121, Fig. 65.

particular circumstances such as the topography of the surface. (Note e.g. the variation of the angular relation between elongation—respective compression—and direction of flow in a piedmont glacier: On the sloping surface the elongation is parallel to the flow lines, on the piedmont flat elongation is parallel to the advancing front and thus normal to the flow lines as demonstrated experimentally by the author, 1964*b*.)

We shall return to a more detailed study of the dynamic relationships of piercement domes and the overburden in connection with the development of nappes.

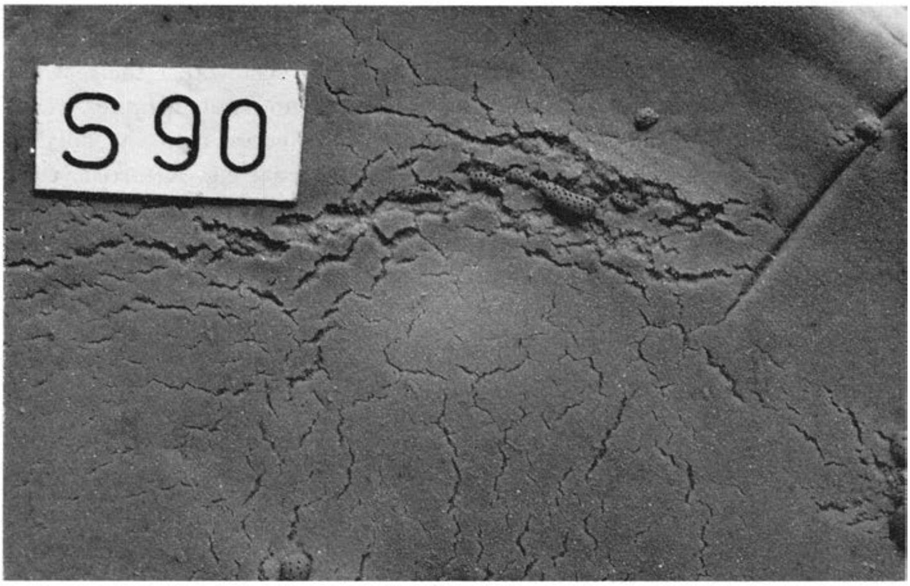


Fig. 48. Tension cracks in overburden of putty through which domes of silicone putty are piercing (dotted). White label about 1.5 cm long.

Concluding remarks on the dynamics of basal culminations as buoyancy phenomena

Based on the experimental studies reported above the following model for the evolution of the basal gneissic or granitic culminations is offered for considerations.

The culminations are portions of sial that rose toward the surface because of buoyant forces. The buoyancy is due to the density of the rising bodies being less than that of the average crustal rock. In this connection it is significant that experiments show that doming bodies need not exhibit lower density than the adjacent material at a high horizontal section through the system, see Figs. 49, 50 and 51. The buoyant force may also arise from density contrasts at deeper levels. It is not yet known whether the lavas and metamorphosed sediments of the Caledonides were sufficiently dense to give buoyancy to the submerged sialic basement, or whether large masses of heavy basic rocks once occurred in the geosyncline, and that they were the masses which made the light sialic material buoyant. The energy increment represented by the elevation of the culmination would then be balanced by the energy decrease represented by the sinking heavy masses.

If such masses existed in or below the Caledonian syncline they may at present be found deep below the marginal synclines—e.g. below the Jotun nappes and below the Trondheim synclinorium, and/or west of the culminations, off the coast of Norway, see similar situations in some of the experiments, e.g. Figs. 54 to 60.

One also notes that increasing temperature toward the depth makes a system potentially convective even if the system is not otherwise unstably stratified. Even in a chemically and mineralogically uniform sialic crust thermal convection would be possible provided the crust does not show a finite strength because the geothermal gradient is substantially steeper than the adiabatic gradient. (Concerning terrestrial convection currents see e.g. KNOPOFF, 1964).

Moreover, at deep crustal or subcrustal levels partial melting may occur, a process which decreases the density by up to 10 % and consequently supplies a strong buoyant force to the system.

Fig. 49. Domes of silicone putty (light and darker grey), $\varrho = 1.14$ g/cm^3, which have risen through three layers of black silicone putty, $\varrho = 1.35$ g/cm^3, see Fig. 50.

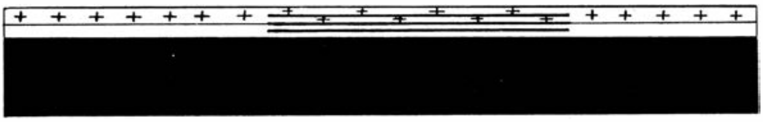


Fig. 50. Model S 122 prior to run in centrifuge. Black: black silicone putty, $\varrho = 1.35$ g/cm^3; white: red silicone putty, $\varrho = 1.14$ g/cm^3; crosses: pink silicone putty, $\varrho = 1.14$ g/cm^3.

It is suggested that all or some of these conditions act together so as to make the sialic basement buoyant under the cover of geosynclinal strata.

Our experiments show that unstable crustal stratification results in elongate anticlinal ridges whose axes conform to the boundary of the overburden or/and to the edge of the source layer. Actually any linear discontinuity in source layer or overburden, such as the edge of an excess mass on or in the overburden, is apt to control the shape and orientation of the domes. So is a variation in thickness of overburden and/or source layer. If thus the thickness has a gradient along the layering then buoyant anticlines or rows of more equidimensional domes are apt to develop with their trend normal to the thickness gradient. It is significant that anisotropic lateral stress is not necessary to produce elongate culminations, an experimental observation also of great consequence for salt

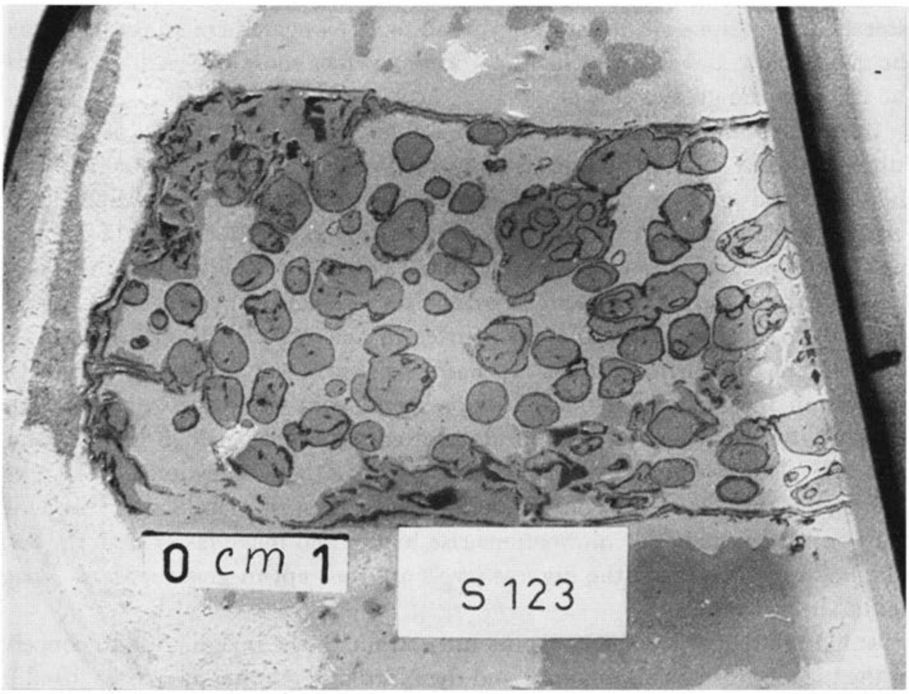


Fig. 51. Model practically identical to S 122 seen from above after run for 10 minutes at 2400*g*. Dark grey: domes of red silicone putty, $\varrho = 1.14$ g/cm^3; light grey: overburden of pink silicone, $\varrho = 1.14$ g/cm^3. Dark mantle around domes is remnant of black deep-seated overburden with density 1.35 g/cm^3, *i.e.* the buoyancy-generating overburden, see Fig. 50.

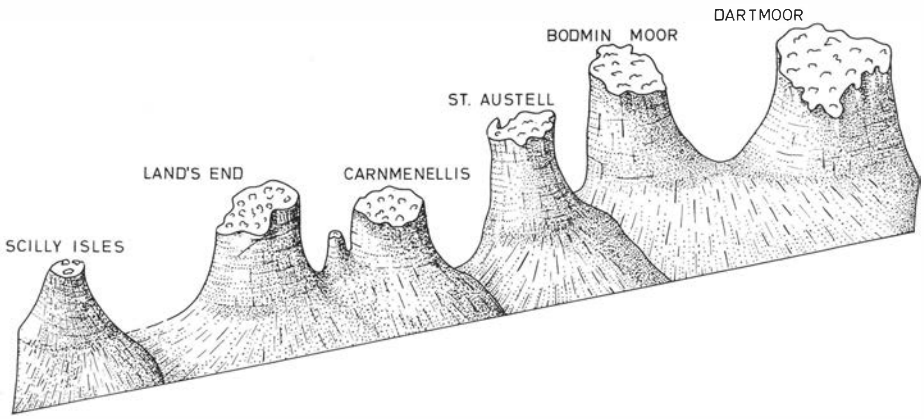


Fig. 52. Schematic indication of the row of granite intrusions between Scilly Isles and Dartmoor, England. After HOLMES, 1965, p. 279.

tectonics. So-called salt walls or salt ridges are generally thought of as formed in response to unidirected lateral compression.

The edge of a geosyncline is a curvilinear discontinuity in the crust and the thickness gradient of the crust and its sedimentary cover in general points normal to the geosynclinal axis. Therefore basal culminations rising, not due to lateral compression, but to buoyancy, tend to be elongate parallel to the edge of the geosyncline and they are likely to occur in rows more or less lined up with the trend of the geosyncline.

The present author feels that the general geometric distribution of the basal culminations in the Caledonides is controlled by the conditions noted above. If the Nordland granites are included amongst the basal culminations (note remarks on the Svartisen granite, p. 6), possibly as protuberances of the most mobile components from a continuous basal geanticline below, then there is no question about the arrangement of basal rises parallel to the orogen. (Compare the row of granites from the Scilly Isles to Dartmoor in England, Fig. 52.) Some of the Nordland granites are more homogeneous and massive than the gneisses in the Grong–Vestranden–Möre complexes—this is e.g. true for the Binndal granite. Many of the Nordland granites appear also to have invaded higher levels in the geosynclinal column than have the unquestioned basal complexes. But this fits well in our model because the most mobile, and less dense components in the basal culminations rise higher and move faster than the rest. The bodies which reach the highest levels are thus apt to give the most "magmatic" impression.

