Bull. geol. Instn Univ. Upsala N. S. 3, 4: 37—78, 1971

GRAVITY TECTONICS ON ÖLAND

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(Submitted April 15, 1971)

CONTENTS

ABSTRACT

The present paper deals with different types of diapiric structures from the island of Öland, southern Sweden. The structures vary in size and in age and include different lithological units from the Middle Cambrian to the Middle Ordovician.

The most striking structures of southern Öland are a number of large ridges — called main ridges — which radially diverge from the western part of the island. The ridges have widths of 50—300 m and the largest one is almost 30 km in length. Steeply dipping older beds of early Ordovician limestone are found in the ridge centre, which is flanked by younger and more gently dipping strata. Culminations on the ridges are a common feature associated with changes in the directions of sets of joints in their vicinity. The term "single-layer anticlinal diapir" is suggested for this type of structure, i.e. one caused by gravitational instability due to density differences in the strata. The term 'single-layer' implies one layer of buoyant material, here consisting of the low-density Upper Cambrian and Lower Ordovician alum shales. The main ridges are absent from the northern parts of Öland due to the decreasing thickness of the alum shales.

The anticlinal structures next in size are the so-called 'hump-back ridges'. These are upright, subhorizontal anticlines rising above the flat-lying surroundings. Their length varies between a few hundred metres and about two kilometres; their wavelength is 10 m on an average and their height sometimes reaches 2 m. An orientation parallel with major joints or sets of joints is a common situation. The term "multi-layer anticlinal diapir" is suggested for this type of structure since the emplacement is due to gravitational instability of several layers of low-density marly and shaly intercalations in the limestone units.

Domes with diameters of about 200 m have been found at two places on southern Öland. The domes are "single-layer domical diapirs" and result from buoyancy of the alum shales. Intermediate domes, with diameters of 5 to 15 m, are rather common structures. Also, domains with five to ten regularly spaced domes occur. The intermediate domes are of multi-layer type where intercalations of marl and shale are the driving agencies of doming. The initiation of the growth is explained by stratification irregularities. An outline of possible age of formation is presented for the different diapiric structures on southern Öland.

Two large structures are described from central Öland. The Mossberga dome, with a diameter of about 4 km, is concentrically built of outward dipping Lower Ordovician limestones in the random part with successively older rocks towards its centre. The dome-generating irregularity at Mossberga is a monadnock of Archaean quartzite which in all probability protrudes into the overlying bituminous shale of Middle Cambrian age. The gravitational instability of light shales overlain by denser limestone made the doming possible. The tectonic style in the area of Borgholm differs from other places on Öland. Faulting and rather strong tilting of limestone units are found, e.g. along the Gestadås structure.

Small domes and ridges with diameters and wavelengths of around one meter are the most characteristic structures on northern Öland. Single-layer domical or anticlinal diapirs dominate and marl is the most common buoyant material. The very early occurrence of diapirism in the history of the sediments is readily deduced from the almost horizontal bedding in the very next marl or limestone bed above the bent one.

INTRODUCTION

Ever since Professor Hans Ramberg organized the tectonic laboratory at the Institute of Mineralogy and Geology at the University of Uppsala in the early sixties, studies there have revolved round different types of gravity tectonics. For the experimental model studies of gravity tectonics the use of a large-capacity centrifuge has yielded new and important results (Ramberg, 1963a, 1967). Theoretical analysis within the field of geodynamics has proceeded hand in hand with the experimental work. In the search for a field area displaying structures where theory, experiment and observation in the field could be intimately linked together the island of Öland is found to be well suited.

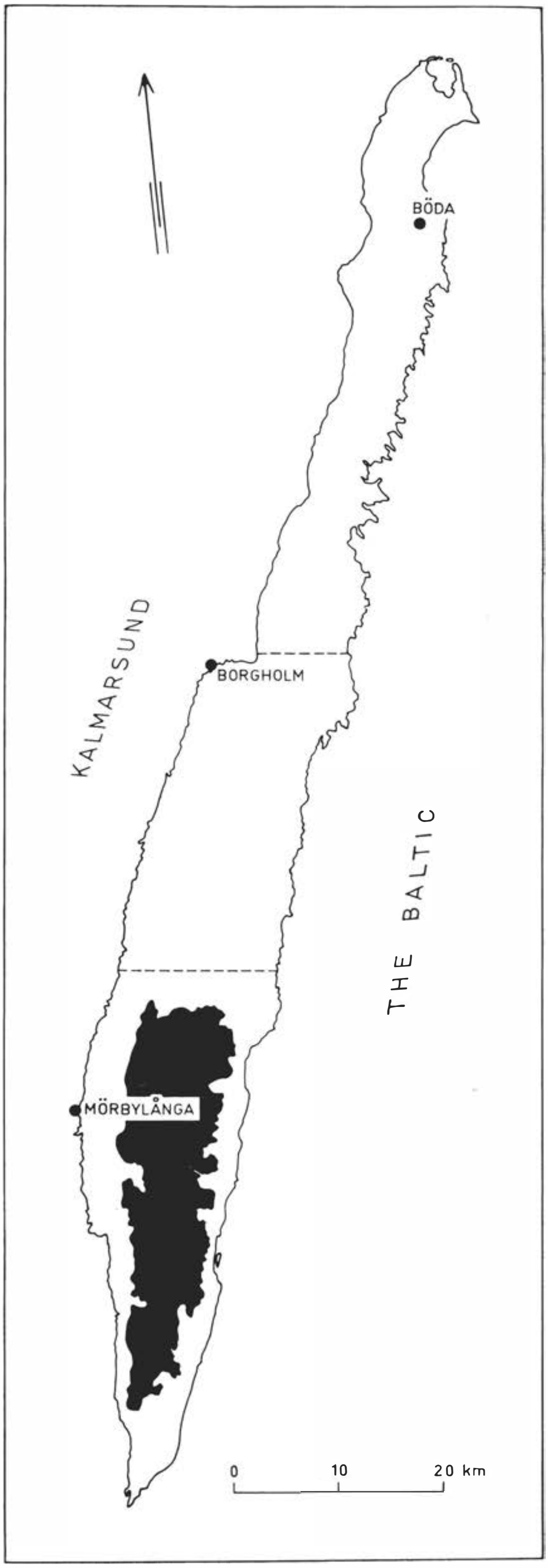
The possibility of Öland as a field area, where some of the theoretical and experimental results on gravity tectonics could be tested, arose through personal communications with Dr. K. Palmqvist in February 1968. In 1965 Palmqvist described small dome structures on southern Öland. In the spring of 1968 I also paid attention to some of the structures — mainly the ridges — and the tectonic map presented by Königsson (1968).

In the summer of 1969 I visited Öland and examined the small domes described by Palmqvist (1965), the ridge Hog's Back (Svinryggen) mapped by Bergsten and Svensson (1963) and the structures mentioned by Königsson (op. cit.) including the Resmo-Ekelunda zone and the small ridges at Gösslunda and west of the Hog's Back. Further, the folds described by Lindström (1963) from several localities on Öland were analysed. The structures showed many striking similarities with those obtained in centrifuged models. This prompted detailed mapping of the Öland structures.

The Great Alvar on southern Öland was mapped in 1969 with support of aerial photographs on the scale of 1:5000, kindly placed at my disposal by Professor L.-K. Königsson (Fig. 1). At that time the idea of gravitational emplacement of the large structures on the Great Alvar, due to a density inversion in the strata, was supported by the report of true shale domes in the Gulf Goast area of the United States (Gilreath, 1968). Results from a series of rock density determinations of different stratigraphic units on Öland, made by Dr. B. Dahlman at the Geological Survey of Sweden, clearly show that density inversion exists in the strata (Dahlman, personal communications).

Mapping of tectonic structures on the central and northern part of the island was performed during the summer of 1970. As might have been expected the large ridges or anticlinal diapirs mapped on southern Öland were not found on other parts of the island due to the small thickness of the buoyant layer of alum shale. On central Öland special attention was called to the two large structures at Mossberga and Gestadås; both were investigated in the early thirties under the guidance of Dr. O. Meier. On northern Öland the investigations have been concentrated on the well-exposed cliff of Västra Landborgen along the shore of Kalmarsund and the area immediately east of the cliff. The fold structures at Horns Udde, described by Lindström (1963), were analysed in detail.

Since the structures to be found on Öland appear in different litho-stratigraphic units, and more-



over, as the information concerning these units is spread in different publications, there was need of a stratigraphic review. This has been compiled mainly from the works of Hedström and Wiman (1906), Westergård (1936, 1947), Bohlin (1955), Jaanusson (1960a), and Martinsson (1965).

Some of the large structures mapped on Öland are interesting from a hydrogeological point of view. Hence, the major results of the mapping have been communicated to Drs. B. Dahlman, H. Möller, and J. Pousette at the Geological Survey of Sweden.

REVIEW OF THE STRATIGRAPHIC SETTING AND THE APPEARANCE OF DIAPIRIC STRUCTURES ON SOUTHERN ÖLAND

The island of Öland on the eastern coast of Sweden is built up of Cambrian and Ordovician rocks, the latter represented mainly by limestones. In general the strata dip very gently ESE and the boundaries between the different units run roughly along the island. The oldest rock found on Öland is the top of the Lower Cambrian Sandstone. This rises above sea-level within a small area on the shore at Mörbylånga. The bottom part of the Middle Cambrian, the *Eccaparadoxides oelandicus* Beds, consists of more or less arenaceous shales, Westergård (1936). South of Mörbylånga the unit rises above sea level and occupies a narrow strip along the shore, Fig. 2. According to Westergård (op. cit.) the thickness of the *Oelandicus* Beds is supposed to be around 25 m in the area of Mörbylånga. However Martinsson (1965) found that if the sequence is not disturbed, the *Oelandicus* Beds at Mörbylånga can hardly exceed 7—8 m in thickness. Towards the north the thickness of the *Oelandicus* Beds increases and a bituminous shale forms the bottom part of the beds. The outcrop of Lower Cambrian sandstone together with the moderate thickness of the *Oelandicus* Beds in the

Fig. 1. Sketch map of Öland showing the main division into southern, central, and northern parts. The shadowed area is the Great Alvar, compiled from the physiognomic map by Königsson (1968).

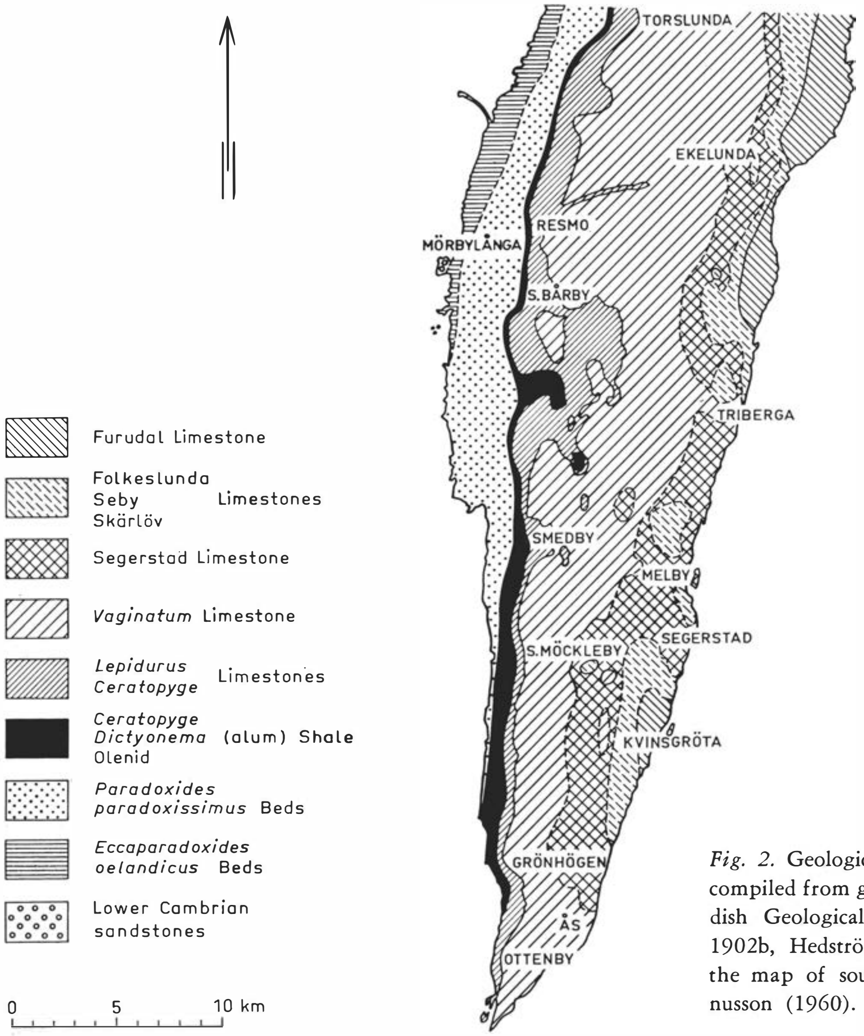


Fig. 2. Geological map of southern Öland compiled from geological maps of the Swedish Geological Survey (Munthe 1902a, 1902b, Hedström & Wiman 1906) and the map of southernmost Öland by Jaanusson (1960).

area of Mörbylånga seem to be intimately related to the main ridges or anticlines on the Great Alvar. The strike of these ridges radially converge towards the Mörbylånga area.

The upper boundary of the *Oelandicus* Beds is lithologically and faunistically sharply marked by the *Acrothele granulata* Conglomerate. Above this are the *Paradoxides paradoxissimus* Beds which according to Martinsson (op. cit.) comprise two typical facies, one with a considerable content of siltstone beds with shaly intercalations called the *Paradoxissimus* Siltstone and the other a grey shale without distinct siltstone beds called the *Paradoxissimus* Shale. Owing to the Quaternary cover it is difficult to strictly distinguish the two facies, and their stratigraphic relationships are fairly obscure. The extremely easily weathered shaly facies are very similar to those of the *Oelandicus* Beds (Martinsson, op. cit.).

The main strip between the *Oelandicus* Beds and the escarpment of the "Västra Landborgen" is occupied by the *Paradoxissimus* Beds, Fig. 2. The total thickness of these is about 60 m in the southern part of the island. Above this the *Paradoxides forchhammeri* Zone is represented on Öland by a thin and sporadic conglomerate, the *Exporrecta* Conglomerate which is covered by the Upper Cambrian Olenid Series. The alum shales embracing the Olenid, *Dictyonema,* and *Ceratopyge Shales* (the two latter belonging to the Ordovician) will be treated in more detail since they are of great importance in the formation of the main ridges and some of the dome structures. Our knowledge of the alum shale is mainly based upon the work of Westergård (1944, 1947).

The Olenid Shale has a thickness of around 13 m at Ottenby in the southernmost part of the island which gradually decreases to about 5 m at Mörbylånga. The isopachytes run in an east-west direction. Stinkstone is very common in the Olenid Shale, 24—63 % per volume, but is almost lacking in other beds of the alum shale. The isopachytes of the Lower Ordovician *Dictyonema* Shale have a direction more or less coincident with the length of the island. The thickness along the eastern coast line is about 8 m in the southernmost parts and the unit vanishes 4 km E of Mörbylånga. Finally, the *Ceratopyge* Shale, developed as alum shale extends half way between Mörbylånga and Ottenby. The thickness of the *Ceratopyge* Shale in the boring at Ottenby is 2.3 m. The isopachyte map, Fig. 3, shows the total thickness of the alum shale. This figure was constructed from the isopachyte map of Westergård (1947) together with information provided by The Water Well Archives of the Geological Survey of Sweden. The isopachyte map shows a remarkable thickening of the alum shale along the main ridges. From a well at Kvinsgröta in the centre of the Bårby—Ås Ridge the thickness is known to be 32.7 m. Further, small local outcrops of *Dictyonema* Shale are found in the central part of the same ridge just south of Bårby. Outcrops of alum shale are not known from the Triberga and Resmo—Sandby ridges but narrow zones of *Limbata* and *Planilimbata* Limestone in stratigraphically younger limestones are found, and are an indirect proof of piercement and hence a thickening of the alum shale. There is supposed to be a slight divergence of the normally NNE striking isopachytes in the northern part of the Great Alvar at the eastern part of the Dröstorp ridge and the Vickleby ridge.

The upper part of the Tremadocian *Ceratopyge*

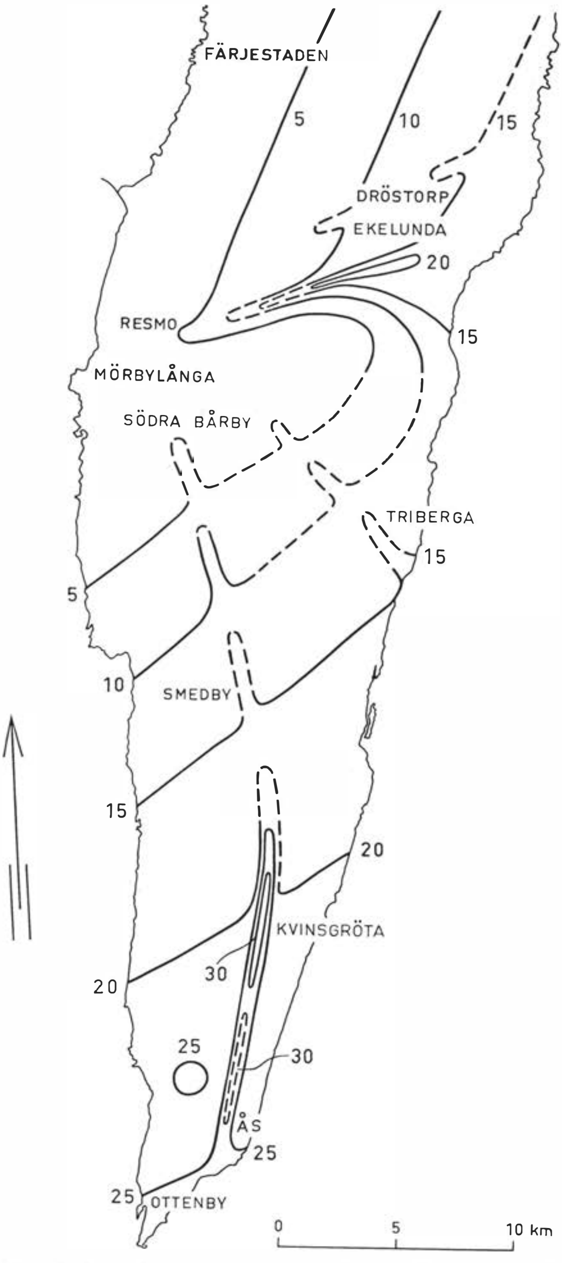


Fig. 3. Isopachyte map of the Upper Cambrian and Lower Ordovician alum shales. The thicknesses are compiled from the well record office of the Geological Survey of Sweden; 5 m interval of isopachytes.