In light of our experiments it puts little strain on the imagination to conceive of the plastic central culminations and the Nordland granites as masses lifted by buoyant forces. It is perhaps more difficult to assume a similar dynamic history for the marginal culminations or basement windows closer to the eastern edge of the mountain range—the row of windows along the southeast and east border

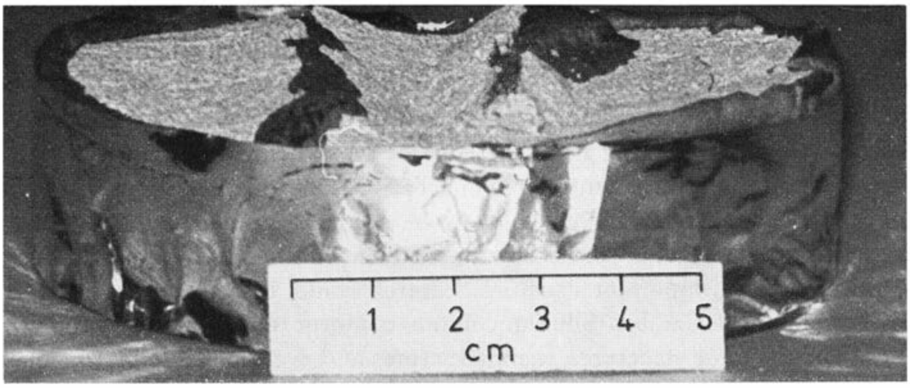


Fig. 53. Two anticline ridges of stitching wax protruded through an overburden of painter's putty with sheet of concrete on the surface.

of the Trondheim synclinorium between Atnasjö in the south and Olden in the north. These exposures of basement rocks show no or but negligible sign of recrystallization and plastic flow, though mylonitization and dynamometamorphism are often considerable. If they represent locally lifted portions of the basement—such as demonstrated by LJUNGNER (1950) east of the Lönsdal–Nasafjell window—rather than original bulges on the geosynclinal floor the movement must have occurred by means of fracturing and faulting. But in this connection we refer to the model experiments in which brittle sheets were placed on the top of the overburden (Fig. 53) and stiff sheets placed immediately above the layer of plastic source. Such sheets became faulted and broken above doming protuberances from a plastic source layer. In view of these results the author does not at all find it unlikely that the cold, brittle top part of the pre-Cambrian floor of the Cambro-Silurian sedimentary and volcanic column would yield to local buoyant forces by faulting. See also Fig. 19.

The Tömmerås dome recently studied quite thoroughly by PEACEY (1964) furnishes an example on a culmination which has been elevated by a combination of faulting and plastic flow. "... the faults are arranged around the Tömmerås block, and, that in nearly every case, the displacement on them is such that the massif is raised at the center relative to the sides". (PEACEY, op. cit. p. 75). Yet the plastic deformation structure or the recrystallization structure due to regional Caledonian metamorphism is conformable in basement and Cambro-Silurian cover. The basal gneiss occurs as a conformable core in a gentle elongate dome.

Beside having been raised in Caledonian time the structure appears also to have existed as a bulge on the geosynclinal floor already during the sedimentation of the basal strata of quartzite and arkose according to PEACEY'S interpretation. But such an initial bulge tends also very definitely to localize an active domal rise in a gravitational unstable situation as demonstrated by a large number of our experiments.

The volumes of the basal culminations and the Nordland granites add up to considerable magnitudes. When these bodies rose substantial lateral flow in the source layer outside the roots of the domes must have occurred. This flow was converging toward the culminations, that is the flow lines in the deepseated source layer were pointing from the edge of the syncline toward the rows of domes, and from some neutral regions between neighboring domes toward either of the latter, see Fig. 39. As lineation or structural elongation as well as the fold axes are parallel to the flow lines in this kind of convergent flow (see p. 37) these two important structural features would, in the basement and the basal strata of the Cambro-Silurian column, conform to the pattern of the flow lines. That is, cross structures, main structures and every possible intermediate attitude would be produced. But because of the control over the geometrical arrangement of the culminations by the pattern of the geosyncline, two preferred directions will prevail, viz. one parallel to the geosynclinal axes (i.e. parallel to the row of culminations) and one normal to the geosyncline (i.e. parallel to the direction of the undercurrent from the edge of the geosyncline to the first row of culminations, and further between the rows of culminations).

This pattern of lineation or elongation and folds is characteristic for the source layer. The source for the rising culminations is in the first place the sialic part of the basement of the geosyncline, but conceivably also the mobilized arkosic sediments (e.g. the sparagmites) and acidic lava. However, flow would be significant only below certain depths where temperature is sufficient to cause recrystallization and plastic mobility. The source has to be located at or below this level.

Now, the source layer is not often exposed outside the culminations except as ridges between some culminations or ridges otherwise radiating out from culminations, formed in part because of buckling about radial axes under convergent flow toward the culminations. There is therefore limited possibilities for testing some of the experimental–theoretical predictions above. However, some checks can be made in connection with the Möre–Vestranden–Namsos complexes. Firstly the Vestranden is a sort of connecting ridge between the two culminations, secondly the cross anticline with basal core at Vågå is an extension from the Möre culmination into the source layer and toward the eastern edge of the geosyncline. The already mentioned NE–SW elongation and fold axes in Vestranden and the strong E–W striking elongation and fold axes in the Vågå ridge both fit well into the model pattern.

The consequence of this picture is that the rise of the basal culminations in the central parts of the Caledonian orogen was associated with a plastic undercurrent from the southeastern edge of the geosyncline, and probably also from the western edge which is not exposed. Similar undercurrents are postulated in other orogenetic theories such as the convection-cell theory. Differences and similarities between the two theories are discussed in a later section, p. 60.

Similarity between nappe structures in orogens and experimental models of spreading domes

It is a remarkable similarity between the pattern of the spreading domes and anticlines in our experimental models and alpine-type nappes, especially the Pennine nappes with their core of deformed and recrystallized basal gneiss (Figs. 54 to 60). The experimentally produced spreading domes are often developed as recumbent folds with inversion of strata and nearly horizontal axial plane, just like some of the most well-developed nappes in the Western Alps. Such striking similarity of geometry makes it tempting to suggest an equally close similarity of dynamics of the two systems, in other words that the acting forces and the movements are similar in model and in the natural structure. In this connection it is worth realizing that the only driving force operating in the models is the centrifugal force which plays the same role in the models as does gravity in natural tectonic systems.

In the Scandinavian Caledonides the more or less well studied nappes along the eastern and southeastern edge of the orogen seem to be thrust sheets rather than recumbent folds of Pennine type. This impression acquired through field studies may, however, in part be due to the deep erosion of this old mountain chain as compared with the Alps. The upper limb of recumbent folds of nappe dimension is easily lost by erosion. Incidentally, our experiments show that intensive stretching of the upper limb usually occurs in nappe-like recumbent folds (=advancing lobes of spreading domal structures), see e.g. Figs. 31, 33, 54 and 55. The upper limb is therefore often torn and the pieces separated by considerable distances. Moreover, deep erosion combined with metamorphic recrystallization makes it difficult to unravel the true structure within such regions as the Trondheim- and the Nordland synclinoria, this is particularly true if the natural orogenic structures are as complex as some of our model structures, cf. Figs. 54 to 60. Recent works in the Oppdal district at the eastern edge of the Möre culmination (Holmsen, 1960) have shown the presence of complex recumbent folds with cores of the basal gneiss. The nappe structures in the Möre Culmination as reported by Muret (1960) are also highly interesting, Fig. 61.

In this connection it may not be out of place to mention the large region with nearly horizontal axial-plane schistosity in the low-grade schists between Malvik and Levanger on the southeast side of Trondheim fjord. Here interlayered strata of shale, sandstone, limestone, conglomerate and greenstone of the Hovin-(flysch) and Stören (ophiolites) groups exhibit numerous small- and medium-scale folds. (That is folds easily recognized in single outcrops without mapping.) Throughout most of this region the axial plane, which coincides in a general way with the schistosity, shows but gentle dip. It seems not unlikely, particularly in view of the apparent inversion of the Stören–Hovin groups at Forborfjell and Foldsjö (Carstens, 1960), that these smaller folds are second-

Fig. 54. Sections through model of domes of black, white and dark grey silicone putty penetrating overburden of painter's putty (inclined hatching), and top layer of white and red silicone with thin sheet of modelling clay. Run for 11 minutes at 2000g. See Figs. 55 and 60.

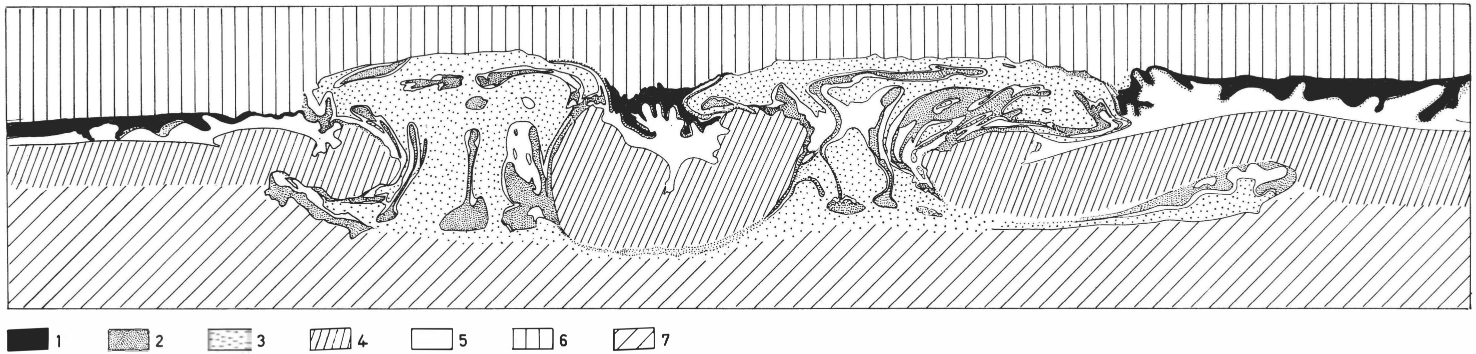


Fig. 55. Trace drawing of model S 112 as shown in the upper section on photo, Fig. 54. 1: red silicone putty, $\varrho = 1.14$ g/cm³; 2: black silicone putty, $\varrho = 1.35$ g/cm³; 3: grey silicone putty, $\varrho = 1.25$ g/cm³; 4: painter's putty, $\varrho = 1.87$ g/cm³; 5: white silicone putty, $\varrho = 1.14$ g/cm³ ;6: oil-wax mixture, $\varrho = 0.9$ g/cm³; 7: painter's putty, $\varrho = 1.87$ g/cm³. Thin sheet of modelling clay causes buckles in surficial strata.

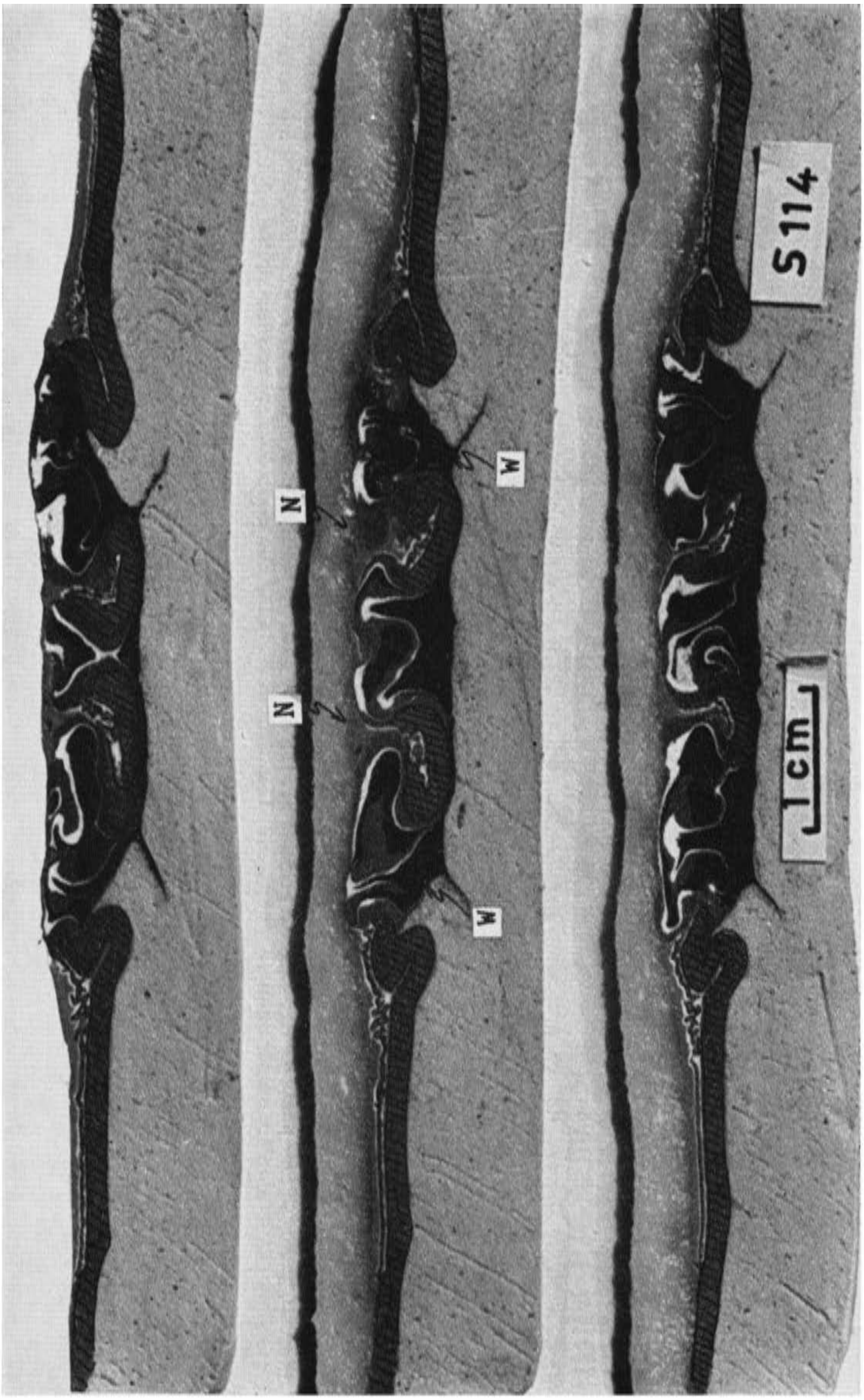


Fig. 56. Sections through model S 114 of domes and nappes of grey, white and black silicone risen through overburden of painter's putty (inclined hatching) and thin "sedimentary" layers of silicone and modelling clay under cover of soft wax. Run for 8 minutes between 1300g and 2200g. N: "Narben" zones; W: "Wurzel" zones. See Figs. 57 and 60.

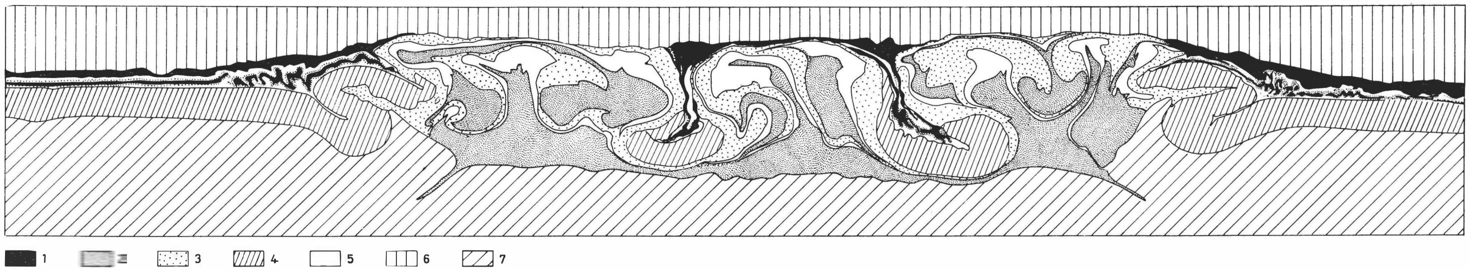


Fig. 57. Trace drawing of model S 114 as shown in lower section on photo, Fig. 56. 1: red silicone putty, $\varrho = 1.14$ g/cm³; 2: black silicone putty, putty, $\varrho = 1.35$ g/cm³; 3: grey silicone $\varrho = 1.25$ g/cm³; 4: painter's putty, $\varrho = 1.87$ g/cm³; 5: white silicone putty, $\varrho = 1.14$ g/cm³; 6: oil-wax mixture, $\varrho = 0.9$ g/cm³; 7: painter's putty, $\varrho = 1.87$ g/cm³. Thin sheets of modelling clay cause buckles in surficial strata.

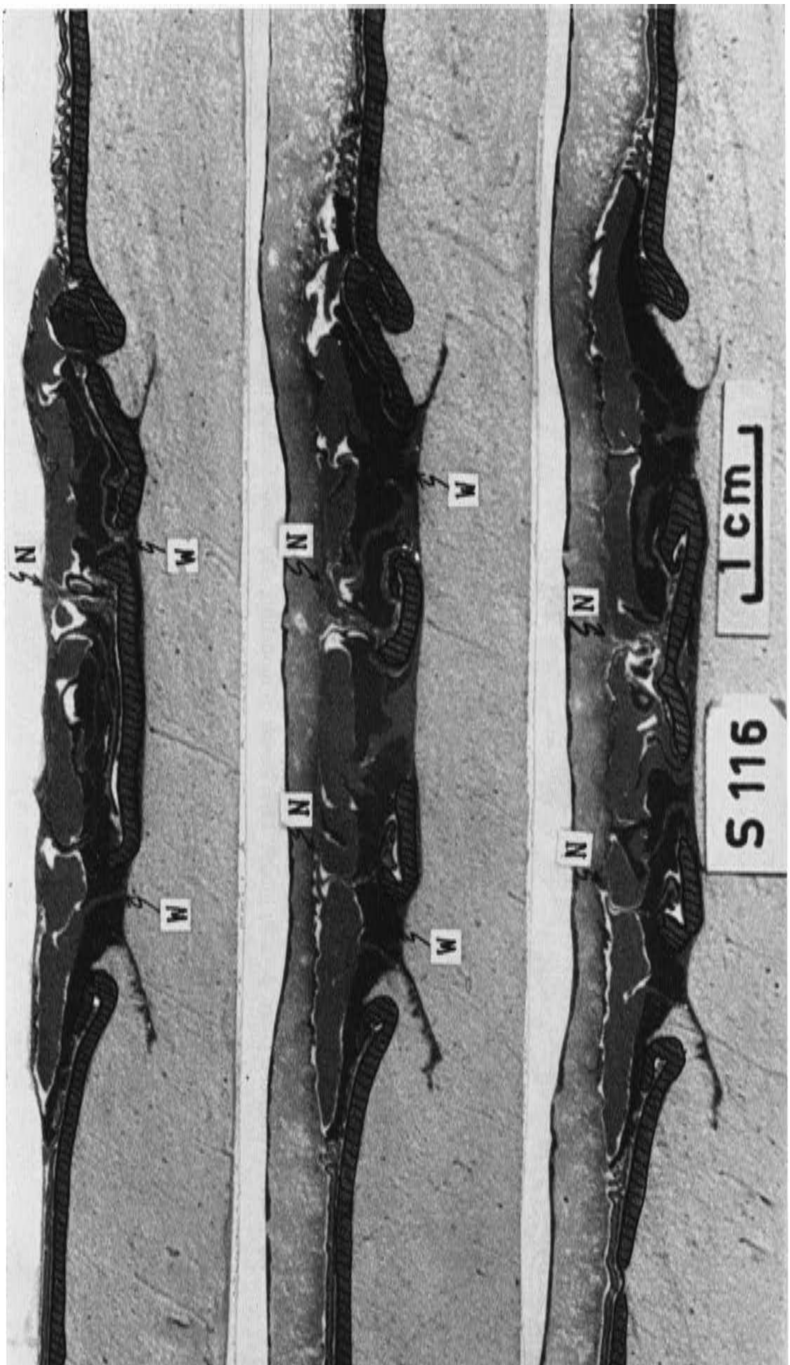


Fig. 58. Sections through model S 116 which is similar to model S 114, Fig. 56, but in S 116 the dark and light grey silicone is inverted (compare Figs. 59 and 60), and the overburden of putty (inclined hatching) is less thick. Run for 15 minutes between 2400*g* and 2900*g*. N: "Narben" zones; W: "Wurzel" zones.

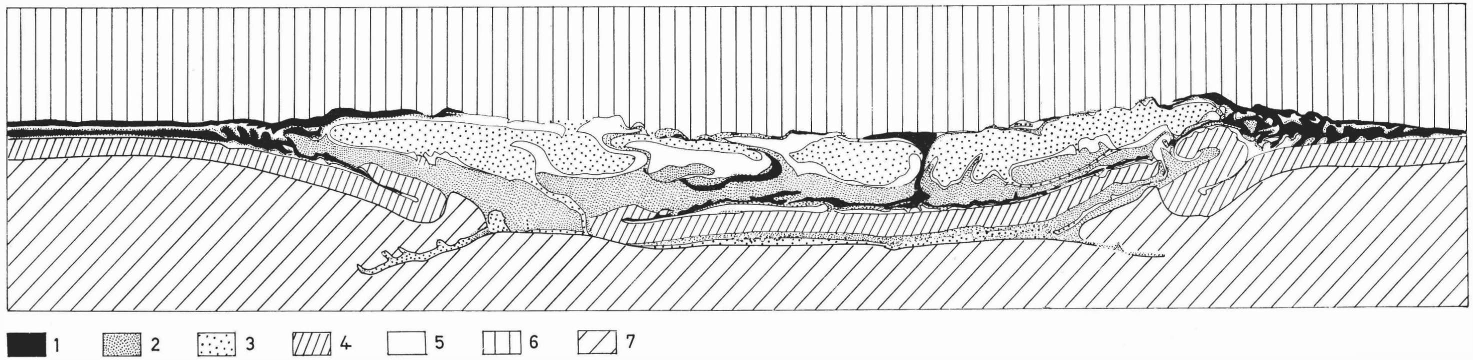


Fig. 59. Trace drawing of section of model S 116. The drawing refers to a profile slightly different from those shown in photo, Fig. 58. 1: red silicone putty, $\varrho = 1.14$ g/cm³; 2: black silicone putty, $\varrho = 1.35$ g/cm³; 3: grey silicone putty, $\varrho = 1.25$ g/cm³; 4: painter's putty, $\varrho = 1.87$ g/cm³; 5: white silicone putty, $\varrho = 1.14$ g/cm³; 6: oil-wax mixture, $\varrho = 0.9$ g/cm³; 7: painter's putty, $\varrho = 1.87$ g/cm³. Thin sheets of modelling clay cause buckles in surficial layer, see also Fig. 62.

order folds on a huge recumbent nappe-like structure. The nearly horizontal axial-plane schistosity signifies a compression in vertical direction such as would happen when a large recumbent fold spreads under its own weight.

In this connection one notes that ophiolitic nappes are a characteristic feature of fold mountains according to AUBOUIN (1965) who points to overfolded or overthrust nappes of ophiolites in the internal regions of the Alps, the Apennines, the Hellenides and the orogens of the Sunda Islands.

It is interesting how some significant detail structures of orogens have evolved in the models without the author's intention. For example the steeply downpulled "Narben" zones (KRAUS, 1951, p. 65) and the equally steeply risen "Wurzel" zones (KRAUS, op. cit. p. 65) are clearly developed in the models shown in Figs. 54, 56 and 58, *N* and *W*.