Shale is a glauconitic shale; the glauconite content sometimes reaches 85 % of the mass of the rock according to Hadding (1932). The thickness of the glauconitic shale is 0.3 m on the shore W of Ottenby and 0.9 m at Bårby (Hadding, op. cit., fig. 3). The late Tremadocian *Ceratopyge* Limestone makes up the basal part of the so-called orthoceratite limestone. It consists of beds of grey glauconitic limestone intercalated with layers of greenish gray shale. Outcrops in the quarry of Degerhamn and investigations by Hadding (1932) and Tjernvik (1956) indicate a thickness varying between 0.3 m and 1.5 m. The intercalated shales of the limestone beds of Tremadocian and Arenigian ages and the underlying Tremadocian glauconitic shale are supposed to be the driving agencies in the formation of the intermediate dome structures along the western part of the Great Alvar and some of the saddle-shaped ridges on its western and central part.

According to Tjernvik (1956) and Lindström (1963) the Arenigian of Öland is from the top divisible into:

Lepidurus	Limestone
Limbata	Limestone
Billingen	Stage
Hunneberg	Stage

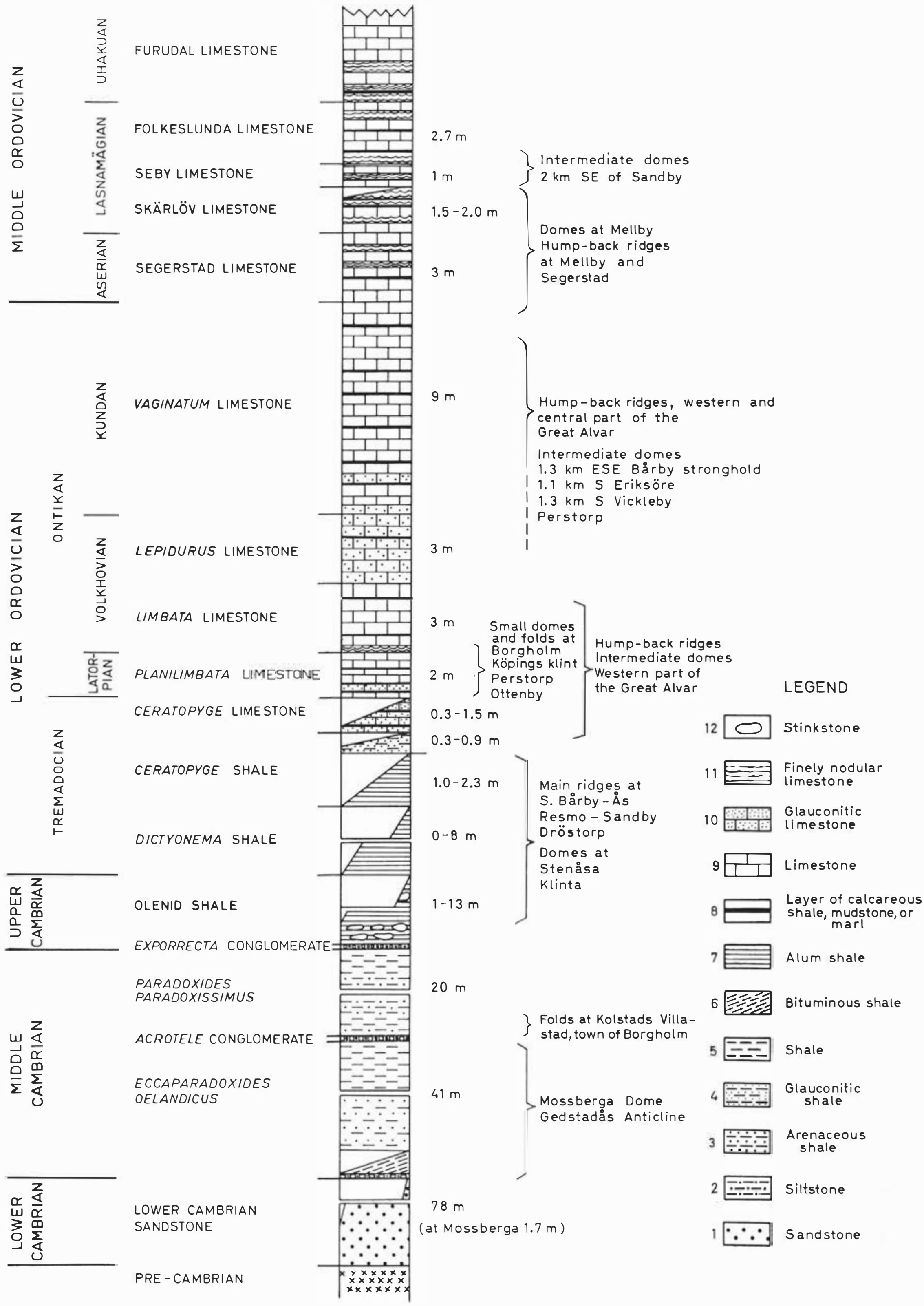
The Billingen and Hunneberg Stages are equivalent to the previously used term *Planilimbata* Limestone. The total thickness of the *Planilimbata* Limestone is about 8 m, and the *Limbata* and *Lepidurus* Limestones are each around 3 m in thickness. The lower part of the Arenigian limestones are layered with marly and shaly intercalations. According to Jaanusson (1960a) the Arenigian limestones, except for the *Lepidurus* Limestone, can be classified as calcilutites and calcarenitic calcilutites. At the base of the *Lepidurus* Limestone more sandy sediments have undergone a strong recrystallization and most of the limestone has therefore been classified as a recrystallized calcarenite. In the sense of Jaanusson (1952) the limit between calcarenite and calcilutite is drawn at about 20 % per volume of the "sand fraction" in the limestone. Lindström (1963) has investigated fold structures with wavelengths of about 0.5 m at several places on the island. The folds were found in the entire lower part of the Arenigian but were most common in the lower part of the Billingen Stage. The Lower Ordovician below the Kunda Stage occupies a narrow strip along the "Västra Landborgen" except in the area of Smedby and east of Södra Bårby, Fig. 2.

The Vaginatum group of limestones belonging to the Kunda stage of the Lower Ordovician consists of light gray pale brownish to brownish red limestones (recrystallized calcarenites). In the southern part of the island the total thickness is around 9 m. The *Vaginatum* Limestone — very capable of resisting denudation — occupies the main part of the Great Alvar. A thorough description of the *Vaginatum* limestones is available for northern Öland (Bohlin 1955) but is still lacking for the southern part.

The stratigraphy and lithology of the Viruan (Middle Ordovician) limestones of the bedrock of the northern and southern part of Öland has been worked out by Jaanusson (1960). He has presented a combined litho- and bio-stratigraphic classification. The Segerstad Limestone which is of Aserian age forms the bottom part of the Viruan of Öland. The limestone is a reddish brown, mostly thick-bedded calcarenite sometimes with intercalations of finely nodular limestone. The uppermost parts consist of coarse extensively recrystallized calcarenites with occasional intercalations of beds of finely nodular limestone or marl, Jaanusson (op. cit.). The entire succession is known from the borings at Skärlöv and Gammalsby where the thickness is around 3 m. Throughout the whole of Öland the Segerstad Limestone is overlain with bedded calcarenites which belong to the Skärlöv Limestone. The thickness of the Skärlöv Limestone varies between 1.5 and 2 m. The marl layers belonging to the Segerstad and Skärlöv Limestones are supposed to have actively participated in the formation of the saddle-shaped ridges and intermediate domes on the eastern part of the Great Alvar at Segerstad and Mellby.

The Skärlöv Limestone is overlain by the Seby Limestone, a reddish brown bedded limestone (mostly calcarenite) intercalated with fine nodular

Fig. 4. Stratigraphic position of tectonically active layers in the formation of different diapiric structures on southern central Öland. The thicknesses of the stratigraphic units are approximate and refer to southern Öland.



MIDDLE ORDOVICIAN
UHAKUAN
LASNAMÄGIAN
ASERIAN
LOWER ORDOVICIAN
ONTIKAN
KUNDAN
VOLKHOVIAN
LATOR-PIAN
TREMADOCIAN
UPPER CAMBRIAN
MIDDLE CAMBRIAN
LOWER CAMBRIAN
FURUDAL LIMESTONE
FOLKESLUNDA LIMESTONE
SEBY LIMESTONE
SKÄRLÖV LIMESTONE
SEGERSTAD LIMESTONE
VAGINATUM LIMESTONE
LEPIDURUS LIMESTONE
LIMBATA LIMESTONE
PLANILIMBATA LIMESTONE
CERATOPYGE LIMESTONE
CERATOPYGE SHALE
DICTYONEMA SHALE
OLENID SHALE
EXPORRECTA CONGLOMERATE
PARADOXIDES PARADOXISSIMUS
ACROTELE CONGLOMERATE
ECCAPARADOXIDES OELANDICUS
LOWER CAMBRIAN SANDSTONE
PRE-CAMBRIAN
2.7 m
1 m
1.5 - 2.0 m
3 m
9 m
3 m
3 m
2 m
0.3 - 1.5 m
0.3 - 0.9 m
1.0 - 2.3 m
0 - 8 m
1 - 13 m
20 m
41 m
78 m
(at Mossberga 1.7 m)
Intermediate domes 2 km SE of Sandby
Domes at Mellby
Hump-back ridges at Mellby and Segerstad
Hump-back ridges, western and central part of the Great Alvar
Intermediate domes
1.3 km ESE Bårby stronghold
1.1 km S Eriksöre
1.3 km S Vickleby
Perstorp
Small domes and folds at
Borgholm
Köpings klint
Perstorp
Ottenby
Hump-back ridges
Intermediate domes
Western part of the Great Alvar
Main ridges at
S. Bårby - Ås
Resmo - Sandby
Dröstorp
Domes at
Stenåsa
Klinta
Folds at Kolstads Villa-stad, town of Borgholm
Mossberga Dome
Gedstadås Anticline
LEGEND
12 Stinkstone
11 Finely nodular limestone
10 Glauconitic limestone
9 Limestone
8 Layer of calcareous shale, mudstone, or marl
7 Alum shale
6 Bituminous shale
5 Shale
4 Glauconitic shale
3 Arenaceous shale
2 Siltstone
1 Sandstone

limestone; the thickness is around 1 m. The superimposed Folkeslunda Limestone is a grey calcarenite with intercalations of finely nodular limestone and marl. Following the litho-stratigraphic division of Jaanusson (op cit.) the Skärlöv, Seby und Folkeslunda Limestones belong to the Lasnamägi Stage. The entire succession of this stage is known from the boring of Gammalsby and is 5.3 m in thickness. The limestones have a range of exposures close to the eastern boundary of the Great Alvar. Domes of small and intermediate size — diameter 0.5 to 5 m — appear in the Seby Limestone on the Baltic sea shore north of Hålnäs situated 2 km SE of Sandby.

The stratigraphically youngest unit exposed on southern Öland is the Furudal Limestone which belongs to the Uhaku Stage. The rock is mainly calcilutite with intercalations of finely nodular limestone and marl. The calcilutitic development and the marl intercalations are concentrated in the lowermost parts with an areal extension over the whole of Öland.

The stratigraphic column, the main lithology of the beds and the stratigraphical position of the tectonically active layers in the formation of the different diapiric structures on southern and central Öland are presented in Fig. 4. The stated thicknesses of the stratigraphic units are approximate and refer to southern Öland.

DESCRIPTION AND CLASSIFICATION OF THE RIDGES ON SOUTHERN ÖLAND

The Great Alvar of Öland on which most of the anticlinal and domal structures have been investigated is a slightly ESE-dipping plateau mainly consisting of the so-called orthoceratite limestone. A general description of the morphological patterns of the area is presented by Königsson (1968) in a comprehensive work on the Holocene history of the Great Alvar.

The generally very thin soil and the lack of soil in several areas makes the whole Great Alvar a karst area characterized by strong weathering and intensive jointing in sets of different frequency, size, and degree of opening. The presence of open joints and soil-filled joints, in which growths of grass, herbs and bushes is a common phenomenon, favoured the study of sets of joints using aerial photographs. Master joints and sets of major joints presented on the structural map (Fig. 5) were drawn from aerial photographs at a scale of 1:5000. Although the small thickness of soil favoured the discovery of the structures on the surface the often very slight divergences in dip of most of the ridges and domes from the gently dipping plateau made it difficult. However, the measurements of strike and dip of the limestone beds were facilitated due to the presence of ephemeral swamps or temporary pools, i.e. shallow water-filled depressions in the bedrock surface mostly dammed by scarps or sometimes gravel ridges. Here the extent and the depth of the shallow water in different parts of the pools or swamps clearly indicated the strike and dip of the bedding surfaces.

The main ridges

The divergent dips of the limestone beds at the ridges contrasting with the almost flat-lying surroundings were mentioned by Munthe (1902a, 1902b) in the description of the gological maps of southern Öland. In a description published somewhat later Hedström and Wiman (1906) made several references to divergent dips of the limestone beds at separate localities along what is now found to be the continuous ridge or anticline between Södra Bårby and Ås. This main ridge with a total length of almost 30 km strikes NNW—SSE in its northern part but turns somewhat towards a N—S direction in its southern part. The width of the ridge varies between 50 m and 300 m and the more or less horizontal denudation surface shows in several places rather steeply dipping older limestone beds in the centre flanked by successively younger and more gently dipping beds. Along the Södra Bårby—Ås Ridge several marked culminations occur which are indicated on the structural map by curves in the border lines, Fig. 5.

Close to the northern end of the Södra Bårby —Ås Ridge we notice a collection of regularly spaced circular domes with an average diameter of 10 m. In the same area we also find a row of round or slightly elongated depressions with smooth inwardly dipping edges. These depressions, often more than one metre deep, have been classified as dolines by Nilsson (1965). Similar structures are also found closely connected with the main ridge at Resmo. Many of these so-called dolines show strong evidence of a deep-seated denudation process due to glacier movements affecting the often strongly jointed limestones close to the main ridges. Hence, quasi-dolines would be a better term for these structures. Local abnormal dips of limestone beds due to weathering and wash out of loose material is a fairly common feature on the top of the scarps associated with the limbs of the main ridges.

Hedström and Wiman (op. cit.) also mentioned the main ridge between Resmo and Ekelunda which they classified as a "broken saddle" characterized by upwarping of older limestone beds into younger. However, this ridge is found to continue further east of the village of Ekelunda till it becomes covered by gravel at the Alvar border, 1.3 km W of Sandby church. Hence, the ridge is hereby called the Resmo—Sandby Ridge. In the area of Ekelunda the ridge bends slightly towards the north and becomes narrower. This narrowing accompanied by less pronounced dips of the limestone beds of the ridge as we move towards the northeast appears to be due to a thicker overburden of limestone. This ridge also shows several culminations.

The Dröstorp Ridge, cut by the outlet of the Dröstorps Mose and running in an ENE—WSW direction, shows a remarkable narrowing towards the southwest. This part of the ridge also differs somewhat from the ordinary pattern of the main ridges as regards the denudation and dimensions. Here, a narrow intact anticline rises about 2 m above the flat-lying surroundings and one and the same limestone bed encloses the structure. Hence, this part of the ridge belongs to a type of structure which is here called a hump-back ridge. Except for the southwestern part of the Dröstorp ridge, the main ridges as a whole show very similar structural patterns and they are all supposed to have had the same type of origin, i.e. gravitationally controlled piercement of the overlying, tectonically passive limestone beds by the underlying, tectonically active shales. The strong vertical forces associated with the vertical movements have led to the formation of *bending folds* (Ramberg, 1963b). Another striking characteristic is the remarkable change in direction of joint sets or sometimes joint systems (intersecting sets of joints) when we move towards the ridges from the outside (cf. the structural map, Fig. 5).

The main ridges with their strongly jointed hinge regions have offered a good opportunity to denudation during the last ice movements. These eroded parts form depressions in the almost flat-lying surroundings and have served as a suitable area for deposition of moraine and glaciofluvial materials when the land ice left Öland. Consequently, we now often find gravel deposits situated in, or close to, the centres of most of the main ridges (cf. the physiognomic map and figs. 10, 15, and 18 in Königsson (1968). Königsson found that many of the deposits still preserved were tectonically controlled and in a sketch map he also presented possible tectonic zones on the Great Alvar.

Looking at the main ridges as tectonic zones implies amongst other things a high frequency of joints and sets of joints and automatically gives the ridges a great importance for the subterranean drainage of large areas of the Great Alvar. Several productive springs are known from the main ridges, and wells situated within the ridges are never dry. Further, looking at the map showing the archaeological monuments within the Great Alvar (Königsson, op. cit. fig. 104, p. 147), we notice a strong concentration of groups of old house foundations, strongholds, and graves close to the main ridges and hump-back ridges, i.e. in areas which have offered reliable water resources.