The structure of some of the models would be strikingly similar to Caledonian thrust sheets if the upper half of the spreading domes, overriding as they are the doubled-up strata, was cut away, Figs. 56 and 58. One notes for example the inversion of layering in the sense that the deeper part of the original source layer has become the top part in the overturned anticline or dome, and that the spreading part of the anticline or dome is overriding the younger sediments along the edge of the geosyncline, Fig. 62.

In nature this would correspond to inversion of regional metamorphism such as found in many nappe localities in Sweden (e.g. KAUTSKY, 1952) and Norway, the gneiss-capped schist nappes or klippen in the Hardangervidda region being traditional examples, see e.g. HOLTEDAHL *et al.* 1960, pp. 200–202. In this connection one may recall the extensive zone of inverted regional metamorphism in the lower Himalayas below the overthrust vast masses of the crystallines of the higher Himalayas (see Fig. 145 in GANSSER, 1965).

The folds in the sedimentary cover in front of the nappe in model S 116 (Fig. 62) compare well with the folds described by SKJESETH (1963) from the Cambro-Silurian sediments in the Mjösa district, Fig. 63.

It is of course one of the most generally accepted views among students of orogenic tectonics that the lateral movement of nappe-type recumbent folds is due to gravitational forces. Sliding down-hill from e.g. geanticlines and gravitational spreading are the chief types of movements in this so-called secondary tectonics. But the upheaval of the culminations and the piling-up of the thick packages of strata needed for sliding and spreading are phenomena generally assumed to derive from lateral compression of the geosyncline between the jaws of the adjacent cratogens, the necessary lateral stress arising for example from subcrustal convection currents.

The picture arising from our model experiments is different in one important respect; the very rise of the culminations and anticlines which are to become the core of the nappes—not only their lateral creep and folding—is caused by the body force of gravity acting on an unstably stratified crustal region. Other authors have expressed similar views, particularly VAN BEMMELEN in a number

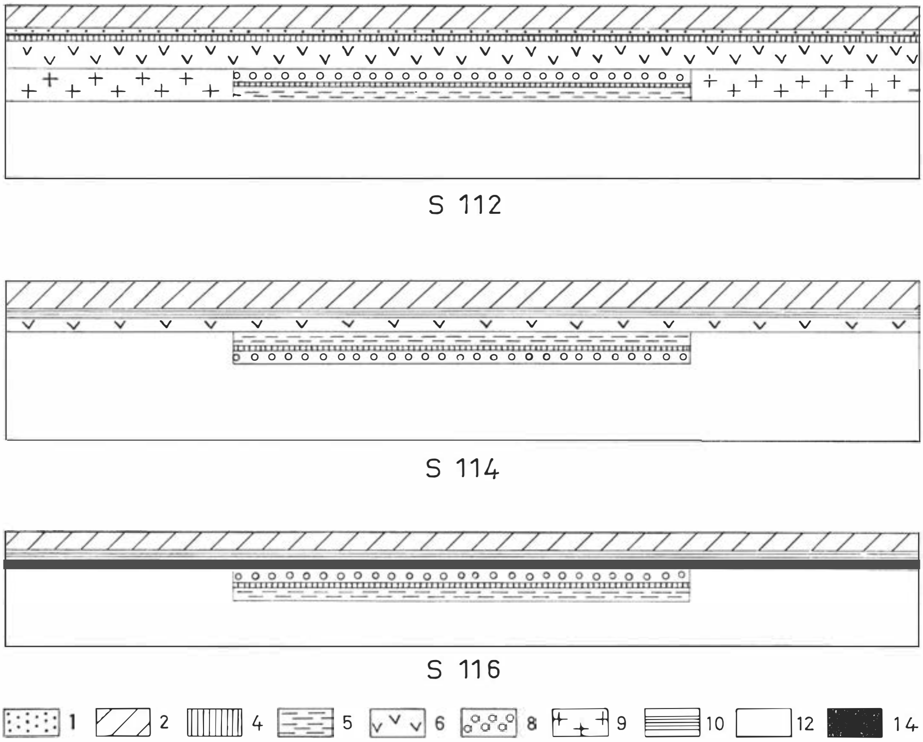


Fig. 60. Cross sections through models S 112, S 114 and S 116 prior to run in the centrifuge. See Figs. 54 to 59. 1: red silicone putty, $\varrho = 1.14$ g/cm^3; 2: oil-wax mixture, $\varrho = 0.9$ g/cm^3; 4: white silicone putty, $\varrho = 1.14$ g/cm^3; 5: grey silicone putty, $\varrho = 1.25$ g/cm^3; 6: painter's putty, $\varrho = 1.87$ g/cm^3; 8: black silicone putty, $\varrho = 1.35$ g/cm^3; 9: painter's putty, $\varrho = 1.87$ g/cm^3; 10: silicone with thin layers of modelling clay; 12: painter's putty, $\varrho = 1.87$ g/cm^3; 14: thin sheet of modelling clay in S 112.

of thought-provoking articles (van Bemmelen, 1960), but the experimental support has been lacking.

According to the model experiments on which this paper is based the total geosyncline needs not be compressed, instead the local compression between or adjacent to culminations may be compensated by the lateral expansion produced by the domal upwelling of the unstable basement. Moreover, the buckling due to sagging of strata down in rim synclines and due to the push in front of lobes of piercement domes and overturned anticlines (see e.g. Fig. 62) easily gives an illusion of a net compression of the geosyncline with its stratified content between the jaws of adjacent cratogens. But our models were always run in a rigid container whose boundaries were fixed and did not permit any net shortening in horizontal direction. The folds were either simply bending folds due to alternating rise and subsidence of stratified structures, or local compression folds associated with lateral widening of adjacent regions.

With special reference to the Scandinavian Caledonides the widespread so-

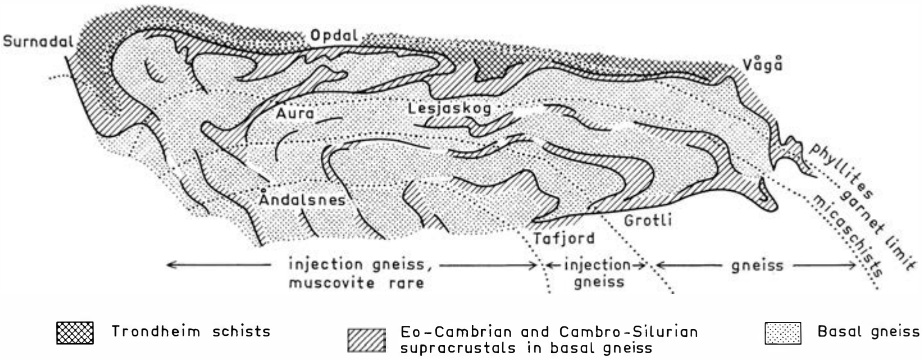


Fig. 61. Profile across the eastern region of the Möre culmination as interpreted by MURET, 1960.

called cross folds with axes making obtuse angles with the main trend of the orogen are not easy to explain by a general compressive stress normal to the geosyncline. In the model advocated in this account local variations of fold axes, of elongation and of other structural properties indicative of movement and strain, are necessary details of the complete picture. Just as the model calls for structural variations in space along the Caledonian chain so it calls for variations in time. The model is not consistent with Caledonian-wide tectonic phases—pulsations—which supposedly have occurred simultaneously throughout the entire orogen, but predicts instead that smaller or larger parts of the chain may be tectonically active while other parts are inactive, and that the latter may awake to activity while the former assume a state of relaxation.

Our experiments show, however, that in comparatively homogeneous models

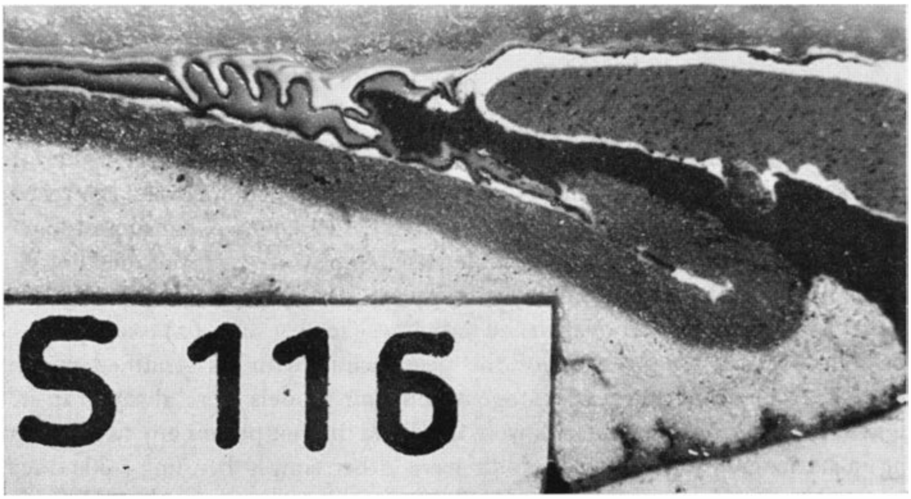


Fig. 62. Enlarged edge of model S 116 showing front of pennine-type nappe with buckling of "sediments" and inversion of layering. Note e.g. detachment folds of upper thin sheet of modelling clay in front of nappe.

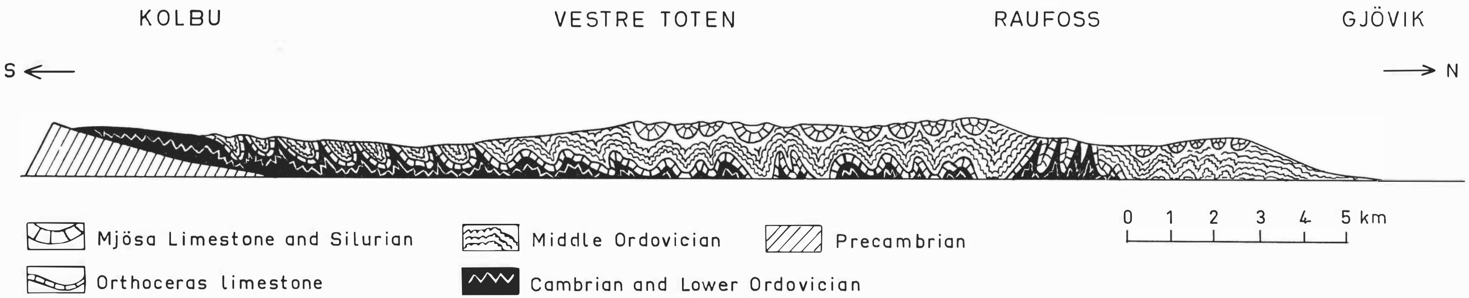


Fig. 63. Profile of sediments folded in Caledonian time at the peripheral southeastern part of the orogen. After SKJESETH, 1963.

—i.e. models with rather uniform overburden and uniform source layer—the ascent and spreading of a number of domes occur roughly simultaneously. That is, the unstable model has a limited period of tectonic activity during which the more stable arrangement develops. When a certain stability level has been reached the remaining mechanical potential is too small to cause further development, and the model becomes inactive (though not usually mechanical stable because the model materials have a certain finite strength, at least some of the materials). On the scale of the model the period of general activity is to be compared with the general orogenic period, e.g. the Caledonian period, in nature, but within this general period of activity there are little which indicates a shorter periodicity that affects the whole model.