The hump-back ridges

The proposed division of the ridges into main ridges and hump-back ridges is based upon differences in morphology, dimensions and to some

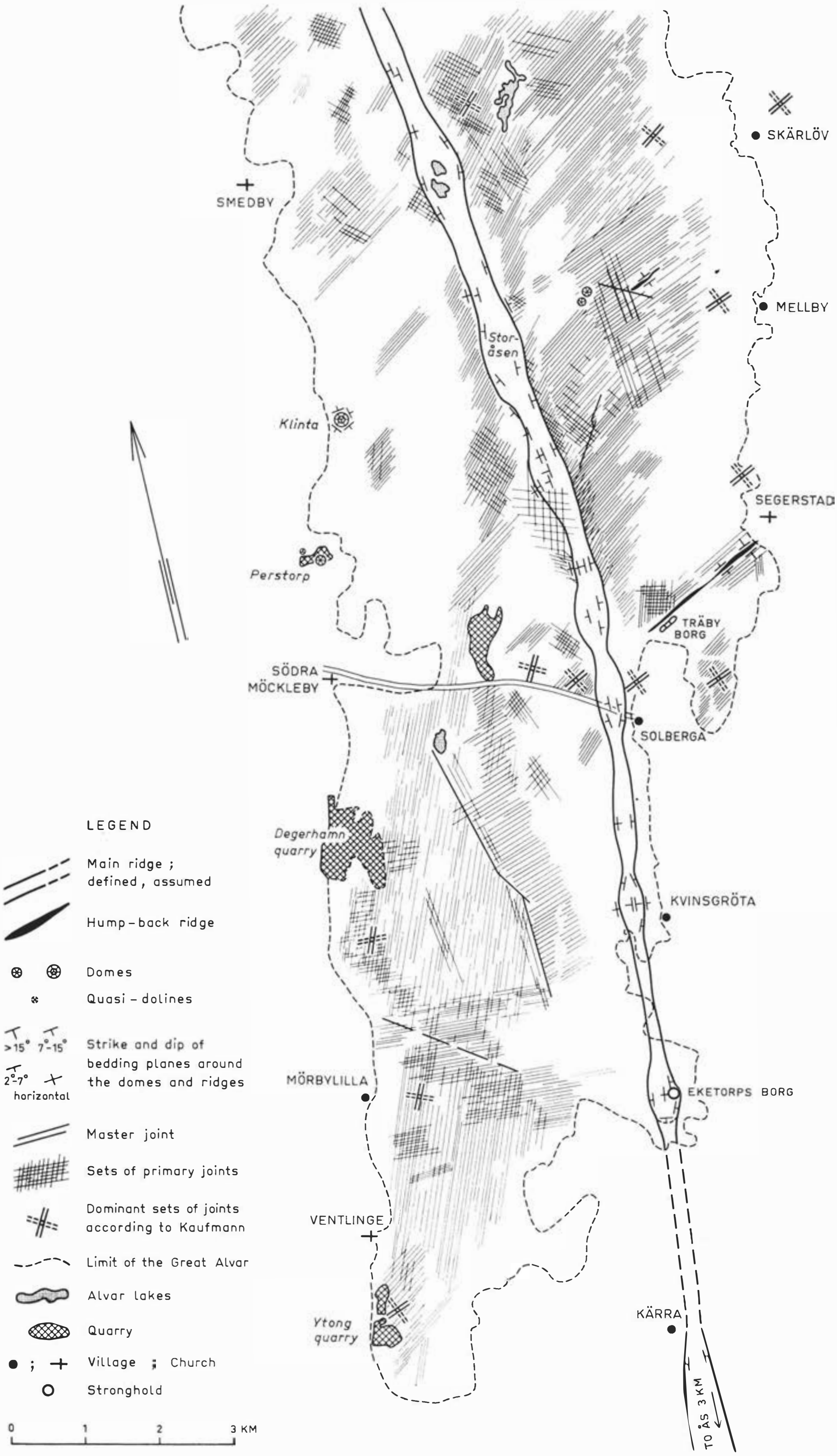
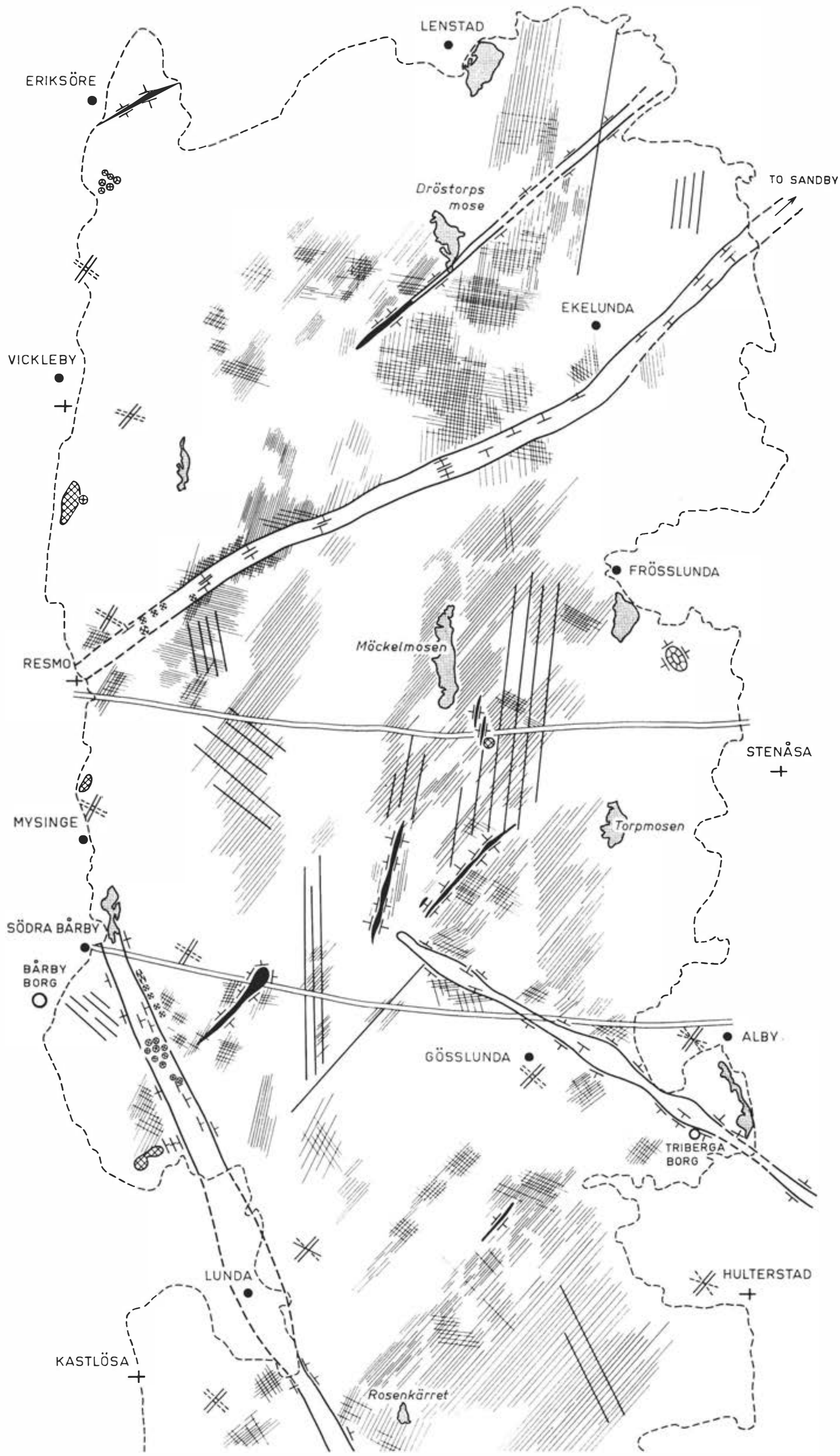


Fig. 5. Tectonic map of the Great Alvar, southern Öland. Mapped with the aid of aerial photographs. Notice the 'pinch-and-swell' structure of the main ridges and the change in direction of sets of joints in their vicinity.



Hump-back ridges are often oriented parallel with a surrounding set of joints. The domes have tangential and radial joints. The direction of primary and secondary joint sets according to Kaufmann (1931) are shown.

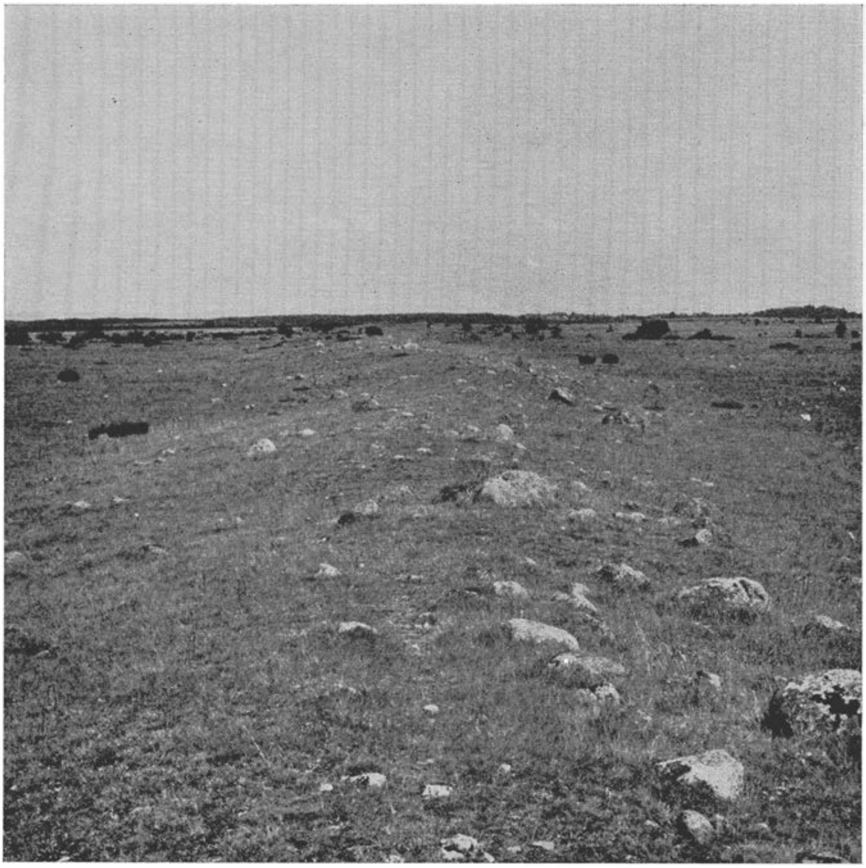


Fig. 6. Hump-back ridge. The so-called Hog's Back, situated 2.5 km NW of Gösslunda, central Great Alvar. A morphological description of the ridge is given by Bergsten and Svensson (1963).

extent origin. In contrast to the main ridges the hump-back ridges are intact anticlines consisting of upright folds of bedded limestones. These anticlines rise above the almost flat-lying surroundings and often a single limestone bed covers the whole structure. The length of the ridges varies from a few hundred metres up to almost two kilometres. The width or wavelength seldom exceeds 50 m and an average value of 10 m is rather representative for most of them, although values as low as a few metres do occur. The height or amplitude measured from the almost flat-lying surroundings varies from a maximum of 2 m for the largest ridges down to a few decimetres for the small ones. The areal distribution of the ridges is rather scattered although we notice a certain concentration in the middle part of the Great Alvar (Fig. 5). We also find that most of the hump-back ridges are located near and orientated parallel with the major single joint or joint set which occurs in the vicinity of the ridges. However, some of these parallel joints situated close to the ridges have either developed or become accentuated during the formation of the anticline. The majority of the ridges are straight except for two which show a slight and rather narrow offset. Culminations are also present along some of the ridges. Normally the ridges gradually die out in both ends. A morphological description of the most significant ridge, the so-called Hog's Back situated 2.5 km NW of Gösslunda, the central Great Alvar, is presented by Bergsten and Svensson (1963).

Origin of the two types of ridges

Besides the differences in morphology and dimensions between the two types of ridges there is also some slight difference in origin. The joint pattern close to the hump-back ridges indicates that the ridges were initiated at the site of early formed sets of joints. These joints have lowered the strength of the overburden and favoured growth of the ridge. Some of the sets of joints now observable in the vicinity of the hump-back ridges were certainly developed during the process of growth or folding and represent failure followed by a plastic deformation. A strict distinction between these two main types of joints is almost impossible due to the strong weathering.

The dimensions and general appearance of the hump-back ridges in relation to their stratigraphical position indicate relatively thin and shallow source layers. Since the thickness of the intercalations in the different limestone units seldom exceeds the scale of decimetres there must have been several layers of marl and shale which cooperated in the formation of the hump-back ridges. The same is true for the domes of intermediate size found at several places on Öland. The stratigraphical positions of the source layers for the hump-back ridges are shown in Fig. 4.

Along the main ridges we do not find any marked sets of joints running parallel with the ridges which could have initiated the formation of the ridges. From the structural map, Fig. 5, we notice a remarkable change in the number and direction of sets of joints as we near the main ridges. This characteristic is mainly thought to be connected wth the growth of the ridges and represents failure following the plastic growth.

The present position and direction of the main ridges seem to have been controlled by early tectonically formed weak zones radially diverging from the area of Mörbylånga just west of the border of the Great Alvar on the level with Resmo. Hence, some of the different sets of joints now visible in the vicinity of the main ridges could also be of another type developed in connection with the formation of the tectonic weak zones.

The dimensions and tectonic style of the main ridges together with their stratigraphical and geographical position indicate a formation where the light and ductile alum shales act as buoyant source layers.

Classification of the ridges

In this section we will first establish a description and classification of the geometry and morphology of the ridges which does not involve any genetic terms. Later we will find that in the case of the ridges on southern Öland there are also possibilities of introducing genetic terms.

Following the non-genetic description and classification of folds according to Ramsay (1967, p. 345) the terms that describe the parts and shapes of folds can be divided into two groups: those which refer to a single folded surface within the structure and those which describe the relationships of adjacent surfaces. In our case the *description of the single folded surface* will be mainly restricted to the exposed surfaces of the intact hump-back ridges. Generally, the crest and hinge lines here are coincident and run rectilinearly except for two of the hump-back ridges which show a slight offset. Culminations and depressions are fairly common and can be found along both types of ridges. The fold profiles of the single hinge folds, i.e. the cross sections of the folded surfaces perpendicular to the hinge line, are mainly of two types: (a) those with a narrow angular hinge zone often associated with a marked open joint running parallel with the hinge line, (b) those with a gently curved hinge zone, cf. Fig. 6. Both types of fold profile can be found in one and the same ridge. Very often both types of folds show straight limbs. The interlimb angle varies between gentle (180° to 120°) and open (120° to 70°) according to the fold description suggested by Fleuty (1964). Interlimb angle, wavelength and amplitude of the folds are to a large extent dependent upon the level of the exposed surface in relation to the top of the source layer. Wavelengths and amplitudes of the two types of ridges, presented in the previous section, are in both cases referred to the almost flat-lying surroundings as a median surface.

Information about the *relations of adjacent surfaces in the ridges* is almost lacking for the hump-back ridges whereas the planed off main ridges offer some information. From dip measurements across the main ridges the results point to almost vertical axial surfaces which bisect the interlimb angle which in turn implies that the ridges are symmetrical folds. Deviations from this rule are known from a few places, e.g. along a distance of about 2 km south of the bend of the Bårby—Ås Ridge and at some places on the bend of the Resmo—Sandby Ridge. Using the terms to describe the *attitude of folds* based on the dip of the axial surface and on the plunge of the hinge line (Fleuty, op. cit.), the ridges found on S. Öland can be classified as upright sub-horizontal folds. Because of the fact that décollement structures are often associated with *buckling* of one or several competent layers underlain by a much less competent layer the ridges show a striking similarity with a décollement structure. As in the décollement structures the limbs of the folded layers are flattened out and become horizontal on both sides of the axial surface. This feature together with a gradually increasing dip of the limestone beds towards the centre of the main ridges indicates a convergence of the trough and inflection surfaces towards the interior part of the structure.

The *morphological classification* of folds involves the shape of the profile section of the folded layers. According to Ramsay (op. cit.) it is always possible to express the changes in shape within the fold and to classify the folded layer using any of the parameters: orthogonal thickness, thickness parallel to the axial surface, and inclination of the dip isogons. From field observations there is clear evidence that the curvature of the inner fold arc of the main ridges always exceeds that of the outer arc. This is turn implies that the dip isogons converge on each other and on the axial trace of the folded layer as they are followed in towards the inner arc of the fold. Ramsay assigns folds with convergent dip isogons to his Class 1 which in turn can be subdivided into three subclasses depending on the strength of the convergence of the isogons. The folds of the ridges found on southern Öland can be classified among the two subclasses: 1 A, folds with strongly convergent dip isogons, and 1 B, parallel folds. In folds belonging to the subclass 1 A the orthogonal thickness of the limbs always exceeds that at the hinge of the structure. In parallel folds, subclass 1 B, the layers have a constant orthogonal thickness and the thickness of any bed measured parallel to the axial trace of the fold is always greater than that at the hinge. For further characteristics of the two subclasses of folds the reader is referred to Ramsay (op. cit., p. 367).

In order to obtain a successively increasing dip of the originally horizontally bedded limestone layers as we move on an arbitrary horizontal plane towards the central part of the main ridges it is likely that the layers possess one of the arrangements schematically shown in Fig. 7. In a folded sequence with no intercalations of plastic layers of marls and shales within the limestone pile the structure is likely to form by a successive thinning of each bed towards the hinge, e.g. case a in Fig. 7. The result is a fold shape in accordance with subclass 1 A. In a limestone sequence with intercalations of marls and clays the same main structure can be obtained by a plastic yielding of the incompetent intercalations and a brittle or semibrittle failure of the more competent limestone beds, cf. case b in Fig. 7. The shape of the profile section of the intercalations will here correspond to subclass 1 A. The stretching of the limestone beds due to the folding is at the same time compensated for by an equal amount of opening up of the joints. The morphology of the more competent limestone beds here fits into the subclass 1 B, i.e. a parallel fold. Notice that in both cases the structure dies out away from the buoyant layer.

In those parts of the Great Alvar where the main ridges occur the stratigraphic sequence mainly consists of alternating units of bedded limestones without intercalations and bedded limestones with intercalations of marls, clays and shales. Hence, it is most likely that the limestone beds in one and the same main ridge consist of both 1 A and 1 B types of folds. The final shape of each single layer and the whole structure is to a very great extent dependent upon the shape and dimensions of the underlying buoyant layer of alum shale. This is shown schematically in Fig. 7 c, d. Notice that in all cases there is an increasing dip of the limestone surfaces going horizontally towards the axial plane.

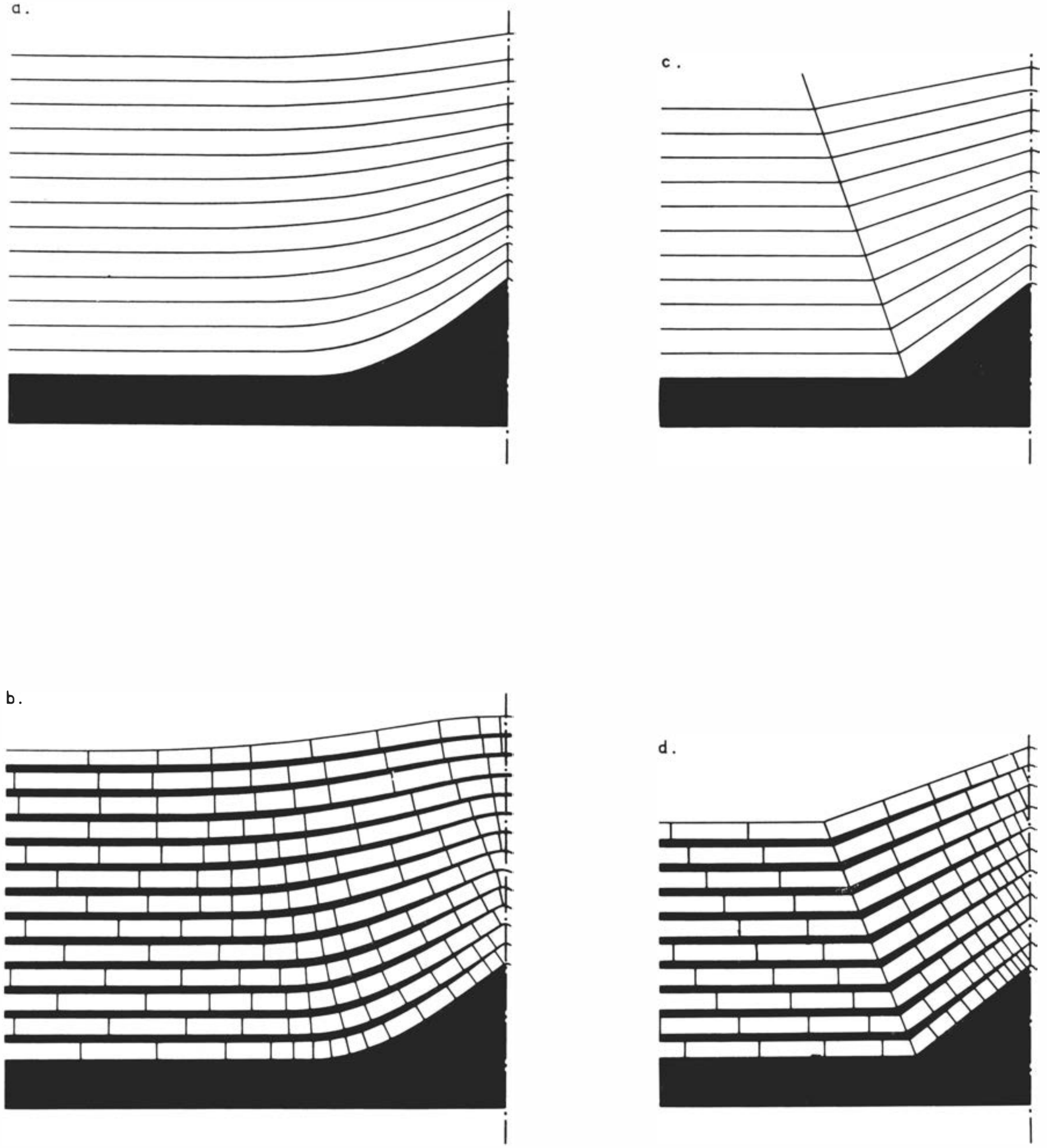


Fig. 7. Schematic sections of a main ridge.

a. Folded sequence of limestone without intercalations; each layer has convergent dip isogons, subclass 1 A.

b. Folded sequence of limestones with intercalations of plastic layers of marls or shales; the limestone layers form parallel folds, subclass 1 B; the intercalations have a fold geometry belonging to subclass 1 A.

c. Same sequence as in (a) but with another shape of the surface of the buoyant layer.

d. Same sequence as in (b) but with another shape of the surface of the buoyant layer.