In this connection it is interesting to see the opinion on the evolution of the British Caledonides as expressed by NEVILLE GEORGE (1963, p. 30) "Except in a general Caledonoid alignment, implying a system of large-scale crustal forces, Caledonoid structures in Britain are markedly heterotropic. They reveal a fluctuating impulse and style of movement that does not conform to a simple pattern correlated with an idealised Lower Paleozoic geosyncline or with a unitary mechanical frame of struts and strain".

The same could well be said of the Scandinavian Caledonides. The writer believes that this picture of the Caledonides built up by field work of a large number of geologists is in good agreement with the experimentally supported model presented in this paper.

Relation between the presented buoyant-basement model and the conventional convection-cell model

Of the various proposed models of orogenic evolution the convection-cell theory seems the most widely accepted in current geophysical literature. It is therefore of interest to compare the here presented buoyant-basement model with the convection-cell theory though we realize that there are serious—perhaps insurmountable—discrepancies between the latter theory and geologic-geophysical facts and inferences. Some of these discrepancies are for example recently discussed by KNOPOFF (1964). On the other hand the convection-cell theory and the author's buoyant-basement theory have some basic features in common, the most important being that both theories assume the earth's thermal energy (possibly produced by decay of radioactive nuclei) as the chief driving agency of orogenesis. But there are differences in the mechanisms and ways by which the thermal energy is coupled with the essentially mechanical process of orogeny.

The convection-cell theory assumes the presence of thermal convection currents in the earth's mantle. The upper branches of these currents flow laterally underneath the crust which by friction coupling is compressed, supposedly with resulting buckling and thrusting, in regions above the downward-moving branch of the convection cell. According to this model geosynclines and orogenic belts are consequently to be found above the subsiding colder branch of the convection cell.

In the buoyant-basement model the primary assumption is that the gravitationally stable state of the light sial arranged above the heavy sima is not established within geosynclines (and other oceanic tracts), and that the orogenic processes such as folding, thrusting and other rock deformations as well as the bodily and diffusional rise of quartzo-felspathic matter, are processes leading toward a more stable arrangement, or some secondary results of such equilibripetal processes. (The folds and faults do not necessarily represent more stable structures than the undeformed rocks, but such deformations are regarded as by-products of the equilibripetal rise of sialic- and subsidence of simatic matter.)

The unstable arrangement of crustal and possibly subcrustal matter in geosynclines is either a remnant feature of the primeval undifferentiated earth, or, what is more common, the unstable geosynclinal state develops from time to time during the evolution of the globe.

Now, at this point the buoyant-basement model appeals to the same energy source as does the convection-cell model. In the latter the geothermal gradient is the propelling agent as it is related to a continuous variation of density in a uniform mantle. In the author's model the geothermal gradient also produces a density variation, but a discontinuous one and a considerably greater one inasmuch as an important feature of the model is that differential melting takes place within the mantle and/or within the deep part of sima. Due to the low density of the melt, both because of the change from crystalline to liquid state and because of the contrasted chemical composition of melt and residual solids, the body of basic magma is strongly buoyant and rises to levels determined in part by the density contrast between melt and surrounding crystalline rocks through which it passes. If the rising column of melt is continuous through a considerable vertical distance the great density contrast between melt and adjacent rock at the deep part of the magma column produces a mechanical potential that is able to lift the top of the magma column through surroundings that are even less dense than the melt. However, the tendency of the melt to spread laterally increases as the density contrast between melt and adjacent rocks vanishes at high levels in the mantle and crust, the spreading tendency being especially strong where the surrounding rocks become less dense than the melt.

This happens when, for example, a basaltic melt intrudes the granitic layer, or when an ultrabasic magma intrudes sima from below (see experiments and quantitative discussion in RAMBERG, 1963 and 1964*a*). Under these circumstances only a small portion of the basic melt extrudes on the free surface as lava flow—the chief mass must spread laterally as huge sills or laccolites within the granitic layer and in the sedimentary strata in oceanic tracts. There are reasons

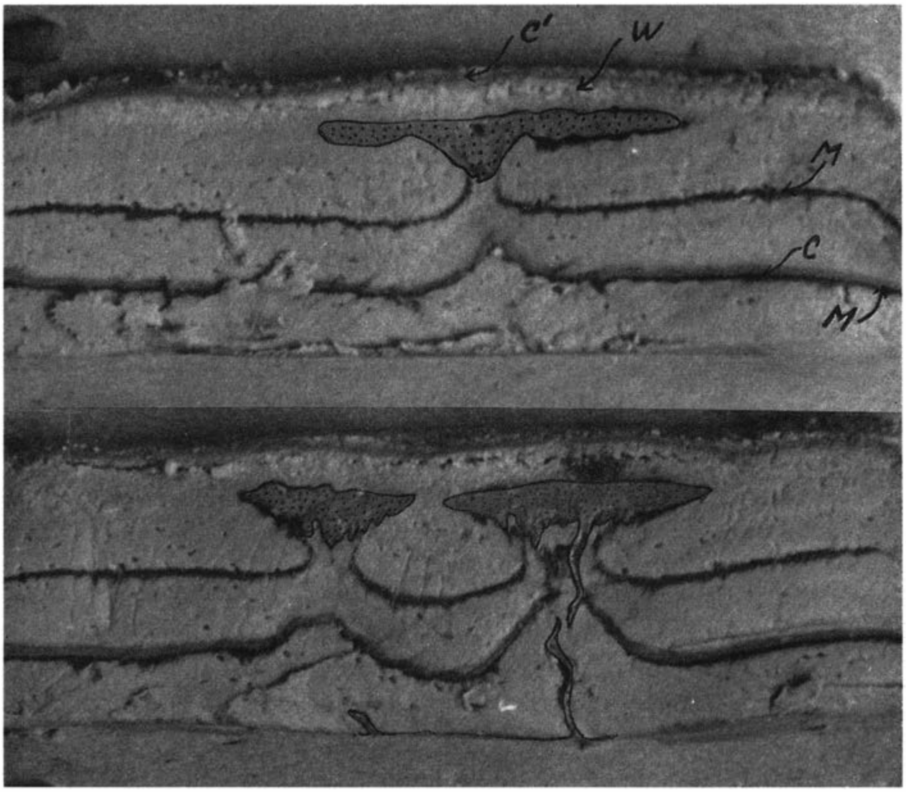


Fig. 64. Model of buoyant body (dotted) which has spread out under a layer with low density (w) (RAMBERG 1963).

to believe that the larger part of the basic melt spreads at the boundary between sial and sima, at least this is suggested by model experiments, Fig. 64.

In like fashion ultrabasic melts produced deep in the mantle may spread in sima or at the boundary between sima and the mantle. Such melts are probably too heavy to rise up through sial.

As the basic magma in the sima-invaded sial, and/or the ultrabasic magma in the dunite-invaded sima solidify the increased density creates a mechanically unstable situation. The first response to this unstability is an isostatic *en bloc* sinking of the whole region such that a topographic depression is produced (a geosyncline). (Incidentally, one may expect pulsation of the surface for when basic magma rises and spreads below or within the granitic layer the surface will rise whereas subsidence follows the solidification of the intruded magma.)

The subsidence is accomplished by sidewise flow of solid sial from the sima-invaded region, and/or flow of solid sima below the intruded large bodies of ultrabasic rocks. But this *en bloc* subsidence of the invaded region is not a final stable state though it may be isostatically balanced at sufficient depth of compensation. The mixing of bodies—layers—of heavy basic rocks with light quartzo-felspathic rocks, partly with complete inversion of density stratification, is a gravitationally unstable situation, which, if the various bodies are large enough, results in spontaneous processes tending to re-establish the more stable crustal stratification of sial on top and sima below. It is exactly this kind of mechanical processes which have been studied in our centrifuged experimental models whose end results bear such a striking resemblance to the structures encountered in orogenic belts.

The buoyant-basement model as applied to the Scandinavian Caledonides in this paper covers only the orogenic and cratogenic periods (folding- and upheaval periods) of fold-mountain evolution. The processes thus ascribed to buoyancy of the basement in geosynclines are by-products of a larger system of flows of energy and matter, namely the cycle of differential melting at great depth, rise of the melt with spreading and solidification at higher levels, and sinking of the heavy solidified bodies with consequent rise of subjacent and adjacent rocks as domes and buoyant anticlines.

The first-order cycle may be regarded as a discontinuous convection cycle—discontinuous because it involves melting and crystallization as essential processes, convection in the sense that heat transport is of prime significance in the model.

Appendix

Models of tectonic evolution at a continental border

Models S 118, S 119 and S 121 as shown in Figs. 65 and 66 are pertinent in a discussion of mountain building along continental borders.

The models are based on the assumption that along some continental borders, especially around the Pacific ocean, the continental shelf which consists essentially of light continental sial, is overlain by heavy basic lava. Such basic lava furthermore occurs alternating with sedimentary strata in the shelf deposits. The initial stage of our models are accordingly constructed, see Fig. 66. Painter's putty in the models simulates basic lava and also deep-seated sima while silicone putty simulates sial and the incompetent strata of the sedimentary column whose competent layers consist of modelling clay.

After a few minutes run in the centrifuge these layered structures change to the rather more complicated pattern as reproduced in the photos in Fig. 65.

There are in these models two structural features which are particularly reminiscent of circumpacific conditions, and so also the conditions prevailing during early stages of many mature mountain ranges, namely (1) a row of island arc (A), and (2) the well known huge batholithes (B) along the edge of some continents, especially along the West coast of both South- and North America.

Depressions have formed during the rise of the island arc and the batholithes in the models, both in the region between the island arc and the continental

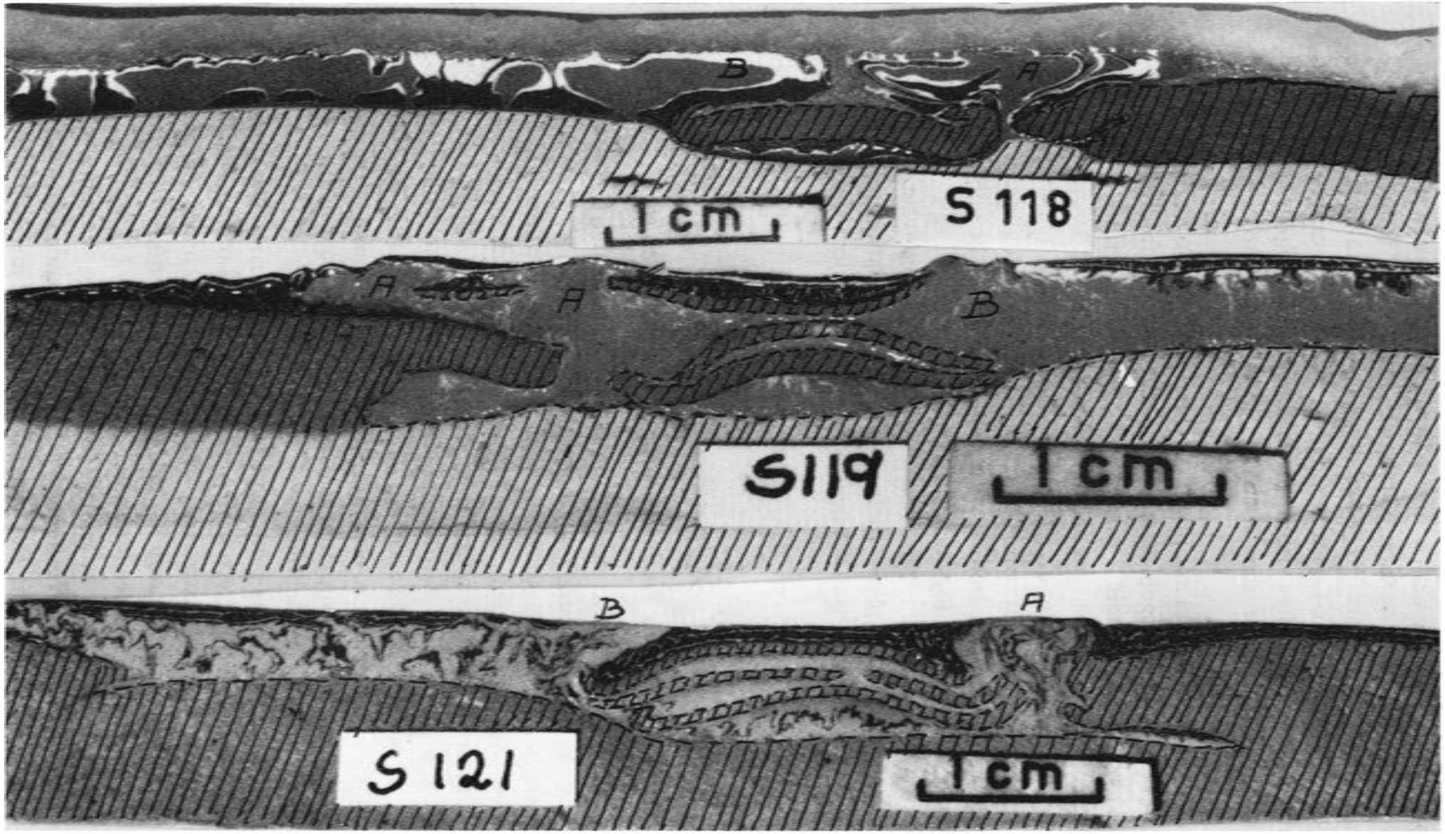


Fig. 65. Profiles through three centrifuged models of possible tectonic evolution within the border-region between continent and ocean of circumpacific type. Ocean on the right-hand side in S 118 and S 121, on the left-hand side in S 119. S 118 run 2 minutes at 700*g* and 14 minutes at 2000*g*. S 119 run 15 minutes at 2500*g* and 10 minutes at 2900*g*. S 121 run 3 minutes at 700*g* and 7 minutes at 2650*g*. *A:* island arcs, *B:* batholithes at continental edge. Inclined hatching: painter's putty (with unlike color), $\varrho = 1.87$ g/cm^3. In S 118*:* grey: silicone putty, $\varrho = 1.25$ g/cm^3; black: silicone putty, $\varrho = 1.34$ g/cm^3; white: silicone putty, $\varrho = 1.14$ g/cm^3; greyish white layer on top: oil-wax mixture. In S 119*:* grey with white irregular spots: silicone putty, $\varrho = 1.5$ g/cm^3 and 1.14 g/cm^3 (white); black with light-colored folded sheets on top of model: silicone putty with sheets of modelling clay. In S 121: light grey with dark irregular "sliren": silicone putty, $\varrho = 1.34$ g/cm^3; black with light colored buckled sheets on top of model: silicone putty with sheets of modelling clay. For information on initial structures see Fig. 66.

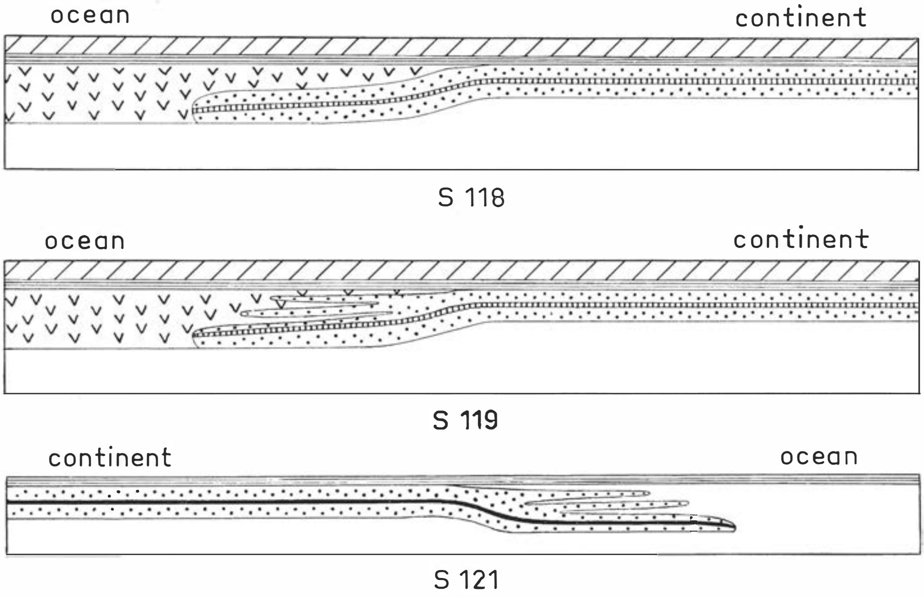


Fig. 66. Sections through models S 118, S 119 and S 121 prior to run in centrifuge. White: painter's putty, $\varrho = 1.87$ g/cm^3; angular marks: painter's putty (dark color), $\varrho = 1.87$ g/cm^3; dotted: silicone putty, $\varrho = 1.25$ g/cm^3, 1.34 g/cm^3 and 1.41 g/cm^3 (see Fig. 65); vertical lines: white silicone putty, $\varrho = 1.14$ g/cm^3; black sheet in S 121: dark silicone putty; thin horizontal lines: sheets of modelling clay in silicone putty; inclined hatching: oil-wax mixture. Note that continent-ocean sides reversed in S 118 relative to the photo, Fig. 65.

margin as well as just outside the islands, on their oceanic side in a position where the deep-sea troughs occur in nature.

Stratified deposits in these depressions have been more or less buckled chiefly due to push from the advancing lobes of the spreading continent and the rising arc system. Buckling due to sagging of surficial layers down into the depressions is also evident as indicated on the photos though perhaps not clearly visible in the reproductions. However, a greatly enlarged portion of a similar model (S 120) shown in Fig. 47, p. 43, exhibits clearly the Jura-type detachment buckling of the surficial strata in front of an advancing nappe.

Buckling of surficial strata in the backwash of a heavy body sinking through the crust

A sheet of heavy rock, thick and otherwise huge enough to sink with geologically significant speed in response to the pull of gravity, cannot escape producing severe tectonic disturbances in the adjacent and overlying crustal portions. A backwash of tremendous dimensions, though of course evolving with exceedingly slow movements in the sluggish materials of the crust, must be generated by the sinking body. One of the more conspicuous consequences of the backwash phenomena should be a buckling and lateral shortening of

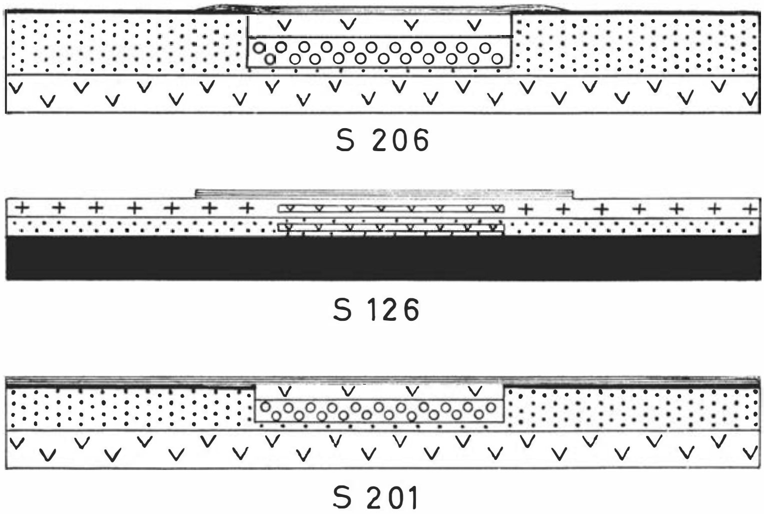


Fig. 67. Sections through three models of unstable density stratification prior to run in centrifuge (see Fig. 68). Angular marks: painter's putty, $\varrho = 1.87$ g/cm³ in S 201 and S 206, $\varrho = 1.95$ g/cm³ in S 126; dotted: silicone with $\varrho = 1.33$ g/cm³ in S 201 and 206, $\varrho = 1.14$ g/cm³ in S 126; open circles: dark silicone, $\varrho = 1.47$ g/cm³; thin horizontal lines: silicone putty with sheets of modelling clay; solid black in S 126: silicone putty, $\varrho \approx 1.35$ g/cm³; black sheet in S 206: modelling clay.

superincumbent strata just above the foundering body, and possibly a complementary lateral stretching in the surface region outside the sinking body.

To test his theoretical prediction concerning the evolution of such an event (RAMBERG, 1945), the author had several models constructed of the kind shown in Fig. 67. After being run to various stages of maturity in the centrifuge the following tendencies were observed in all instances: (1) The heavy plastic sheets (of painter's putty = basic igneous rocks) would bend into a gentle anticline while sinking. Because of the particular pressure distribution and flow pattern below such a sheet the edges must sink faster than the central regions (see RAMBERG, 1963 p. 45) thus generating an anticlinal shape. (2) Domes may or may not burst through the crest of the anticline, depending upon strength and viscosity of the heavy sheet, and of course depending upon the length of run in the centrifuge and the acceleration employed. (3) The subjacent and adjacent materials (sial) would flow centrifugally out from underneath the subsiding sheet, the flow lines bending around the edge of the sheet and continue along the upper surface of the sheet in a centripetal sense, just as one could readily predict (RAMBERG, 1945). (4) Surficial strata became buckled and laterally shortened above the subsiding sheet or in the regions between domes which may rise through the sheet. The strata were stretched in regions outside the sinking body and above spreading domes.

Fig. 68 shows cross sections through models S 126, S 201 and S 206. In the shown section of model S 126 the core of the wide anticline of the sinking sheets

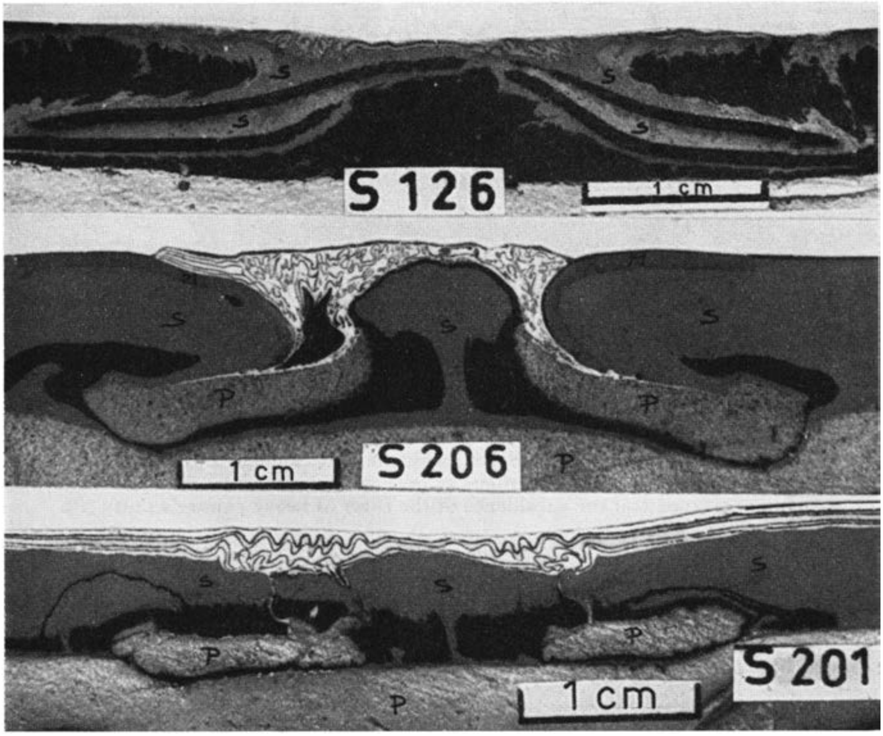


Fig. 68. Sections through models S 126, S 201 and S 206 after run in centrifuge (see Fig. 67). Black layers in S 126: heavy painter's putty. Black massive regions: dark silicone putty; white or light grey with dark lines: silicone putty with sheets of modelling clay; grey, *S:* silicone putty; *M:* sheet of modelling clay; light grey, *P:* painter's putty.