Notice that in all cases the structure dies out away from the buoyant layer of alum shale; further note the increasing dip of the limestone surfaces as we move horizontally towards the axial plane.

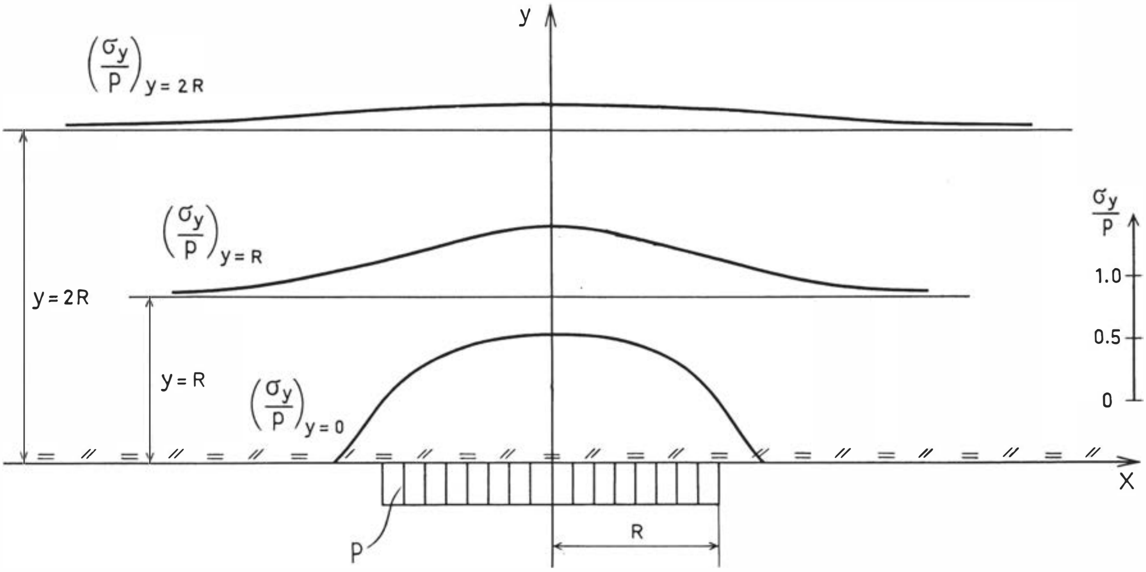


Fig. 8. Normal stress distribution in the centre of an elastic, homogeneous structure loaded perpendicular to the surface; R is the radius of the load, p is the load and σ_y is the normal stress.

In principle there are two ways in which an originally straight layer may be bent into folds; longitudinally or transversely, with respect to the layer. In the case of longitudinal bending the bed will buckle under a compressive deviatoric stress that acts parallel to the bed, and the folds developed are called buckling folds. If the bed is bent transversely by deviatoric stresses normal to it bending folds are formed. These two mechanisms and accompanying types of folds are fundamental in the *genetic classification of folds.* The characteristic features of these two classes of folds in layered rocks have been successively studied by Ramberg (1963b). Further, a third class of folds generated by accentuation of initial curvatures by homogeneous strain has been suggested by Flinn (1962) and Ramberg (1964). The genetic classification of folds, mainly based upon the results obtained by Ramberg, has also been discussed by Ghosh (1968) in connection with buckling experiments of multilayer systems.

According to Ramberg (1963b) buckling folds may be distinguished from bending folds by (a) the pattern of the contact strain, (b) the contrast in competence of the folded layer vis-à-vis the adjacent rock and (c) the commonly existing simple relationship between the length of arc and layer thickness of bucking folds.

Dealing with structures formed by a mechanism of bending, a characteristic feature of the heterogeneous strain around a competent body enclosed in less competent rocks is the slight penetration of the contact strain into the surroundings. "The tendency of the incompetent rocks to conform to the contours of the competent body is not noticeable outside a boundary zone that is not thicker than the 'wavelength' of the irregularities", says Ramberg (1963b, p. 7). Bending folds associated with competent inclusions are rather common structures e.g. in stratified rock adjacent to rows of boudins.

Bending folds can also originate from local deviatoric stresses acting perpendicular to a certain bed in a sequence of layered rocks. Since the beds remain in their original position at a certain distance from the applied deviatoric stress the surface of the beds that are bent becomes greater than it was in its original position; hence, each bed undergoes extension as the bending structure becomes accentuated. The deformation of any individual bed is caused by the transmitted pressure from the underlying strata applied over its entire surface. This pressure will gradually decrease as we move away from the area of the applied deviatoric stress. The situation can be illustrated by means of the Boussinesq equations which are well known among strength-of-material scientists. In Fig. 8 the normal stress distribution is shown parallel with the y-direction of an infinite two-dimensional elastic and homogeneous structure subjected to a distributed load along a limited part of its free surface. For a layered structure each such curve of normal stress as shown in Fig. 8 will penetrate slightly deeper into the structure than the equivalent curve in the homogeneous structure (Sonntag, 1957). The gradually decreasing stresses in the y-direction away from the applied load are reflected in the decay of the normal stresses parallel with the x-direction, although this is not illustrated in the figure.

The net effect of decreasing stresses away from the loaded area is a decreasing amount of contact strain. This in turn leads to a gradually decreasing bending amplitude of each layer as we move away from the area of applied deviatoric stress. In Fig. 8 we also notice the close similarity between the length of the loaded part of the structure and the distance to a region of greatly reduced normal stress in the y-direction. This situation is in close agreement with the conclusions obtained by Ramberg (1962, p. 411 and 1963b, p. 7); namely, that the finite displacements in the area surrounding a single buckled layer or a bent competent body are scarcely noticeable outside a boundary zone that is thicker than the wavelength of the fold or the competent body. Applying the theoretical results concerning contact strain in different geological structures to the two types of ridges we conclude that the commonly open interlimb angle (120° to 70°) and the moderate wavelength (average 10 m) of the hump-back ridges indicate a rather shallow source area for the deviatoric stresses. Hence, the source layer for the gravity formation of the hump-back ridges must be searched for among the marly and shaly intercalations in the different limestone units. The main ridges on the other hand, with their larger dimensions and gradually decreasing amplitude in the direction of increasing overburden thickness point to a deeper source layer. In the case of the main ridges the source layer must be searched for in the Upper Cambrian and Lower Ordovician alum shales.

The change of thickness of originally uniform layers subjected to bending forces is directly opposite to the changes of layer-thickness in folds formed by buckling (cf. Ramberg, 1963b, figs. 7 and 8). Results from model experiments of buckling multilayers consisting of competent sheets with considerable or restricted ease of sliding often show chevron-type folds in the centre of the model giving way to conjugate and single-hinged, sharp-crested folds away from the centre (Ghosh, 1968, fig. 5 and 14). Almost identical chevron-like folds were earlier obtained by Ramberg (1962) in rubber models consisting of multilayers of thin sheets of stiff rubber interlayered with rather thick sheets of soft rubber. Most of these buckling folds are parallel-sided with more or less constant interlimb angles, thus differing distinctly from bending folds.

The theoretical results combined with results from model experiments are to a great extent applicable to the ridges found on southern Öland. Solely on the basis of their geometrical and morphological patterns the general appearance of both types of ridges indicates an origin according to the mechanism of bending. On the basis of a genetic classification the ridges are bending folds or supratenuous folds. The evidence for a bending mechanism in the formation of the ridges is presented in a later section.

Another genetic classification of the ridges can be obtained if we introduce the concept of diapirism. As early as 1907 Mrazec introduced the term diapiric fold for those folds in which the sediments of the core pierced the overlying beds (Mrazec, 1915). To-day the terminology of diapiric structures is rather confusing. However, some of the terms introduced by Hoen (1964) are applicable to the structures on Öland. Hoen classified anhydrite diapirs on Axel Heiberg Island into three main groups, based on the structural setting in which they occurred — domical diapirs, anticlinal diapirs and fault diapirs. By making use of the term anticlinal diapir in describing the ridges on southern Öland both the main morphology and the mechanism of origin of the ridges are evident. Hence, the term single-layer anticlinal diapir and multi-layer anticlinal diapir are suggested for describing the main ridges and the humpback ridges respectively. Single-layer here implies one layer of buoyant material.

DESCRIPTION AND CLASSIFICATION OF THE DOMES OF SOUTHERN ÖLAND

The dome structures found on the island of Öland range in size from several kilometres in diameter — such as the Mossberga Dome situated on central Öland — down to not more than a couple of decimetres. The domes of southern Öland appear in nearly all the different stratigraphic units of the limestone sequence from the Lower Ordovician Hunneberg Stage to the Middle Ordovician Seby Limestone. Further, if we take into consideration the source layers for some of the large domes still older units are affected.

Most of the domes are found along the western part of the Great Alvar. Stratigraphically they are all situated between the limestones of the Hunneberg Stage and the *Lepidurus* Limestone. The exposed parts of the two domes at Melby and the slightly elongated dome just NW of Stenåsa on the eastern part of the Great Alvar all belong to the Segerstad Limestone, Fig. 4. Outside the Great Alvar domes have been found along the shore of the Baltic sea just north of Hålnäs, 2 km SE of Sandby. The exposed parts of the domes belong to the Seby Limestone. Small and very gentle dome structures appear along the low sea-cliff at Ottenby on the eastern shore of the Baltic on southernmost Öland. The structures, which have been assigned to the Billingen Stage, are mentioned by Lindström (1963).

Description of the domes

One of the largest domes found on southern Öland, apart from a possible dome-like culmination or uplift in the area of Mörbylånga, is situated close to the western border of the Great Alvar at the village of Klinta. The almost circular dome has a diameter of around 200 m. Most of it is covered with thin soils, but on aerial photographs and to some extent in the field the almost circular form is observable from scarps. From the outer border of the dome towards the centre the outwardly dipping beds with their scarps form a structural pattern of concentric rings which are clearly visible on the aerial photographs. The dip of the limestone beds around the border averages 12° with a maximum of 15° towards the east i.e. in the main direction of dip of the whole Great Alvar plateau. From the border towards the centre of the dome we reach successively younger beds of *Lepidurus* Limestone which at the same time are marked by slight increases in dip. The large diameter of the dome in relation to its stratigraphical position indicates a doming process with the alum shales as the active source layers. The thickness of the alum shales in the area of Klinta is estimated to be about 10 m.

A dome with almost the same dimensions as the one at Klinta but with a slight elongation in a NW—SE direction is situated close to the eastern border of the Great Alvar halfway between the durch of Stenåsa and the village of Frösslunda.

The domes next in size are situated near the central part of the Great Alvar 2.5 km east of the village of Mellby, Fig. 5. The three domes which could be discerned with certainty are situated in a row striking almost N—S. The domes have an average diameter of 20 m and the distance between each pair is around 50 m. The height or amplitude of the domes varies between one and two meters measured from the almost flat-lying surroundings. The very smooth and regular shape of the domes is readily seen in Fig. 9. Here, a stone-wall almost one metre high runs right over the centre of the central dome. The domes mostly have a thin soil covering except near their apices. The exposed parts of the domes stratigraphically belong to the Segerstad Limestone (Larsson, personal communications).

The next group of domes of almost equal size, here called *intermediate domes,* have diameters ranging within the interval 5—15 m. They are all situated along the western part of the Great

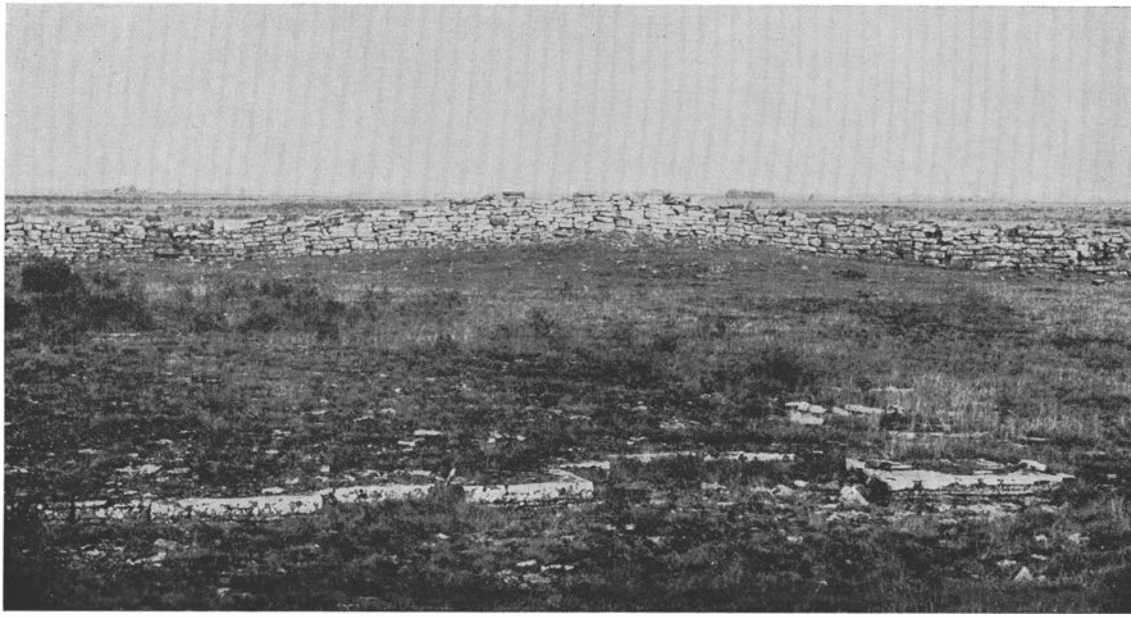
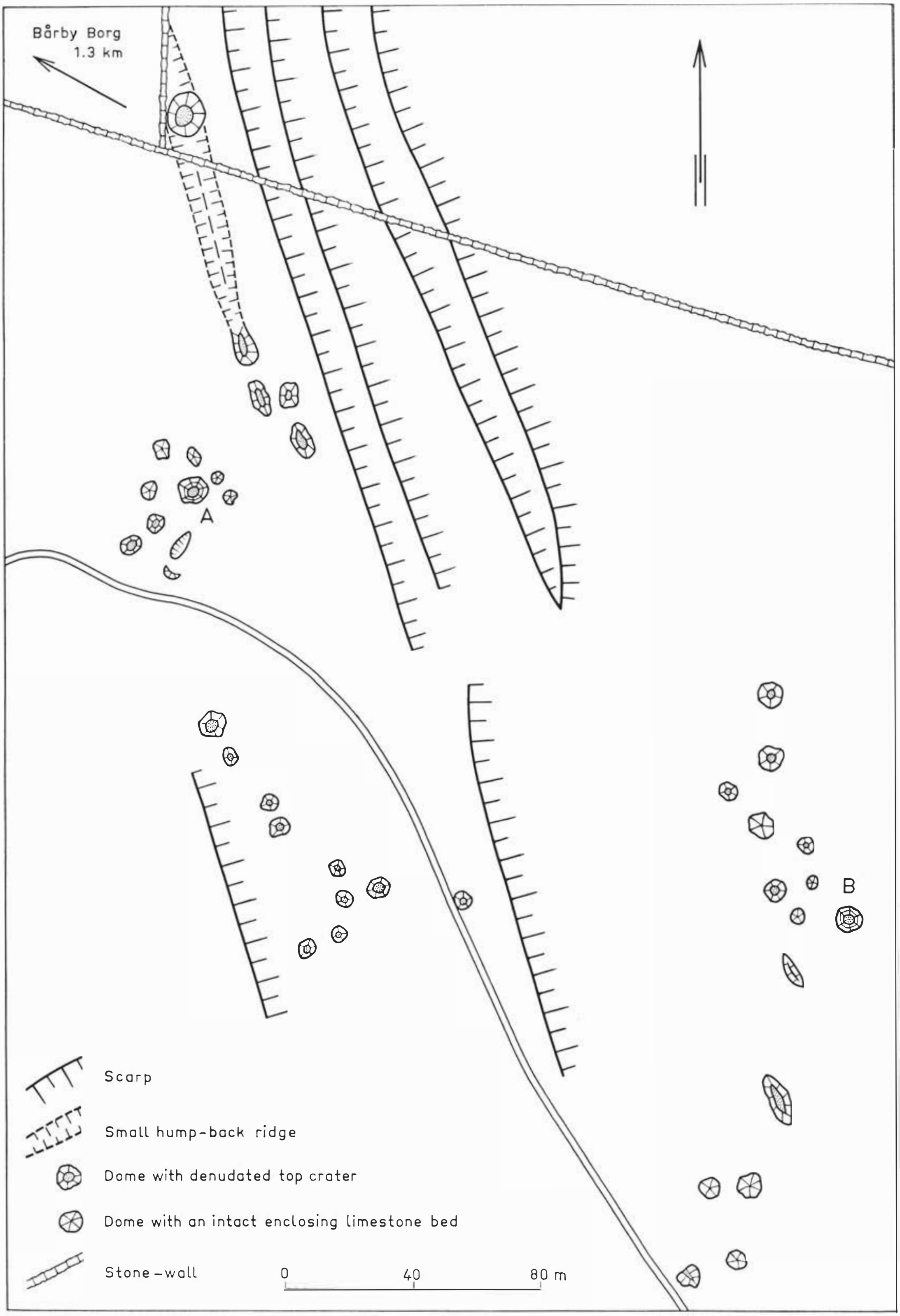


Fig. 9. Dome in Segerstad Limestone 2.5 km E of the village of Mellby. The height of the stone-wall is about one metre.

Alvar and the limestone in which they occur range from the *Ceratopyge* Limestone up to the *Lepidurus* Limestone. In a short article on "Dome-like uplifts on the Great Alvar" Palmqvist (1965) was able to identify not less than 50 separate dome structures within a limited area 1.3 km ESE of the stronghold at Södra Bårby. The domes are more or less concentrated on the central or eastern part of the main ridge which runs between Södra Bårby and Ås. A striking pattern of this area is a certain concentration of several domes into separate domains often with very regular spacing between the domes within each domain, Fig. 10. In general there are two types of surface structure found: (a) domes with denuded top craters and scarps along the flanks, (b) intact domes where one and the same limestone bed encloses the structure. The outward dip of the dome flanks and the height or amplitude measured from the flat-lying surroundings are slightly greater for the first type of domes. On average the outward dip is 15° and the amplitude varies in between 20 and 50 cm. In Fig. 11a dome belonging to the first group is shown; notice the almost axi-symmetrical shape, the outwardly dipping limestone beds with their escarpments and two sets of roughly discernable joints, one set radiating out from the centre of the dome, the other developed tangential to the circular structure. The joint systems associated with the domes are distinctly different from the main sets of joints in the neighbourhood. The gravel deposit in the central crater consists of a mixture of angular fragments of limestone not far transported together with well-rounded pebbles of pre-Cambrian rocks. Although the domes in each domain show a very regular spacing that is slightly more than the mean diameter of any adjacent pair of domes, we notice a certain variation in diameter among them. In most cases we find a couple of big domes of crater type (a) with a diameter of about 15 m surrounded by slightly smaller domes of both (a) and (b) types. Several domes are surrounded by a slightly developed rim syncline and in an area between two closely spaced domes the syncline often becomes strongly accentuated.

In Fig. 10 we further notice the occurrence of elongated domes which are of type (a) and


Bårby Borg
1.3 km
A
B
Scarp
Small hump-back ridge
Dome with denudated top crater
Dome with an intact enclosing limestone bed
Stone-wall
0
40
80 m

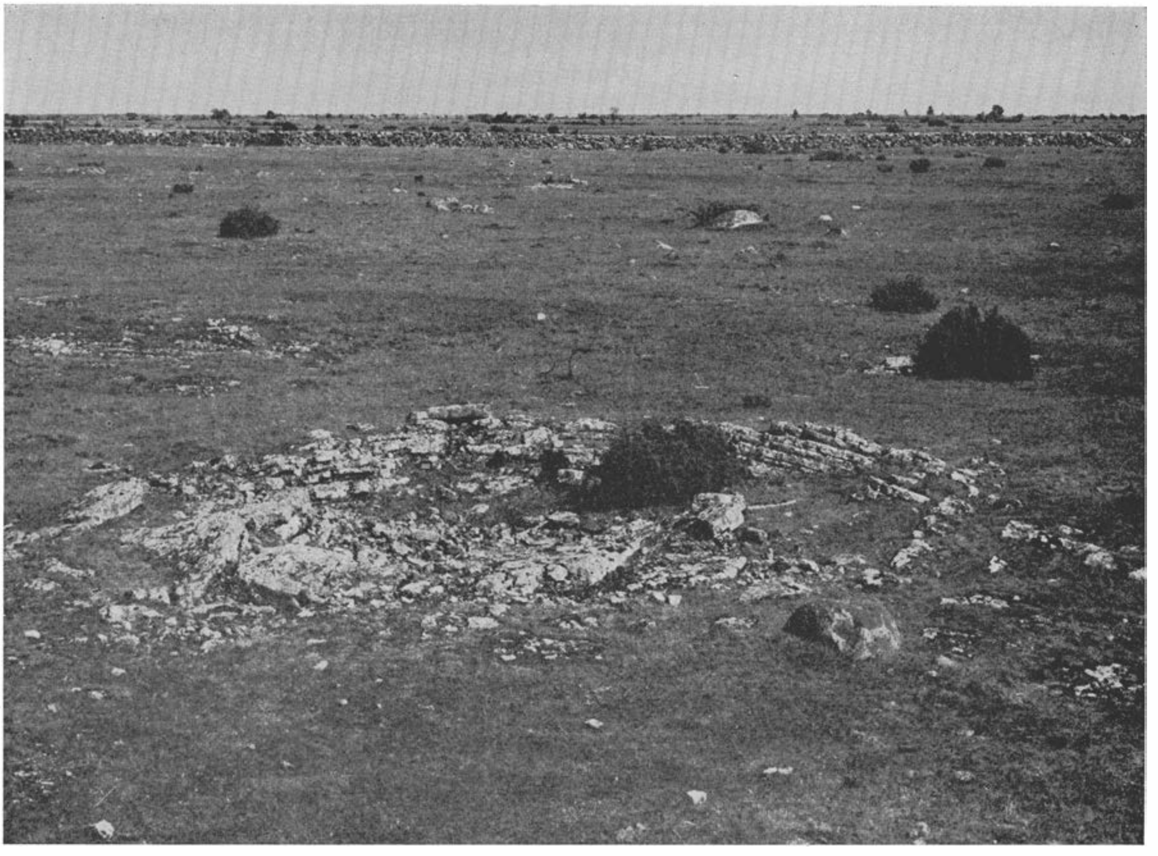


Fig. 11. Dome with denudated top from the dome area 1.3 km SE of Bårby stronghold. The dome is marked A in Fig. 10. It has a diameter of 8 m and is in the *Vaginatum* Limestone.

(b). Within the area of dome development we also find a small hump-back ridge which at both ends are locked by domes of the (a)-type. The appearance of domes in combination with a ridge is of great importance for understanding that the mechanism of origin for both types of structures is similar.

One of the domes, marked with the letter B in Fig. 10, was dug out and described in detail by Palmqvist (1965). At a depth of about one metre beneath the surface he reached a zone of "soft" limestone which could easily be broken into pieces. The clay content increased downwards and no marked changes in dip of the limestone beds were observed down to 1.25 m beneath the surface. Palmqvist believes in dome formation under the action of vertical forces but finds it difficult to explain them as pre-Glacial structures. Extension by freezing of water and weathered materials in cavities — frost tectonics — is a likely mechanism according to him. Undoubtedly, most of the domes and ridges have been affected by such a process which in turn has accentuated their structural pattern, but an origin due to frost tectonics seems quite improbable.

Another part of the Great Alvar with apparent similarity to the dome area at Södra Bårby is situated at the southwestern border of the small hills of Tingstadbackar, 1.1 km S of the village

Fig. 10. Intermediate domes and a small hump back ridge 1.3 km SE of Bårby stronghold. Notice the tendency of domes to form along the hump-back ridge and the rather regular spacing of the domes within each domain. Dome A is shown in Fig. 11, and dome B in Palmqvist (1965, figs. 1, 2).

Fig. 12. Slightly elongate intermediate dome in the quarry 1.3 km south of the church of Vickleby. The levelling staff is 1.4 m.

of Eriksöre, cf. Figs. 5. Here more than 20 separate domes could be counted within an area of 100 m by 50 m. Like the situation at Södra Bårby both types of domes appear (i.e. (a) and (b)). Also, the spacing and amplitude of the domes show great similarities in the two areas.

A three-dimensional exposure of a dome with a surface structure similar to that of the type (b) domes found in the areas of Södra Bårby and south of Eriksöre, was discovered at the southern wall of a small quarry close to the main road at the village of Perstorp. Here a vertical section, almost three metres high, through the central part of a circular dome was exposed along an east-west striking joint surface. Unfortunately, the dome structure is now covered with refuse.

A dome of intermediate size and partly exposed in three dimensions was found in the easternmost parts of a small quarry situated close to the eastern side of the road 1.3 km south of the church of Vickleby. The dome is slightly elongated in an ESE—WNW direction with its westernmost parts exposed in the quarry, Fig. 12. The very top of the Limbata Limestone consists of a hard reddish-brown calcilutite, most capable of resisting weathering. A thin layer of soil covers the structures outside the borders of the quarry. At the time of quarrying the limestone slabs, the stone-breakers exposed part of the dome and then stopped due to the large number of joints. The steadily increasing frequency of joints towards the dome is readily seen in Fig. 12. Also notice the tendency for the development of sigmoidal tension gashes or pinnate tension joints on the left hand wall. These have developed more or less *en échelon* in some of the limestone beds. The simultaneous appearance of pinnate joints together with the almost planar sub-vertical joints indicates a combined tension and simple shear mechanism for the deformation of each layer in the limestone sequence. At the exposed part of the curved flank of the dome the radial and tangential sets of joints are clearly visible and the exposed flank consists of piled-up 5—10 cm thick limestone slabs with an outward dip averaging 30 degrees. A slight rim syncline is also detectable around this dome.

Outside the Great Alvar, domes of intermediate size have been found along the shore of the

Baltic sea at the small village of Hålnäs, 2 km SE of Sandby. The morphology of the shore where the domes appear distinctly differs from the huge, planar and gently dipping limestone surfaces so characteristic along most of the eastern coast of Öland. At Hålnäs wave action has broken up blocks from the strongly jointed domes. These blocks now form a border along the shore. Along the exposed part of the shore and in shallow water areas a total of nine domes were discovered all in the Seby Limestone. The general appearance of these domes is similar to the one previously described from Södra Bårby and south of Eriksöre.

The type of domes that come next on a basis of size are the *small domes* on the scale of decimetres. On southern Öland small domes appear along the sea-cliff at Ottenby, situated on the eastern shore of the Baltic, southernmost Öland. The domes at Ottenby and similar structures at Perstorp (now filled with refuse) have been described by Lindström (1963). Since the type locality of small domes is situated at Horns Udde on northern Öland these structures will be treated in more detail in a later section.

Classification of domes

The present terminology dealing with diapirism or dome formation is limited, confusing and mainly restricted to areas of salt deposits. The terms most commonly used are salt dome, salt anticline and salt uplift. Each has been used in a different sense by different writers. Piercement or diapiric structures can have many different forms. There are waves, pillows, domes, mushrooms, teardrops and plugs developed mostly in areas of little or no orogenic activity, e.g. the United States' Gulf Coast or the Hannover basin in northern Europe. In orogenic belts several more types of diapiric structure appear. Even within areas which have been tectonically stable there may be great differences in forms among individual diapirs as the results of variation in the original thickness of the source layer or overburden. Salt plugs formed in areas of thick salt, and salt pillows formed in areas of thin salt are common features in the basins of the Gulf Coast and in Germany.

If we apply the most commonly used terminology in salt tectonics to describe the domes found on southern Öland they could be classified as pillows. Here, a pillow is characterized by a slightly bent surface of the source layer with gradual decrease in dip away from the centre. In salt tectonics the pillow forms as a primary stage in the process of gravitational instability of the less dense salt layer overlain by a heavier overburden. The truly intrusive domes, e.g. plugs, mushrooms and teardrops, form only when this pillow stage has "matured", thus representing a second phase in the process of gravitational instability.

Hoen (1964) divided the anhydrite diapirs of Axel Heiberg Island into three main classes — domical diapirs, anticlinal diapirs and fault diapirs. According to him the domical diapirs occur in tectonically undisturbed areas and are surrounded by essentially flat-lying strata. None of the terms pillow or domical diapir imply any restriction as to size or dip of the surrounding strata. However, the term domical diapir is more suitable than pillow in classifying the domes on southern Öland. As for the ridges, which are classified as anticlinal diapirs, the term domical diapir gives both the general morphology of the structure and the mechanism of origin. The purely descriptive terms, large, intermediate and small which were used to qualify domes in the previous section only refer to the diameter of the domes.

In the case of salt tectonics a complete sequence of salt commonly forms the source layer for the gravitational growth of a pillow or a salt dome, e.g. the Zechstein evaporites in Germany. A similar situation holds for the main ridges and the large domes on southern Öland, but instead of salt the "mother" bed or source layer consists of alum shale.

The situation is somewhat different for the hump-back ridges (anticlinal diapirs), the intermediate domes and the small domes (domical diapirs). Their dimensions, wavelength or diameter, amplitude, and outward dip of the limestone beds, in relation to their stratigraphic position preclude the possibility of gravitational formation which involves the alum shales. The principal

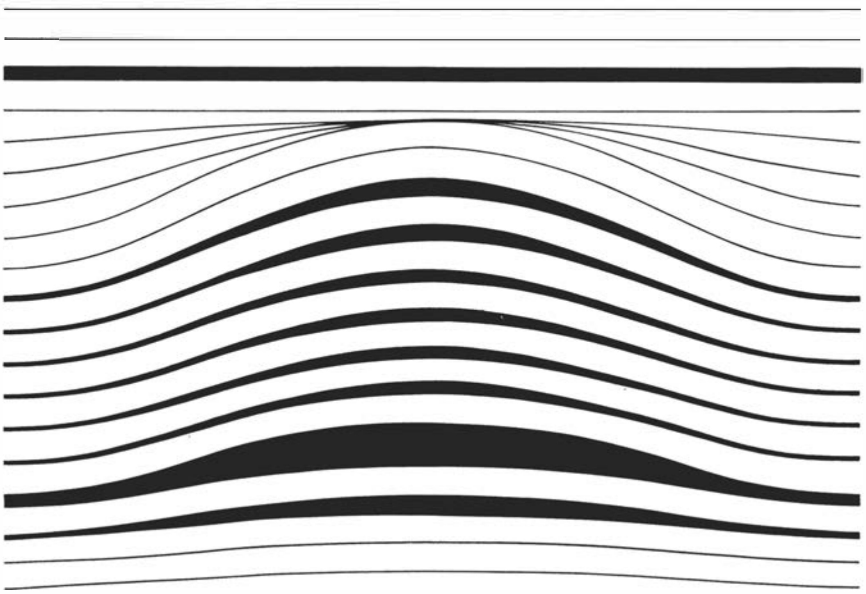


Fig. 13. Schematic section of an intermediate dome. At an early stage the dome occurs in the form of a gentle bulge. The convergent flow of the plastic intercalations will cease when the limestone surfaces come into mutual contact or when the plastic layers reach a certain thickness. A standstill in growth is followed by horizontal sedimentation. The term multi-layer domical diapir is suggested for this structure.

mechanism of origin is the same, i.e. gravitational instability of a layered sequence of rock, but in this case several layers of marl and shale forming intercalations in the limestone units, have co-operated in the doming process. Instead of one source layer consisting of a more or less complete sequence of almost the same rock type the source layer for the hump-back ridges and the intermediate and small domes is composed of alternating layers of limestone and marl or shale.

The spontaneous rise of the buoyant media in the field of gravity is in the form of a perturbation whose wavelength depends upon the total number of layers in the multi-layer system and on the layer thicknesses and rheological properties. Perturbations with a "wavelength" either larger or smaller than a certain characteristic wavelength for the multi-layer system rise less rapidly and are gradually adsorbed by the fastest growing perturbation. Equations for the dynamics of this type of layered system in the field of gravity have been developed by Ramberg (1968).

A sketch of a section of an intermediate dome is shown in outline in Fig. 13. The geometry depends on the stage of evolution of the domal structure. At a very early stage the dome appears in the form of a gentle bulge. As the convergent flow of marly or shaly materials increases towards the center the structure becomes more accentuated. This convergent flow of plastic materials will cease as soon as the limestone surfaces above and below the plastic layers come into contact or when the plastic layers reach a certain minimum thickness. When the doming process has come to a standstill sedimentation with horizontal bedding surfaces will continue. A domical diapir showing subsequent sedimentation during the latest stage of growth was found on northern Öland, cf. Fig. 18, p. 71. Here horizontal sedimentation after the cessation of growth is evident from the horizontal discontinuous bedding surfaces situated directly above the top of the dome. The term *multi-layer domical diapir* is suggested for this type of structure. This is in order to distinguish between domes formed by gravity instability of a multi-layer sequence from domes formed by instability of a single layer.

The hump-back ridges have an origin very similar to that of the domes, except for the phase of initiation. As already mentioned, the ridges were initiated by single joints or sets of joints whereas their actual growth was due to gravitational instability of the marly and shaly intercalations in the different limestone units. Commonly, these ridges gradually die out towards both ends, but sometimes they form closed anti-clines with a well-rounded nose at one end (cf. the hump-back ridges 1 km east of the village of Södra Bårby, Fig. 5). Also, transitional structures of ridges and domes appear such as the small hump-back ridges from the dome area SE of Södra Bårby, Fig. 10. Here, a nose dome is situated at the southern end of the ridge and a domical culmination appears in the center of the ridge. Hence, conforming with the classification of intermediate and small domes, the term multilayer anticlinal diapirs is suggested for these ridges.

AGE OF FORMATION OF THE DIAPIRIC STRUCTURES ON SOUTHERN ÖLAND

The stratigraphic position of the source layers active in the formation of the different structures was presented in Fig. 4. Possible ages of formation of the diapiric structures — in time-stratigraphic terms are summarized in Fig. 14. The time-stratigraphic classification of the Lower and Middle Ordovician is in accordance with the classification of Jaanusson (1960b). The dating of the different types of diapirs is based solely upon the dimensions of the structure in relation to its stratigraphical position, e.g. an intermediate dome which is found to consist of limestone of Kundan age is most probably of Aserian age since a certain amount of overburden material is required before the diapiric growth can start.

Whatever theory geologists have proposed for the origin of diapiric structures they have needed to search for some mechanism of initiating the growth of the diapir. Irregularities in the basement rock beneath the source layer, basement rock movements, variation in thickness or composition of the overburden or source layer and any type of faulting are the most common mechanisms involved. The physical and mechanical aspects of this problem with respect to the structures on Öland will be treated in detail in a forthcoming paper.

The strong similarity in the general pattern of the main ridges indicates simultaneous initiation and formation. The directions of the ridges diverging outwards from a rather limited area precludes the possibility of a formation by a mechanism of horizontal compression, and further, the tectonic style on southern Öland is contradictory to such an explanation. The main ridges are supposed to be initiated by faulting along zones radially diverging from the area of Mörbylånga. There is strong evidence that this faulting can be connected with a local tectonic uplift in the area of Mörbylånga. The appearance of Lower Cambrian sandstone and the possibly abnormal thickness of the *Oelandicus* Beds in this area strengthen this explanation, provided that the observation of undisturbed Lower Cambrian sandstone at Mörbylånga is correct, cf. Martinsson (1965, p. 188). Expressing the age of this faulting more exactly is impossible from local evidence. However, an Idaverean age is suggested because certain tectonic activities are known to have taken place in adjacent areas at that period of time, e.g. the early development of the middle part of the Caledonian geosyncline and spreading of volcanic sediments now appearing as layers of metabentonite in the *Ludibundus* division of the Idaverean Stage (Magnusson et al., 1962).

Before the actual growth of the main ridges could start the overburden material must have reached a certain thickness, and further the thickness ratio of the source layer to the overburden is a controlling factor of the growth. From the isopachyte map, Fig. 3, we know that the thickness of the alum shale gradually increases eastwards. Provided that the whole sequence of alum shale is involved in the gravitational formation of the main ridges this implies a great span in the time of formation. Differences in parameters

Time-stratigraphic units				Anticlinal Diapirs	Multilayer anticlinal Diapirs	Domical Diapirs	Multilayer domical Diapirs
	Series	Subseries	Stages				
UPPER ORDOVICIAN	Harjuan			Main ridges between Södra Bårby—Ås Resmo—Sandby Dröstorp (eastern part)			
MIDDLE ORDOVICIAN	Viruan		Oanduan				
			Keilan				
			Jõhvian				
			Idaverean	Possible initiation of the main ridges by radially diverging faults from the area of Mörbylånga		Large dome 2 km NE of Stenåsa	
			Kukrusean		Hump-back ridges at Segerstad and Mellby	Large dome at Klinta	Domes at Mellby
			Uhakuan				
			Lasnamägian		Hump-back ridges on central part of the Great Alvar and at Eriksöre		Intermediate domes 2 km SE of Sandby
			Aserian		Intermediate hump-back ridge at Bårby stronghold		Intermediate domes at Bårby stronghold and 1.1 km S Eriksöre
LOWER ORDOVICIAN	Oelandian	Ontikan	Kundan				Intermediate domes 1.3 km S Vickleby and at Perstorp
			Volkhovian				
			Latorpian			Small domes at Perstorp and Ottenby?	Small domes at Perstorp and Ottenby?
		Tremadocian					

Fig. 14. Age of formation of the diapiric structures on southern Öland.

such as strength, effective viscosity and density will also influence the growth. Hence, a precise dating of the growth of the anticlinal diapirs must be left open for now.