Fig. 69. Enlarged part of model S 206 showing details of buckling structure, the pulling down of the surficial "sediments" in the regions adjacent to the dome, and the boudins formed above the hat of the dome. The boudins are drawn in ink as they may not otherwise show up in the reproduction. See also Figs. 67 and 68.

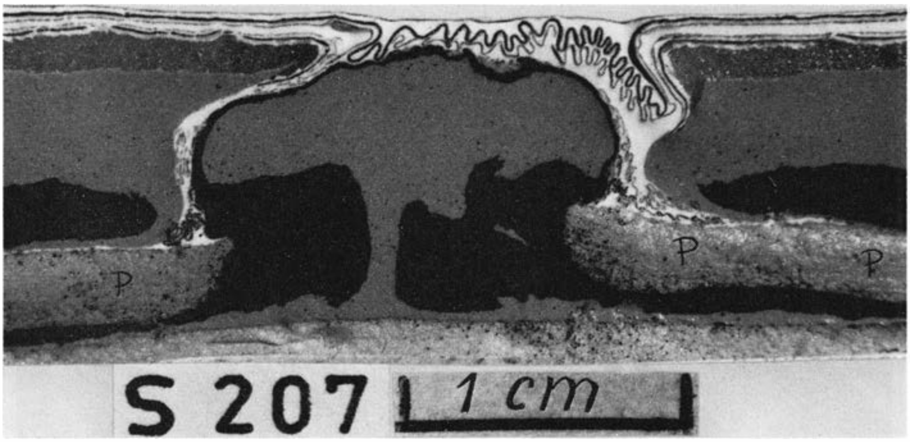


Fig. 70. Cross section through model S 207, which is quite similar to model S 206, showing intensive buckling and stretching of surficial multi-layer in connection with the domal rise of the light silicone and the subsidence of the sheet of heavy painter's putty (P).

has not pierced both sheets. In model S 206 piercement has occurred, and in model S 201 the heavy sheet has sunk to the bottom and practically all "sial" below it has risen as a central dome and as unsymmetric "domes" along the edges.

Note behavior of the piercement domes, the particular flow pattern of the adjacent "sial" and the behavior of the surficial strata. The latter is shown from above in Figs. 71 and 72.

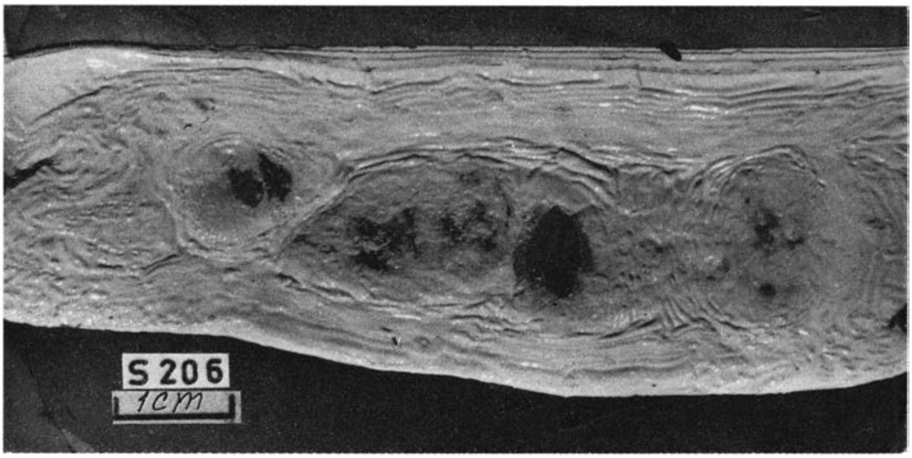


Fig. 71. Model S 206 seen from above after run in centrifuge. Note buckling of surficial layer adjacent to the domes and along the edge of the "geosyncline".

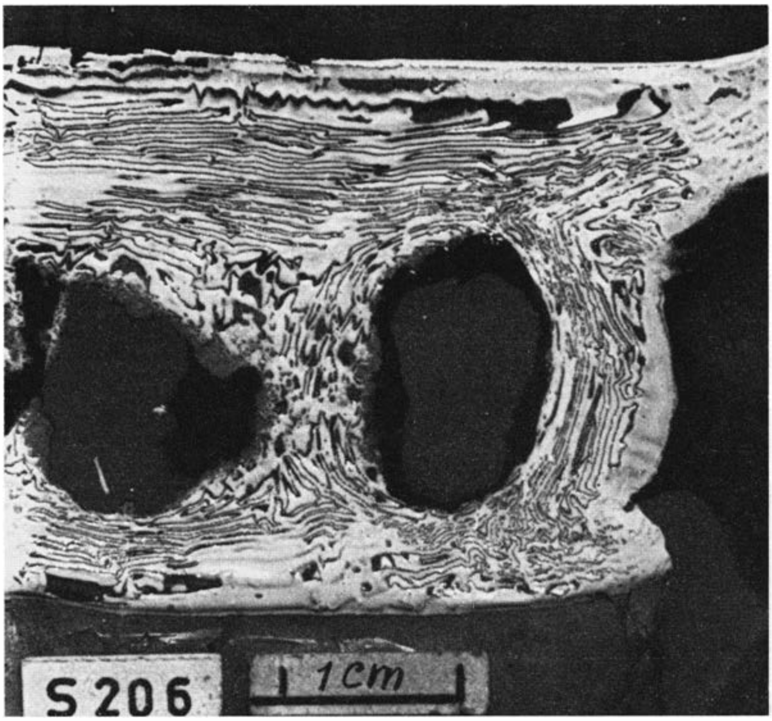


Fig. 72. Model S 206 as seen from above after half of it cut away for sectioning and a thin sheet cut off from the top to show the fold pattern of the surficial multilayer.

References

AUBOUIN, J., 1965: Geosynclines, Developments in Geotectonics I. 335 pp. Elsevier, Amsterdam.

BELOUSSOV, V. V., 1961: The origin of folding in the earth's crust. *Jour. Geoph. Research.* 66, pp. 2241–2254.

BEMMELEN, R. W. VAN, 1960: New view on east Alpine orogenesis. *Int. Geol. Congr. 21 session.* Norden. Part 18, pp. 99–116.

BILLINGS, M. P., 1945: Mechanics of igneous intrusion in New Hampshire. *Am. Jour. Sci.* 243 A, pp. 40–68.

BIRKELAND, T., 1958: Geological and petrological investigation in Northern Tröndelag. *Norsk Geol. Tidsskr.* 38, pp. 327–420.

CARSTENS, H., 1955: Jernmalmene i det vestlige Trondhjemsfelt og forholdene til kisforekomstene. *Norsk Geol. Tidsskr.* 35, p. 211.

— 1956: Geologi. Fosdalen bergverk 1906–1965, p. 149.

— 1960: Stratigraphy and volcanism of the Trondheimsfjord Area, Norway. *Norges Geol. Unders.* 212b (Excursion guide), pp. 1–22.

CHILDS, O. E., and WARREN, B. B., 1963: Backbone of the Americas. *Am. Ass. Petrol. Geol. Mem.* 2. 320 pp.

COLE, G. H. A., 1962: Fluid dynamics. 238 pp. Methuen & Co. Ltd., London.

COLLET, L. W., 1927: The structure of the Alps. 289 pp. E. Arnold & Co., London.

EARDLEY, A. J., 1963: Relation of uplifts to thrusts in Rocky Mountains. *Am. Ass. Petrol. Geol. Mem.* 2, pp. 209–220. (See also Childs *et al.*).
ESKOLA, P., 1921: On the eclogites of Norway. *Vid. Selsk. Skr. Kristiania,* 8.
— 1948: The problem of mantled gneiss domes. *Quart. Jour. Geol. Soc. London.* 104, pp 461–476.
FOSLIE, S., 1941: Tysfjords geologi. *Norges Geol. Unders.* 149, pp. 1–298.
FOSLIE, S., and STRAND, T., 1956: Namnsvatnet med en del av Fröyningsfjell. *Norges Geol. Unders.* 196, pp. 1–82.
GANSSER, A., 1964: Geology of the Himalayas. 289 pp. Interscience Publishers London–New York–Sydney.
GERTH, H., 1955: Der geologische Bau der Südamerikanischen Kordillere. 264 pp. Gebrüder Borntraeger, Berlin.
GJELSVIK, T., 1953: Det nordvestlige gneisområde i det sydlige Norge. *Norges Geol. Unders.* 184, pp. 71–94.
GOLDSCHMIDT, V. M., 1912: Die kaledonische Deformation der südnorwegischen Urbergstafel. *Vid. Selsk. Skr. Kristiania,* 10.
— 1915: Die Kalksilikatgneise und Kalksilikatglimmerschiefer des Trondhjem–Gebietes. *Ibid.* 19.
GROUT, F. F., 1945: Scale models of structures related to batholiths. *Am. Jour. Sci.* 243 A, pp. 260–284.
HAARMANN, E., 1930: Die Oszillationstheorie. 260 pp. Ferdinand Enke, Stuttgart.
HEIER, K. S., 1960: Petrology and geochemistry of high-grade metamorphic and igneous rocks on Langöy, northern Norway. *Norges Geol. Unders.* 207, pp. 1–246.
HERNES, I., 1956: Geologisk oversikt over Molde–Kristiansundsområdet. Kongl. Norske Vid. Selsk. Skr. 1955, 5.
HOLMES, A., 1965: Principles of Physical Geology. 1288 pp. Nelson Ltd., London Edinburgh.
HOLMSEN, P., 1955: Trekk av Oppdalsfeltets geologi. *Norsk Geol. Tidsskr.* 35, p. 135
—, 1960: see Strand and Holmsen.
HOLTEDAHL, H., 1950: Geological and petrological investigations in the NW'ern part of the Oppdal quadrangle. B.U. 1949, 7.
HOLTEDAHL, O., 1938: Geological observations in the Opdahl–Sunndal–Trollheimen district. *Norsk Geol. Tidsskr.* 18, p. 29.
— *et al.*, 1960: Geology of Norway. *Norges Geol. Unders.* 208, pp. 1–540.
JONES, R. W., 1963: Structural evolution of part of South east Arizona. *Am. Ass. Petrol. Geol. Mem.* 2, pp. 140–152. (See also Childs *et al.*).
KAUTSKY, F., and TEGENGREN, F. R., 1952: Die Geologie der Umgebung des Tuoddarjaure am Südrande des Sjangelifensters. *Geol. För. Förh. Stockholm.* 74, pp. 455–474.
KAUTSKY, G., 1952: Der geologische Bau des Sulitelma–Salojauregebietes in den Nordskandinavischen Kaledoniden. *Sveriges Geol. Unders.* 528. Ser. C, pp. 1–228.
KNOPOFF, L., 1964: The convection current hypothesis. *Rev. Geophys,* 2. pp. 89–122.
KRAUS, E., 1951: Die Baugeschichte der Alpen, I. 552 pp. Akademie-Verlag, Berlin.
— 1936: Der Abbau der Gebirge, I: Der alpine Bauplan. 352 pp. Gebrüder Borntraeger, Berlin.
KVALE, A., 1948: Petrologic and structural studies in the Bergsdalen quadrangle. Part II. *Bergens Mus. Årb.* 1. pp. 1–255.
— 1953: Linear structures and their relation to movement in the Caledonides of Scandinavia and Scotland. *Quart. Jour. Geol. Soc. London,* pp. 51–64.
— 1960: The nappe area of the Caledonides in Western Norway. *Norges Geol. Unders.* 212e, pp. 1–43 (Excursion guide).
LANDMARK, K., 1951: Tverfolding i den kaledonske fjellkjede. *Norsk Geol. Tidsskr.* 29, p. 241.