There is strong evidence that the initiation of the multi-layer anticlinal diapirs — the hump-back ridges — took place along almost straight joints, and further, that the subsequent growth was fairly rapid. The intermediate hump-back ridge 1.3 km ESE of the stronghold at Bårby was initiated and grew at the same time as the multi-layer domical diapirs in this area.

The dimensions of the large domes in relation to their stratigraphical position indicate a domical diapirism with the alum shales acting as the source layer. The mechanism of initiation of these structures must be searched for among the known variations in thicknesses and compositions of the material in the source layer and the overburden. Where these domes appear the thickness of the Tremadocian shales is rather limited and the commonly very irregular boundary with the underlying Olenid Shale must have have afforded good opportunities for initiation of the diapirs. The irregular boundary is due to the great proportion of stinkstone in the Olenid Shale.

As in the case of domical diapirs the multilayer domical diapirs have been initiated by irregularities in the source layer or the overburden. The present situation with irregular appearances of shaly and marly intercalations in the different limestone units accounts for the good opportunity for dome initiation during the Lower and Middle Ordovician.

The mechanism and age of formation of the small domes will be treated in detail in connection with the description of the small domes at the type locality Horns Udde, northern Öland. Although the basic mechanism of formation is the same for the different structures shown in Fig. 14 the small domes depart from the general pattern as far the rheological properties of the limestone beds is concerned. The structures of large and intermediate size indicate that the limestone beds possessed brittle or semi-brittle properties during the growth whereas the structure of the small domes clearly points to the limestone bed having had a plastic or semi-plastic property during dome formation. The layers of marls and shales on the other hand always show a plastic behaviour independent of the size of the structure.

The possibility of postulating the age of formation of the structures is sometimes hampered due to accentuations of the structural pattern caused by frost activity and weathering.

STRATIGRAPHY AND TECTONICS OF CENTRAL AND NORTHERN ÖLAND

Most parts of central Öland — here taken as the region between the latitude of Torslunda and at just north of Borgholm — are concealed by Quaternary deposits, and only in small and scattered places are the rocks below accessible for observation. The same situation is true for the eastern and northernmost parts of northern Öland. The main details of the stratigraphy show striking similarities with the situation on southern Öland. The geological map of central Öland, Fig. 15 left part, is mainly compiled from Hedström's and Wiman's map (1906) at the scale of 1:200 000 and Jaanusson's (1960a) notes on the Viruan of central Öland. North of Föra on northern Öland the map (Fig. 15 right part) is compiled *in extenso* from Jaanusson (1960a); here the western part was mapped by B. Bohlin and the eastern part by V. Jaanusson.

The oldest rocks found on the central part of the island — the *Eccaparadoxides oelandicus* Beds — occupy about one half of the narrow strip between the shore of the Kalmarsund and the cliff of Västra Landborgen. A very small isolated outcrop of these beds is also found beneath the soil in the centre of the dome at Mossberga. The youngest strata of the unit consist of dark or almost black shale followed by greenish arenaceous shale. A greenish gray soft shale forms the uppermost part. The total thickness of the unit is stated to be 57 m in a boring 2.5 km S of Borgholm and from this place the thickness decreases both southward and northward (Westergård, 1946). The *Oelandicus* Beds were of great importance in the formation of the Mossberga dome.

The two facies of the overlying *Paradoxissimus* Beds were discussed in the review of the stratigraphic setting of southern Öland. According to Martinsson (1965) the thickness of the *Paradoxissimus* Beds does not exceed 15.1 m immediately E of Borgholm. A thickness of 12 m is known from a drilling 2.5 km S of Borgholm (internal report of the former Electrical Prospecting Company, now the Terratest Company, Stockholm). Towards the north the thickness decreases. Although the stage is still present at Horns Udde on the west coast it has not been found with certainty in the boring at Böda Hamn (Waern, 1952).

Martinsson (op. cit., p. 188) has found a series of very small anticlines with fairly steep limbs (dips reaching 38°) and with amplitudes of

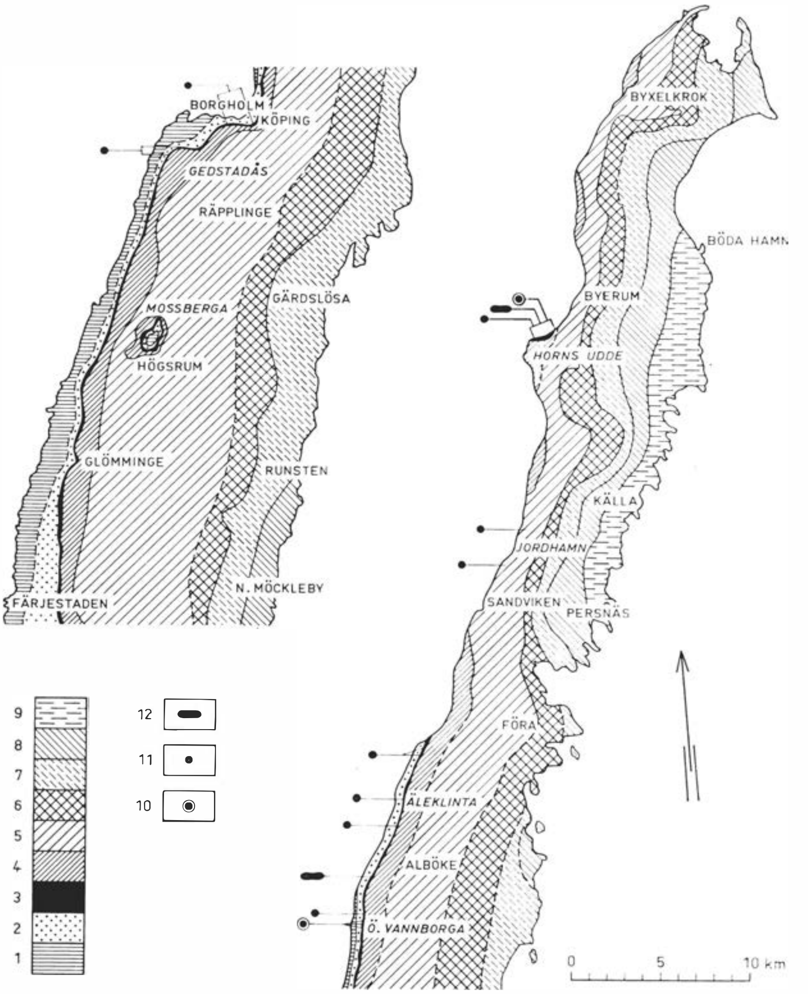


Fig. 15. Geological map central (left part) and northern Öland. The central part and the southernmost parts of northern Öland are compiled from Munthe 1902a, Munthe & Hedström 1904, Hedström & Wiman 1906 and notes from Jaanusson 1960a. North of Föra the map is compiled from Jaanusson 1960a, text fig. 2. Symbols:
1. *Eccaparadoxides oelandicus* Beds; 2. *Paradoxides paradoxissimus* Beds; 3. Upper Cambrian and Lower Ordovician alum shales; 4. *Ceratopyge-Lepidurus* Limestones; 5. *Vaginatum* Limestone; 6. Segerstad Limestone; 7. Folkeslunda, Seby and Skärlöv Limestones; 8. Källa and Persnäs Limestones on northern Öland, Furudal Limestone on central Öland; 9. Dalby Limestone; 10. Intermediate domes; 11. Small domes; 12. Small anticlines in Lower Ordovician limestones.

around 45 cm at Kolstad Villastad in the town of Borgholm. The fold structures were found in the lower part of the *Paradoxissimus* Beds consisting of *Paradoxissimus* Shale with alternating siltstone beds, cf. the drawing handed over by Dr. Martinsson, Fig. 16. The structures were exposed during different excavations but are unfortunately no longer available for examination. According to Martinsson (op. cit., p. 189) the folds are one of several indications that there have been tectonic movements along the bend of cliff line at Borgholm, and subsequent investigations of temporary exposures at the castle ruin of Borgholm exclude the possibility of glacial tectonics (Martinsson, personal communications). The main lithology of the lower parts of the *Paradoxissimus* Beds, i.e. alternating layers of silty and shaly beds, and the formation of boudinage

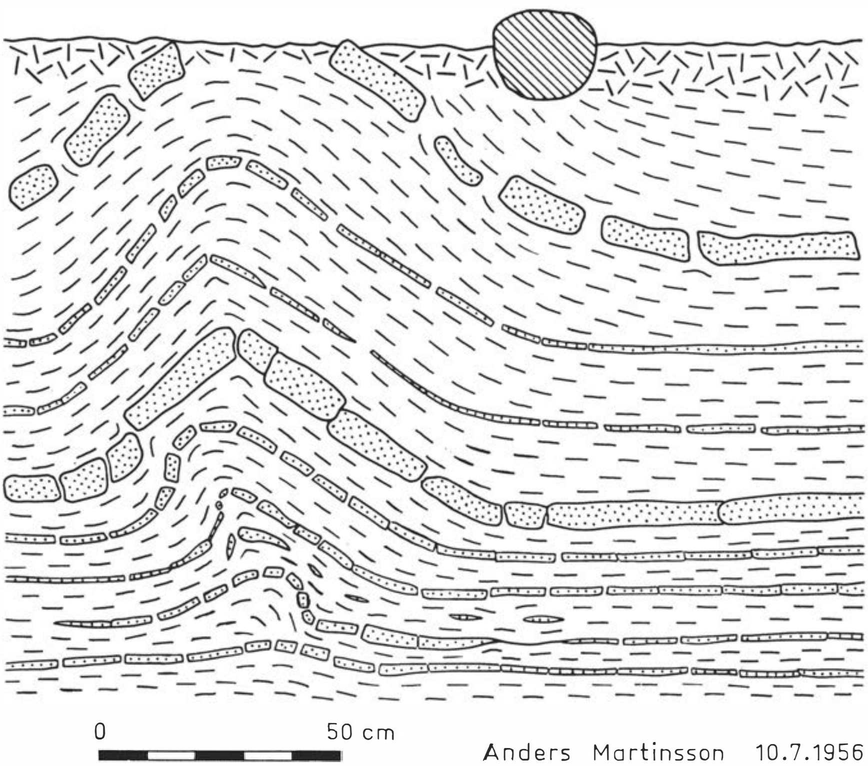


Fig. 16. Structure in the lower part of the *Paradoxissimus* Beds at Vikingavägen, Borgholm, engaging two of the three major siltstone beds in the local section. Axis trend 102°, obliquely intersected by temporary excavation in the street towards 54°. Two lower folds with similar trends were found 8 and about 20 m south of the figured structure. Topsoil and thinner siltstone layers omitted. The top layer of the shale grades into local moraines with scattered allochthonous boulders. Drawn from a field sketch and a colour slide.

structures in the competent siltstone beds of the folds, speak in favour of a gravitational origin. The almost complete lack of outcrops of the lower parts of the *Paradoxissimus* Beds on Öland has prevented a further investigation.

The isopachytes of the Upper Cambrian and Lower Ordovician shales maintain their main direction on central and northern Öland. Along the Västra Landborgen in the south and the cliffs of the west coast towards the north the thickness seldom exceeds 2 m. On the eastern part of central and northern Öland the thicknesses are nowhere known to exceed 10 m and 4.5 m respectively. The collective term alum shale can no longer be used due to a great difference in facies. Thus, the Upper Cambrian Olenid Shale mainly consists of a conglomeratic bank of stinkstone. In the area of Borgholm the thickness is around one metre, but this decreases towards the north and the unit seems to be altogether lacking in the northernmost parts. Thicknesses of the *Dictyonema* Shale are rarely known except at the sea cliffs along the western coast; here they seldom exceed a couple of decimetres. According to Westergård (1947) the zero thickness isopachyte runs in a NNE direction through a point 1 km E of the town of Borgholm. However, a more easterly direction north of Borgholm is likely since the shale is lacking in the drill core obtained at Böda Hamn (Waern, 1953). Known outcrops of *Dictyonema* Shale are of alum shale type.

The Ceratopyge Shale seems to be present all over Öland. In northern Öland the Shale exhibits both alum shale facies and glauconitic shale facies with laminae of bituminous shales. From sections through the Early Ordovician beds in northern

Öland the thickness is found to vary between 0.7 and 1.8 m (Tjernvik, 1956).

The very different facies and small thicknesses of the Upper Cambrian and Lower Ordovician shales almost exclude the possibility of these units to form diapiric structures like main ridges and large domes. Hence, along the well-exposed western part of northern Öland it is not surprising that these types of structures have not been found. The thicknesses of the shales along the eastern part of central Öland do not preclude the possible appearance of diapiric structures but this area is mostly concealed by Quaternary deposits.

Our knowledge of the Lower Ordovician limestones is solely restricted to the northern part of the island, e.g. Regnéll (1942), Bohlin (1955), Tjernvik (1956), Hadding (1958), and Lindström (1963). Like the situation on southern Öland, the lower Arenigian limestone succession is grouped into two stages, the Hunneberg Stage at the bottom and the Billingen Stage above. Both stages consist of layered limestones with shaly and marly intercalations. In the bottommost parts the shales can be said to dominate. The Billingen Stage is overlain by the *Limbata* Limestone and the *Lepidurus* Limestone. The investigations on northern Öland have confirmed the results obtained by Lindström (1963), namely, that small and intermediate domes and anticlines (sedimentary folds according to Lindström) do not appear in units stratigraphically higher than the *Lepidurus* Limestone. The different places where the structures are to be found are indicated on the geological map, Fig. 15.

The *Vaginatum* Limestone of Öland is known mainly from the work of Bohlin (1955). According to Swedish stratigraphy the *Vaginatum* Limestone comprises the beds from the *Expansus* Limestone up to and including the *Gigas* Limestone. From the boring at Böda Hamn the total thickness is known to be 8.4 m. In mapping the northernmost part of the island Bohlin found a zone of slight folding just north of Byxelkrok. The fold axis of this broad fold trends almost NW—SE and the main structure is apparent from the stratigraphical boundaries on Fig. 15.

The stratigraphy and lithology of the Viruan (Middle Ordovician) limestones are described by Jaanusson (1960a). According to him the Aserian and Lasnamägian topo-stratigraphic divisions, i.e. the divisions from the Segerstad Limestone up to and including the Folkeslunda Limestone, have the same lithological characteristics throughout Öland, cf. the description of these rocks given for southern Öland. Differences between northern and southern parts of the island appear in the lithological development of the Uhakuan beds, mainly in the middle and upper parts. Throughout Öland the bottom part of the Uhakuan Beds consist of calcilutite intercalated with finely nodular limestone and marl. In northern Öland the topo-stratigraphic division Källa limestone forms this part. The Källa Limestone is overlain by the Persnäs Limestone, a coarse calcarenite. The highest beds exposed on Öland belong to the Dalby Limestone of Kukrusean age. This sequence is lithologically fairly uniform, and consists of coarse calcarenites with thin intercalations of calcareous shale, mudstone or marl.

The boring at Böda Hamn, northern Öland, has yielded the following thicknesses of the Viruan limestones (Jaanusson, 1960a):

Kukruse Stage	Dalby Limestone	5.80 m+
Uhaku Stage	Persnäs Limestone	5.15 m
	Källa Limestone	2.20 m
Lasnamägi Stage	Folkeslunda Limestone	2.88 m
	Seby Limestone	0.20 m
	Skärlöv Limestone	2.04 m
Aseri Stage	Segerstad Limestone	5.13 m

The marly and shaly intercalations in the limestone pile of southern Öland were supposed to be the driving agency in the formation of the diapiric structures, e.g. the hump-back ridges and the intermediate domes. None of these structures have been found in beds stratigraphically higher than the *Lepidurus* Limestone on northern Öland. Although the bedrock of southern Öland is more exposed, and consequently the possibility of discovering structures is much better, there are lithological differences preventing the formation of these structures on northern Öland. A comparison of the lithology between the Aseri Stage and the Uhaku Stage from the borings at Böda Hamn on northernmost Öland and Gammalsby, 2.5 km E of Kvinsgröta on southern Öland (cf. Fig. 2) shows that the layers of calcareous shale, mudstone and marl so common in the boring of Gammalsby are almost absent in the boring of Böda Hamn, cf. Jaanusson (op.cit., text figs. 3, 4, 18). Hence, the source layers required for diapiric formation are missing. The Dalby Limestone sequence in the boring of Böda Hamn shows several intercalations of mudstone and marl that might have acted as source layers in the formation of diapiric structures, but unfortunately only small and scattered outcrops of the Dalby Limestone are accessible for observation, and here no structures have been found.