LJUNGNER, E., 1950: Urbergsytans form vid fjällranden. *Geol. För. Förh. Stockholm.* 72, pp. 269–300.
LUNDEGÅRDH, P. H., LUNDQVIST, J. and LINDSTRÖM, M., 1964: Berg och jord i Sverige. 345 pp. Almqvist & Wiksell, Stockholm.
MAGNUSSON, N. H., THORSLUND, P., BROTZEN, F., ASKLUND, B., and KULLING, O., 1962: Beskrivning till karta över Sveriges berggrund. *Sveriges Geol. Unders.* 16. Ser. Ba, pp. 1–290.
MURET, G., 1960: Partie S. E. de la culmination de Romsdal, Chaine Caledonienne, Norvège. *Int. Geol. Congr. 1960.* Part XIX, Sec. 19, pp. 28–32.
NAUMANN, C. F., 1849: Lehrbuch der Geognosie, Leipzig. Ed. 1.
— 1858: Lehrbuch der Geognosie, Leipzig. Ed. 2.
NEVILLE GEORGE, T., in JOHNSON, M. R. W., and STEWART, F. H., 1963: The British Caledonides. 280 pp. Oliver & Boyd, Edinburgh–London.
NICHOLSON, R., and WALTON, B. J., 1963: The structural geology of the Navervatn–Storglomvatn area, Glomfjord, Norway. *Norsk Geol. Tidsskr.* 43, pp. 1–58.
OFTEDAHL, CH., 1952: Övre Rendal. *Norges Geol. Unders.* 177. pp. 1–47.
— 1955: Om Grongkulminasjonen og Grongfeltets skyvedekker. *Norges Geol. Unders.* 195, pp. 57–64.
PARKER, T. J., and MCDOWELL, A. N., 1955: Model studies of salt dome tectonics. *Bull. Amer. Ass. Petrol. Geol.* 39, pp. 2384–2470.
PEACEY, J. S., 1964: Reconnaissance of the Tömmerås anticline. *Norges Geol. Unders.* 227, pp. 13–84.
RAMBERG, H., 1943: En undersökelse av Vestrandens regionalmetamorfe bergarter. *Norsk Geol. Tidsskr.* 23. pp. 1–174.
— 1944*a*: Relation between external pressure and vapor tension of compounds, and some geological implications. *Norske Vid. Akad. Oslo*, avh. Mat. Nat. kl. 3, pp. 1–24.
— 1944*b*: The thermodynamics of the earth's crust I. *Norsk Geol. Tidsskr.* 24, pp. 98–111.
— 1945: The thermodynamics of the earth's crust II. *Norsk Geol. Tidsskr.* 25, pp. 307–326.
— 1948: Radial diffusion and chemical stability in the gravitational field. *Jour. Geol.* 56, pp. 448–458.
— 1963: Experimental study of gravity tectonics by means of centrifuged models. *Bull. Geol. Inst. Univ. Uppsala.* Vol. XLII, pp. 1–97.
— 1964*a:* A model for the evolution of continents, oceans and orogens. *Tectonophysics.* 2, pp. 159–174.
— 1964*b:* Note on model studies of folding of morains in piedmont glaciers. *Jour. Glaciol.* 5, pp. 207–218.
— 1965: The effect of gravity on folding and other deformations in the earth's crust. *Bergmekanik, IVA Medd.* 142, pp. 40–59. Stockholm.
— 1966: Experimental models of fold mountains. *Geol. För. Förh. Stockholm.* 87, pp. 485–491.
RAMBERG, H., and STEPHANSSON, O., 1964: Compression of floating elastic and viscous plates affected by gravity, a basis for discussing crustal buckling. Tectonophysics. 1, pp. 101–120.
— 1965: Note on centrifuged models of excavations in rocks. Tectonophysics. 3, pp. 281–298.
REKSTAD, J., 1929: Salta. *Norges Geol. Unders.* 134, pp. 1–73.
ROSENQVIST, I., 1941: The Lönset anticline in the Oppdal area. *Norsk Geol. Tidsskr.* 21, p. 25.
SCHAAR, G., 1962: Petrological and geological investigations in the Sylene–Skardörsfjell region. *Medd.* 273. *Geologisch Instituut, Universiteit van Amsterdam*, pp. 1–74.

SKJESETH, S., and SÖRENSEN, H., 1953: An example of granitization in the central zone of the Caledonides of Northern Norway. *Norges Geol. Unders.* 184, pp. 154–183.

SKJESETH, S., 1963: Contribution to the geology of the Mjösa district etc. *Norges Geol. Unders.* 220, pp. 1–126.

STRAND, T., 1951: The Seland Vågå map areas. *Norges Geol. Unders.* 178, pp. 1–117.

— 1953*a:* Geologiske undersökelser i den sydöstligste del av Helgeland. *Norges Geol. Unders.* 184, pp. 124–141.

— 1953*b:* The relation between the basal gneiss and the overlying metasediments in the Surnadal district. *Norges Geol. Unders.* 184, pp. 100–123.

— 1961: The Scandinavian Caledonides, a review. *Am. Jour. Sci.* 259, pp. 161–172.

— 1964: Geology and structure of the Prestberget area. *Norges Geol. Unders.* 228, pp. 289–310.

STRAND, T., and HOLMSEN, P., 1960: Stratigraphy, petrology and Caledonian nappe tectonics of central southern Norway: Caledonized basal gneisses in a north western area. *Norges Geol. Unders.* 212 i (Excursion guide), pp. 1–31.

STRÖMBERG, A. G. B., 1961: On the tectonics of the Caledonides in the southwest part of the county of Jämtland, Sweden. *Bull. Geol. Inst. Uppsala.* 39, pp. 1–91.

VOGT, T., 1928: Den Norske fjellkjedes revolusjonshistorie. *Norsk Geol. Tidsskr.* 10, p. 97.

WEGMANN, E., 1959: La hexure axiale de la Driva et quelques problèmes structuraux des Calédonides Scandinaves. *Norsk Geol. Tidsskr.* 39, pp. 25–74.

WOLFF, F. C., 1964: Stratigraphical position of the Gudå conglomerate zone. *Norges Geol. Unders.* 227, pp. 85–91.

ZACHRISSON, E., 1964: The Remdalen synclines. *Sveriges Geol. Unders.* 596. Ser. C, pp. 1 –53.

YODER, H. S., and TILLEY, C. E., 1962: Origin of basalt magmas, An experimental study of natural and synthetic rock systems. *Jour. Petrology.* 3, pp. 342–532.

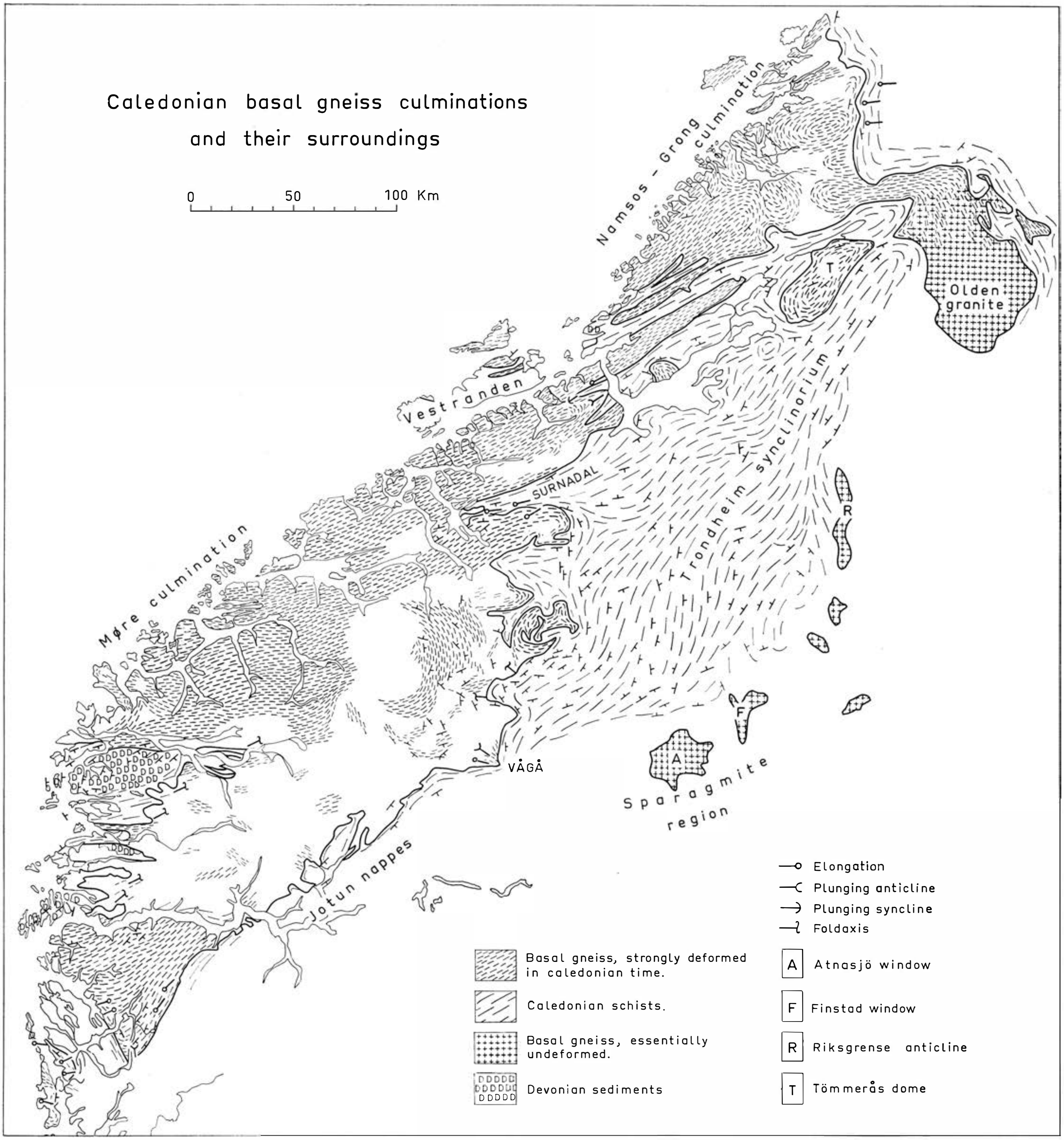


Fig. 7. The Möre culmination, Vestranden and the Namsos–Grong culmination with surroundings, compiled from published works by a number of geologists as cited in the text.