NOTES ON THE STRUCTURES OF CENTRAL ÖLAND

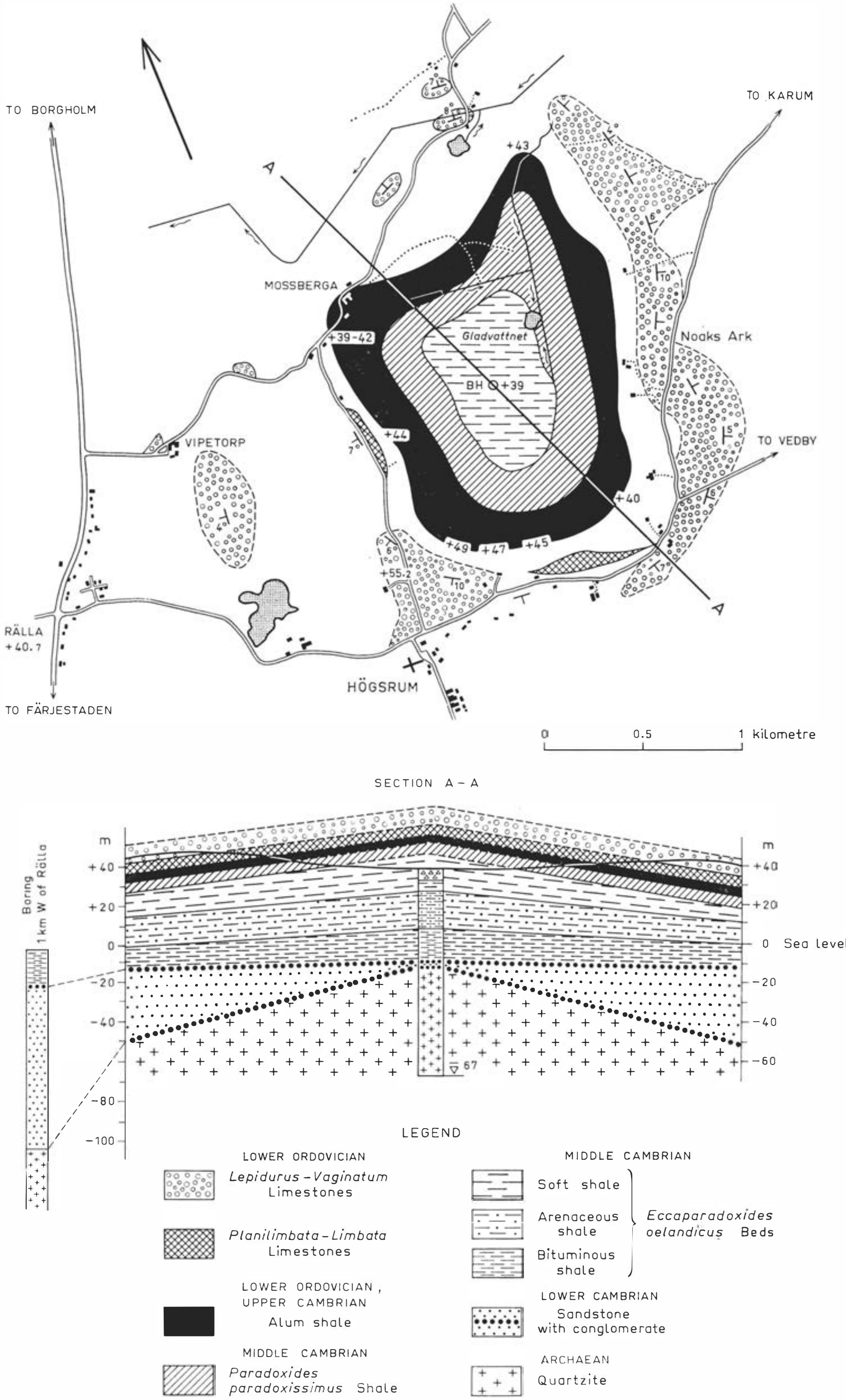
Although the bedrock of central Öland is mostly concealed by Quaternary deposits it appears in it a few interesting large structures. The Mossberga dome is situated about 12 km S of Borgholm and 3 km S of Borgholm the structure at Gestadås can be followed for more than 3 km in a NE—SW direction. Both structures were investigated in the early thirties by the former Electric Prospecting Co. under the guidance of Dr. O. Meier in a search for gas and oil in the Lower Cambrian of Öland. Some of the results were presented by Dr. Meier at a meeting of the Geologiska Föreningen, Stockholm, in March 1935.

The Mossberga dome

The main structure of the dome is apparent on the early geological maps of Öland, cf. Hedström and Wiman (1906). The dome is concentrically built up of outwardly dipping Lower Ordovician limestone in the outer part followed by successively older rocks towards the centre, Fig. 15. Topographically, the landscape rises towards the dome as far as the border of the *Vaginatum* Limestone, and in fact the highest point on Öland is situated just north of the church of Högsrum. The topmost part of the dome has been removed due to denudation and only scattered outcrops are available. Hence, the boundaries between the different stratigraphic units in the central part are mainly based on results of magnetic and electric resistivity measurements, placed at the author's disposal by Terratest Co., Stockholm.

In 1933 a boring was made in the centre of the dome. It reached 106.2 m beneath the surface which is situated 39 m above the sea level. The core was subjected to a close investigation by Westergård (1936) and the sequence he obtained is shown at the centre of section A-A, Fig. 17. The following description of the core is a summary of Westergård's results. From the bottom of the boring up to the 52 m level and 4 m beneath the top of the Archaean, the rock is a uniform true quartzite without any traces of clastic textures. The uppermost 4 m are weathered and the contact between the weathered zone and the overlying conglomerate is sharp. The same type of quartzite occurs in several places in the Kalmarsund area, e.g. the peninsula of Skäggenäs about 10 km W of Mossberga, N of the town of Kalmar, on the isle of Norra Skallarön, NE of Kalmar and on the island of Jungfrun. These scattered occurrences of quartzite in the Kalmarsund area formed small monadnocks on the sub-Cambrian surface when it was invaded by the Cambrian sea. These monadnocks, including the buried one at Mossberga, form the southernmost outliers of the Västervik quartzite of Lower Archaean age. The quartzite surface under the Mossberga dome dips outwards on all sides and from a water drilling 1 km W of Rälla, i.e. almost 4 km from the centre of the dome, the surface of the quartzite is encountered more than 95 m lower down.

The quartzite is overlain by the Lower Cambrian basal conglomerate, about 15 cm in thickness. Then follows a grey mainly fine-grained sandstone which grades upwards into a 60 cm thick conglomerate. The very small thickness of the Lower Cambrian, 1.7 m, is the most remarkable feature of the sequence at Mossberga. In Westergård's opinion the great thickness of the upper conglomerate and the abundance of quartzite pieces found here, indicate that the quartzite formation


TO BORGHOLM
TO KARUM
MOSSBERGA
Gladvattnet
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VIPETORP
TO VEDBY
RÄLLA
+40.7
HÖGSRUM
TO FÄRJESTADEN
A
0 0.5 1 kilometre
SECTION A – A
m
+40
+20
0
-20
-40
-60
-80
-100
0 Sea level
Boring 1 km W of Rälla
67
LEGEND
LOWER ORDOVICIAN
Lepidurus – Vaginatum Limestones
Planilimbata – Limbata Limestones
LOWER ORDOVICIAN, UPPER CAMBRIAN
Alum shale
MIDDLE CAMBRIAN
Paradoxides paradoxissimus Shale
MIDDLE CAMBRIAN
Soft shale
Arenaceous shale
Bituminous shale
Eccaparadoxides oelandicus Beds
LOWER CAMBRIAN
Sandstone with conglomerate
ARCHAEAN
Quartzite

may reach to a higher level somewhere in the vicinity of the boring where it has never been covered with Lower Cambrian sediments. The upper surface of the Lower Cambrian is supposed to have a gentle dip outwards from around the area of the bore hole, mainly due to differential compaction of the sandstone pile. The validity of this assumption is supported by the level of the Lower Cambrian surface observed in the boring 1 km W of Rälla and in other borings S of Rälla, Fig. 17. Although the sandstone is supposed to have behaved passively (i.e. not participated) during the growth of the Mossberga dome under the effect of gravity, there may have been a slight bulging in this layer below the trunk of the dome. From model experiments of dome formation we know that if the substratum below a layer of buoyant material is not rigid it tends to participate in the flow and form a bulge below the dome. Further, the rise of the substratum will increase as the degree of doming increases, cf. Ramberg (1967, p. 112). Hence, the slight domal rise obtained combined with the moderate thickness and rather strong competence of the sandstone under the Mossberga dome suggest that a limited bulging of the sandstone has occurred.

The buoyant layer in the formation of the Mossberga dome is made up of the different shales belonging to the Middle Cambrian *Eccaparadoxides oelandicus* Beds. Westergård's description of the *Oelandicus* Beds from the boring at Mossberga (1936, p. 13) may be quoted *in extenso:*

"The lower portion of the *Oelandicus* beds, from the conglomerate up to the 32 m level, consists of a dark-grey or almost black somewhat bituminous clay-shale, often with an abundance of extremely thin seams of grey sandstone. Between 32 and 29 m, alternate dark-grey shale and lighter greenish-grey fairly arenaceous shale. Above the latter level the greenish-grey arenaceous shale is the only rock present, and it continues up to 11.1 m, where it is covered by a 0.3 m thick stratum of greenish-grey argillaceous and calcareous sandstone. At 10.8 m and thence upwards there appears another greenish-grey shale which is less arenaceous and is softer than the lower one. The highest strata are somewhat displaced by the continental ice, and the covering moraine consists exclusively of fragments of the same kind of shale in a clayey matrix. It may be assumed that the eroded part of the Oelandicus shale originally had a thickness of at least 5 and probably, about 10 m."

The lithological partition of the *Oelandicus* Beds mentioned above can also be observed in small outcrops along the shore of Kalmarsund from the area south of Rälla to Borgholm. Studies of the different types of shales exposed along the shore indicate clearly their ability to have acted as a ductile buoyant substratum in a domal rise.

The thicknesses of the overlying Middle Cambrian *Paradoxides paradoxissimus* Beds are almost unknown in the Mossberga area. From the lateral extent which is mainly based on geophysical results the thickness is estimated to be around 10 m. Whereas the *Paradoxissimus* Shale and Siltstone and the overlying alum shales, about 5 m thick, have actively cooperated with the *Oelandicus* Beds as source layers for the dome formation can not be assessed on present evidence. However, the Upper Cambrian and Lower Ordovician alum shale has served as a low-strength ductile layer of "smearing" at the contact between the overlying limestones and the underlying shales and in this manner has facilitated the dome generation.

On the surface the Lower Ordovician limestones outcrop around the shales with stratigraphically younger limestones away from the centre. Only scattered outcrops of the *Planilimbata* and *Limbata* Limestone are found whereas the younger *Vaginatum* Limestone extensively crops out in the southern and eastern parts of the dome, Fig. 17. The outward dips vary from a few up to ten degrees and the region of the anomalous dips departing from the prevalent ESE direction indicates an outer dome diameter of around 4 km.

In as much as domes with diameters on the scale of kilometres are lacking for the rest of Öland there must have been certain specific factors

Fig. 17. Geologic map and section of the Mossberga dome. The boundaries between the different stratigraphic units in the centre of the dome are based on geophysical measurements. The boring in the centre shows the stratigraphy and main lithology according to Westergård (1936). Data for the boring 1 km W of Rälla are compiled from the well record office of the Geological Survey of Sweden.

predominant only at Mossberga which lead to the dome formation here. Hence, the metastable situation of lighter ductile shales overlain by heavier limestones is not enough. If both source layer and overburden are uniform in thickness, composition and density, and if further the layering is parallel, then diapiric structures should theoretically not start growing from the metastable source layer. From theory, experiments, and nature we know that irregularities either in geometry or property are necessary to generate doming.

The dome-generating irregularity at Mossberga is the quartzite monadnock which in the vicinity of the boring is supposed to project up in the bituminous shale of the lowermost *Oelandicus* Beds. The dynamics of the dome generation is explained by the change of thickness and weight of the source layer along the contact with the quartzite. This creates a pressure difference in the source layer around the monadnock that could initiate horizontal flow. A complete treatment of the dynamics and history of the doming, together with descriptions of centrifuged models, will be presented in a forthcoming paper.

Due to high resistance of quartzite to denudation we have reason to assume that there may exist other monadnocks which may have initiated doming on Öland and in the area of Kalmarsund. On the island of Öland we have so far no clear indication of further doming. The only possible locality is situated just north of Byxelkrok on northernmost Öland, where a slight folding of the limestone units was first reported by Bohlin (1955). Lack of deep well holes and limited outcrops render a determination of the origin difficult, however. Of the monadnocks of quartzite now present in the area of Kalmarsund, the one forming the Skäggenäs Peninsula on the mainland at almost the same latitude as Mossberga, may possibly have initiated a dome. Abundantly fossiliferous Lower Cambrian sandstone beds alternating with gray and reddish shale from a locality in the western part of the Skäggenäs Peninsula have been described by Åhman and Martinsson (1965). From well-sinkings in the area of Rälla it may be seen that the top of the Lower Cambrian is situated 20 m below sea level. With an estimated eastward dip of 5 m per 1000 m and a distance from Rälla to the Skäggenäs Peninsula of 4.5 km, we may reckon that the upper surface of the Lower Cambrian was about 2.5 m above sea level at the the peninsula. Hence, a situation with deposition of *Oelandicus* Beds directly upon the Archaean quartzite seems likely and by analogy with the situation at Mossberga, this implies that a former dome may have existed on top of the quartzite at Skäggenäs Peninsula.

The Gestadås structure

The strong eastward bend of the cliff at Borgholm is the most striking evidence for tectonic movements in this area. This bend is, to some extent, also reflected by the slight bend of the stratigraphic boundaries, cf. Fig. 15. Strong deviations in attitude of the limestone units from their normally flat-lying position can be found around the ruined castle of Borgholm situated at the easternmost part of the bend in the cliff. 150 m south of the castle a zone of strongly tilted Lower Ordovician limestones can be followed for more than 100 m in a WNW—ESE direction. Zones of weakness in the bedrock are evident from the strongly eroded valleys situated on both sides of the main road 1.5 km south of the ruined castle.

Further towards the south, about 2.5 km from the castle, we find the middle part of the Gestadås structure. This structure was first investigated in the early thirties in connection with oil and gas prospecting. Results of eight borings and several excavations along the northern and central part of the ridge have been kindly placed at my disposal by Dr. O. Meier. According to the borings and excavations the northern part of the ridge is a sub-horizontal upright fold with gently dipping limbs (10°—20°). A fault running along the crestal line is found with about 10 m downthrow of the eastern block.

The borings in the eastern limb on the central part of the ridge indicate a repetition of the Upper Cambrian and Lower Ordovician units. The most probable explanation for this repetition is the

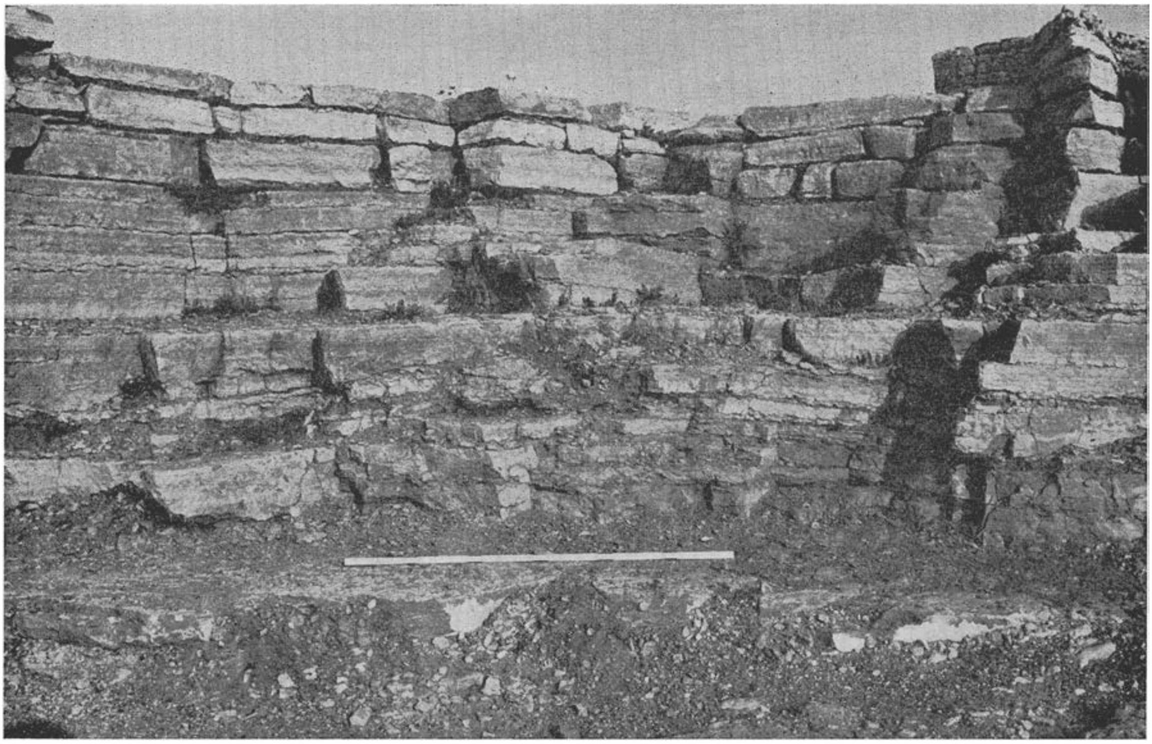

Fig. 18. Intermediate dome of the type multi-layer domical diapir from the old quarry 1.5 km NNW of Ö. Vannborga. The scale is 1 m. Notice the diameter diminishing upwards, the cessation of growth followed by horizontal deposition and finally a slight bulging in a later stage. The uppermost three layers of limestone form a stone wall.

presence of a low angle, westward dipping fault striking almost parallel with the eastern part of the ridge. Whether the faulting along the ridge was preceeded by any anticlinal diapirism can be determined only after further borings.

DIAPIRIC STRUCTURES ON NORTHERN ÖLAND

So far the only large structure described from northern Öland has been the slight fold trending NE—SW just north of Byxelkrok (Bohlin, 1955). The origin of this fold is uncertain. A diapiric process with the alum shales acting as a source layer is out of question due to the small thickness of the shales. Diapirism involving the *Oelandicus* Beds is plausible, but there is still the lack of a mechanism of initiation.

A geological interpretation of aerial photographs by N. B. Svensson (personal communications) has shown several rather wide depressions or valleys crossing the northern part of the island in a NW—SE direction. No strong deviations in dip of the limestone surface, generally gentle dipping to the ESE have been found in the vicinity of these valleys, and an origin due to ice movement seems most likely on the evidence now available. However, it must be emphasized that observations along the valleys can only be made at scattered places and only three glacial striae preserved to indicate the direction of ice movement on northern Öland according to Königsson (1968). According to N. B. Svensson (op. cit.) the valleys may reflect the topographic subsurface of the Precambrian basement.

The most interesting structures on this part of the island are the intermediate and small domes

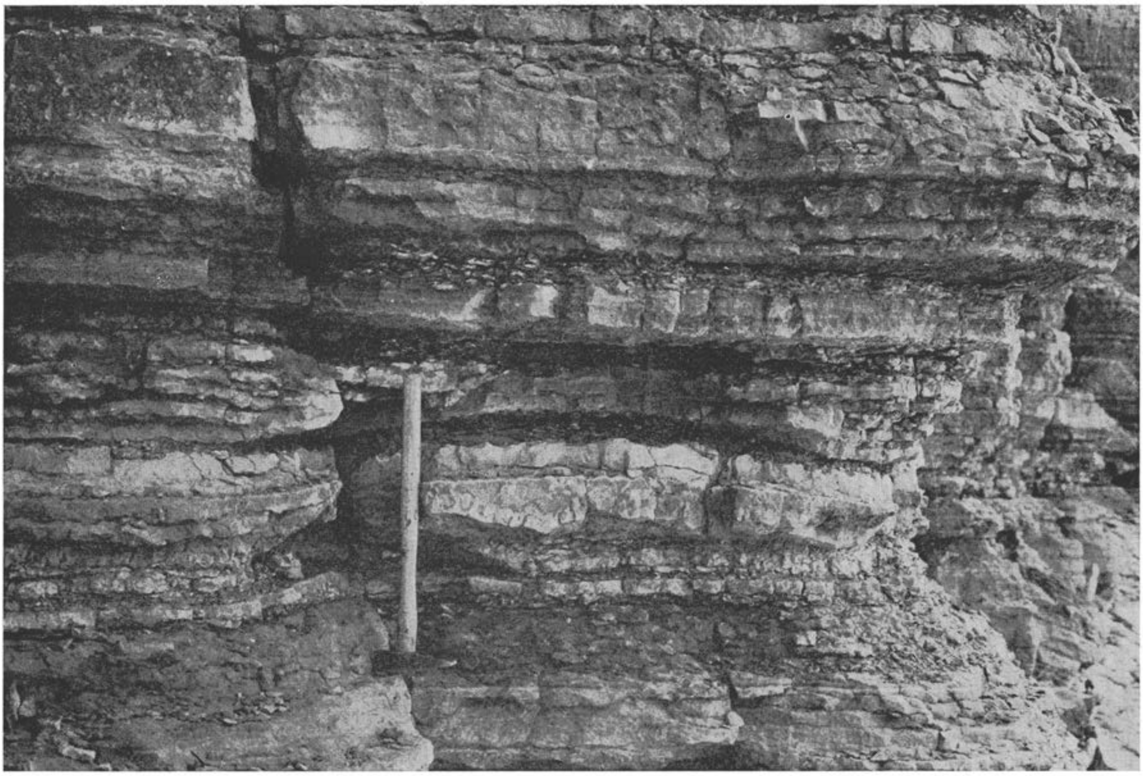

Fig. 19. Small domes from the Billingen Stage, Lower Ordovician, Horns Udde. The strongly fissured buoyant layer of dark marl may be seen in the centre of the upper dome. A bent limestone layer may be seen in the lower dome. Notice the slight bulging of the limestone bed underneath the buoyant layer and the almost horizontal limestone layers above the domes.

and ridges which can be found in great numbers and with varying appearance in the Lower Ordovician along the western coast.

Intermediate domes

Intermediate domes with diameters ranging from a couple to around ten metres have been found at three localities along the cliff of the western coast, cf. the map, Fig. 15. According to the proposed classification of domes (p. 59) the one situated at Horns Udde is of the type *single-layer domical diapir,* cf. Lindström 1963, plate 1, section 0—2.5 m. Here the source layer consists of marl and the overburden of one layer of limestone about 10 cm thick. The other intermediate domes are of the type *multi-layer domical diapir.* Very representative for this type of structure is the southernmost of the domes in the old quarry 1.5 km NNW of Ö. Vannborga, Fig. 18. The bottom part of the dome, where the scale of the figure is situated, has a diameter of 4 m. The diameter gradually diminishes upwards. On both sides of the major dome other small gentle domes appear. The intercalations of marl in the limestone unit were the source layers in the formation of this type of dome, cf. p. 61. On the left-hand side of the dome we notice how younger beds of limestone have been deposited horizontally and against the flank of the dome. Further, the slight outward dip of the discontinuity surfaces on the top of dome indicates slight movement at a later stage.

Small domes

In an article on "Sedimentary folds and the development of limestones" Lindström (1963) described folded limestone beds of Lower Ordovician age from two main areas of southern Sweden. On the island of Öland fold structures were described from four different localities, Horns Udde, Borgholm and Köpings Klint, Perstorp and Ottenby, the two latter localities being on southern Öland. The best exposures of the folds were found along the sea-cliff at Horns Udde on northern Öland. According to Lindström (op. cit.) "A typical fold in the orthoceratite limestone comprises only one limestone bed that it usually a few cm thick. Under the folded bed one will find a bed of marl, which in some cases fills the core of the fold. The limestone beds that follow under the marl are usually not affected by the same folding. The marls have been a zone of 'décollement'. The folding can thus regarded as a model of the classic conception of Jura folds", op. cit. p. 252. Lindström further mentions that the wavelength of a fold is commonly about 0.5 m or little less, the normal variation being 0.2—1.5 m, and the amplitude is generally about 15 cm. At first sight, a folded limestone layer looks like an ordinary fold, but a close examination shows that the structures mostly terminate inwards at right angles to the exposed surface; in reality most of the structures are domes. This can be seen in many places where the intensely fissured layer of marl has been washed out, or after the loose layer has been scraped out, Fig. 19; cf. also Lindström (op. cit., figs. 7 and 13).

Along the almost NE—SW striking cliff at Horns Udde, Lindström measured 82 fold axes and 100 joints and constructed a rosette diagram for the trend of the fold axes and a sum polygon for the joints whose horizontal extent was also included (op. cit., fig. 8). With the reservation that each measurement is somewhat uncertain, Lindström shows that there is a strong maximum for folds trending in an almost NW—SE direction, i.e. perpendicular to the exposed surface. It can be proved that there exist some, although very few, elongate domes or ridges where a measurement of fold axis could confidently be made. However, most of the folded layers have a dome structure. Another direction of the cliff line or exposed surface, thus, would result in a corresponding change of the direction of the maximum in the trend rosette. Lindström also mentions that because the strongest maximum of fold trends is almost normal to the most frequent joints it renders the fold maximum suspect.

The small domes must have started to grow plastically, simultaneously with or very soon after the deposition of the next limestone bed or beds. That the doming was very early is readily seen from the almost horizontal bedding appearing in the very next marl or limestone bed above the bent one, cf. Fig. 19 and the drawing at Horns Udde by Lindström (op. cit., plate I, pp. 277—292). Also notice the tendency for the bent limestone bed to thin at the top of the domes. This deformation took place very early when the marl layer was probably still plastic or slightly hardened.

In several domes at Horns Udde the source layer has pierced through the bent limestone bed. Further, it seems as if the uppermost parts of the source layer material have been dissolved or most probably washed out and later replaced with calcareous sediments. In places where the source layer has pierced through, we often find a layer of marl situated along one or both rims of the dome immediately above the bent limestone bed. This possibly indicates a redistribution of the source layer material that was originally situated underneath the bent limestone bed to a present position above it.

The author fully agrees with several of the conclusions made by Lindström, for example that the structures stood upright on the sea-floor during the deposition of younger sediments, and that the bent limestone layers mineralogically and texturally differ from the normally developed limestone beds etc. However, unlike Lindström I have found that the limestone bed that is situated underneath the source layer in most of the domes has but a fairly regular bulge with an amplitude of a couple of cm. This is a structural pattern which among other features lends support to a theory of gravitational emplacement of the domes and at the same time proves that the underlying bed of limestone was plastic but nevertheless stiffer than the light buoyant layer during the gravitational doming.

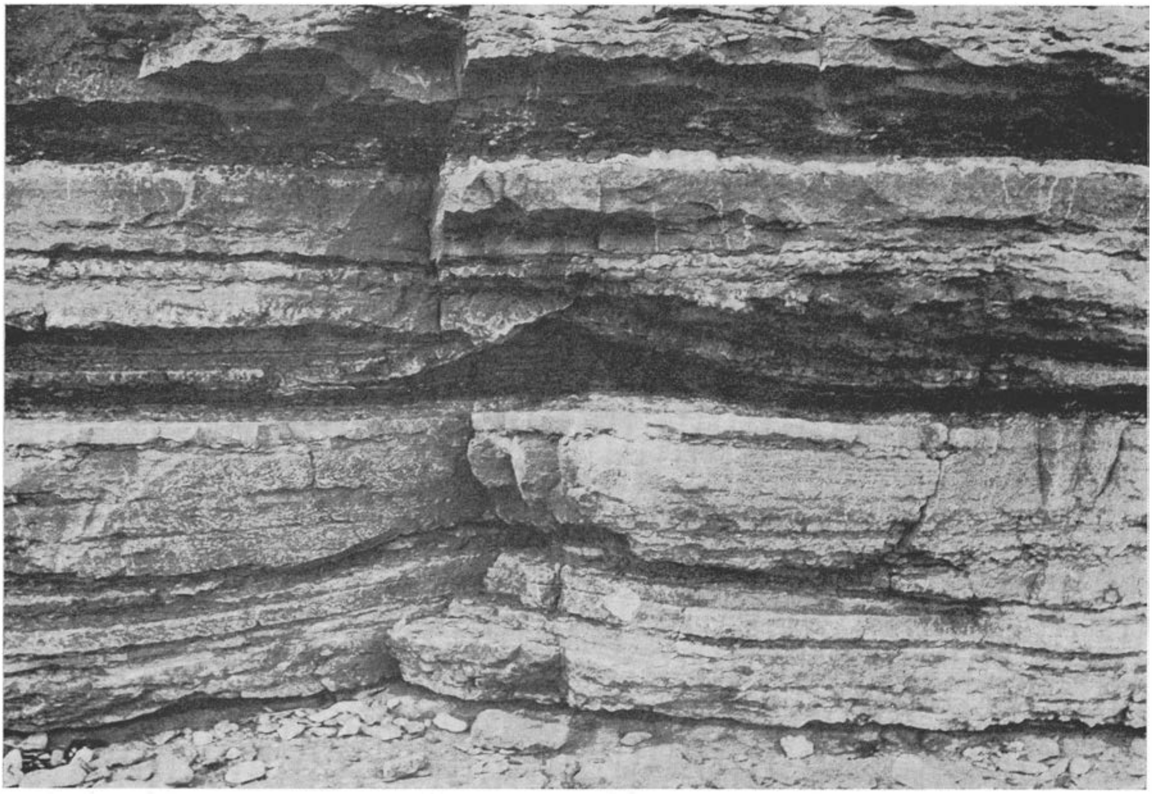

Fig. 20. Section of a single-layer anticlinal diapir from the Billingen Stage, Lower Ordovician, Horns Udde. The thickness of the dark, buoyant layer of marl is 14 cm in the centre of the anticline.

According to Lindström some tectonic movement was the triggering that caused the practically consolidated limestone bed to glide down an almost imperceptible eastward slope and "The folds rose where the bed was pushed together by the movement", op. cit., p. 256. He further mentions that the bigger folds at Horns Udde indicate that some tectonic buckling took place in the Early Arenigian. "Indeed the small-scale folding itself may have been released by earthquake shocks", (op. cit., p. 271).

Later Gidon and Lindström (1965) renounced the belief in submarine gliding as a factor that could have played an important role in the formation of the structures, except for those in the lower part of the section at Horns Udde. Instead they involve a mechanism of buckling due to humidity changes and dilatation of the limestone surface, inspired by the behaviour of a Carrara marble plaque on the wall of the "Hotel de Ville" at Chambéry in France. Although the buckling of the riveted marble plate is in itself an interesting phenomenon, this non-tectonic mechanism is unlikely to have been responsible for the structures found in the Early Arenigian limestone beds on the island of Öland. This type of buckling demands a more or less rigid constraint. The surface of contact between the limestone bed and the underlying plastic substratum of marl could hardly offer that constraint. Although the marl layers sometimes die out and the superincumbent layer comes into contact with the underlying bed of limestone we still have a discontinuity surface which could hardly withstand the buckling forces. At the Horns Udde section we further notice that the distances between these points of contact sometimes reach tens of metres; were a constraint imposed we would expect larger wavelengths of the folds than we in fact can see. Finally, this mechanism of buckling can hardly explain the slight bulging of the limestone bed underneath the marl layer observed in most of the domes.

The basic mechanism of formation of the small domes is a gravitational instability of an inverted density stratification, i.e. marl layers of low density are overlain by limestone layers of higher density. The initiation of the small domes, both of single- and multi-layer domical type, is due to irregularities in the stratification. Hence, at Horns Udde most small domes are found in the lower portion of the Billingen Stage where the stratification of marl, nodular marly limestone and limestons beds is most irregular, cf. Lindström (1963, plate 1).

Small ridges

Some of the structures in the cliff of Horns Udde do not show any sign of termination perpendicular to the exposed surface. These are small ridges or anticlines and in accordance with the proposed classification they are single- or multi-layer anticlinal diapirs depending on whether one or several layers of marl form the source layer. At Horns Udde the type first mentioned dominates, Fig. 20. The mechanism of initiation and growth is the same as that for the small domes, i.e. initiation due to irregularities in the stratification and growth due to an inverted density stratification.

A single-layer anticlinal diapir with a curved crest line was found in the bottom of a small quarry situated 2000 m WSW of the church of Alböke, Fig. 21. The length of the exposed part of the ridge is about 6 m. Towards the NW the structure closes with a well-rounded nose. From here it gradually bends until it reaches a NNE direction at the vertical wall of the quarry. The crest rises almost 40 cm above the flat-lying surroundings, and the width or wavelength varies between 2 and 3 m. The core of the ridge consists of a gray marl and marly limestone and is covered by a 7 cm thick layer of glauconite and goethite bearing red limestone. The dip of the limestone surface increases slightly towards the crest; dips of almost 30° can be found locally. A marked thinning of the covering limestone bed can be found at the crest where the bed is still intact, cf. Fig. 22. At the vertical wall of the quarry we find horizontal younger beds of limestone situated about 20 cm above the top of the source layer of marl.

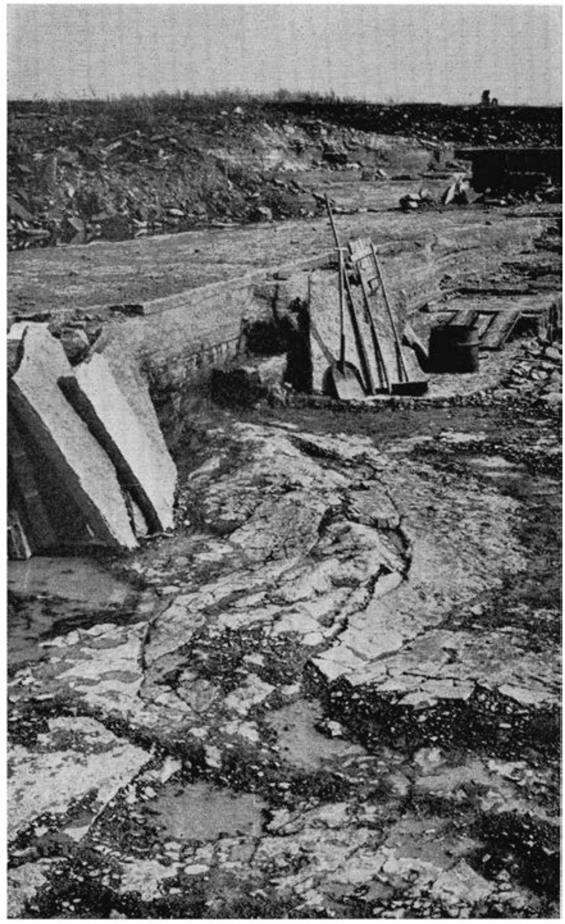

Fig. 21. A curved single-layer anticlinal diapir from a small quarry 2 km WSW of the church of Alböke. The length of the exposed part is about 6 m. The NW end of the structure terminates with a well-rounded nose.

The limited disturbance above the ridge and the marked thinning along the crest of the covering limestone bed both indicate a very early origin for the structure, i.e. when the marl and limestone were still plastic sediments or muds. Although the establishment or initiation of some of the joints can be related to the growth of the ridge most of them are of a later date. Hence, joints

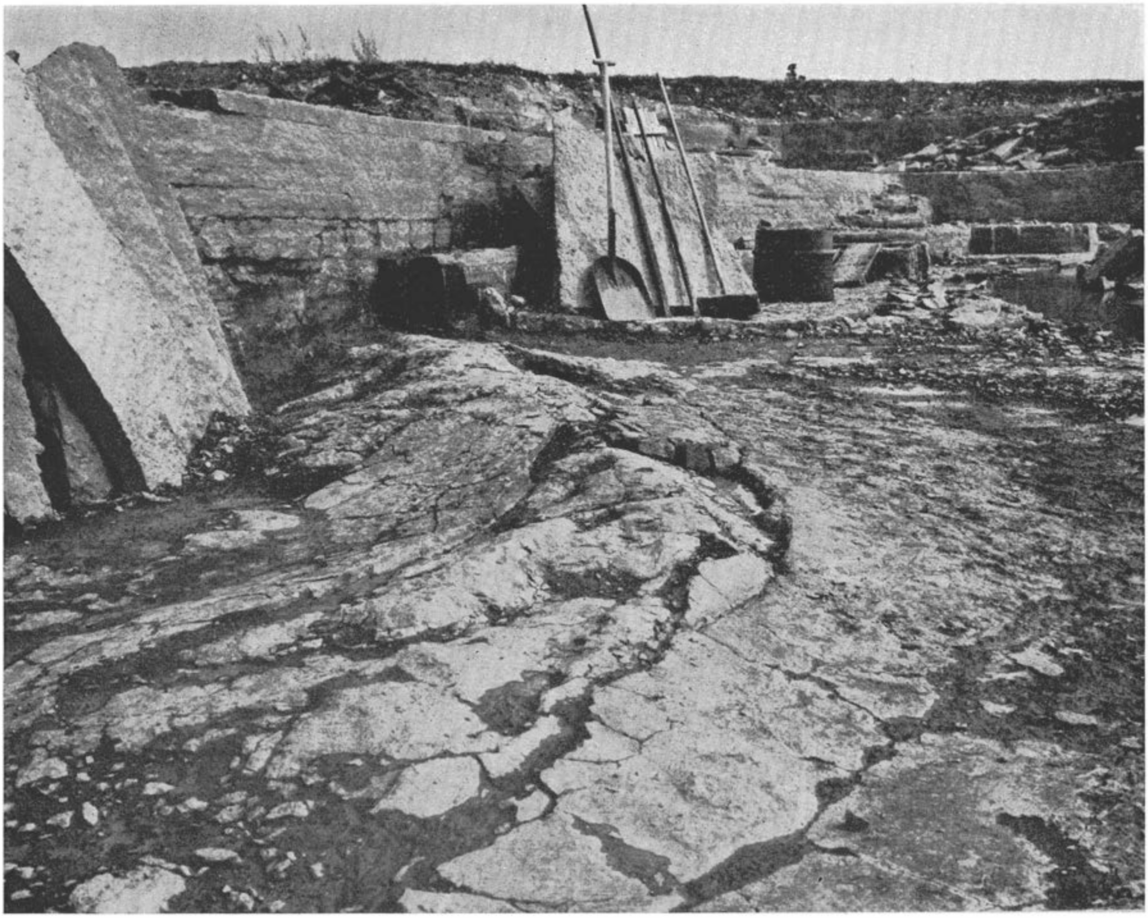

Fig. 22. The same ridge as shown in Fig. 20. The core of the ridge consists of grey marl and marly limestone. A layer of glauconite and goethite bearing limestone forms the cover. Stretching of the limestone layer has caused a thinning of the layer along the crest. Notice the horizontal discontinuity-surface in the stone wall just above the crest of the ridge.

and sets of joints can be followed both in the source layer and on the flanks of the covering limestone bed. From chlorite-coated slicken-sided joint surfaces within the marls we have clear evidence of moderate later movements.

The question why we sometimes get domes and sometimes ridges by the same basic mechanism, is related to irregularities in the source layer or overburden, i.e. irregularities in thickness, composition, density and rheology. We also know that the outer boundary (i.e. lateral extent) of the source layer is a strong controlling factor for the growth and final geometry of diapiric structures. In nature we often find diapiric anticlines trending parallel with the major axis of evaporitic or sedimentary basins, e.g. the Paradox Basin at the border of Utah and Colorado in the United States (Elston and Shoemaker 1963), and the northern part of the Zechstein basin in Germany described by Trusheim (1960). The same observation has been made in model experiments, cf. Ramberg (1967, figs. 12, 13). From the experiments we know further that a source layer with a curved outer boundary favours the development of curved anticlinal rises. Hence, the geometry of the small ridge described above may well be controlled by a curved outer limit and related thickness variations of the source layer of marl.

ACKNOWLEDGEMENTS

In preparing this paper I have had the great advantage of discussing various questions with Professor Hans Ramberg, Director of the Institute of Mineralogy and Geology, University of Uppsala. The field investigation was to a great extent inspired by the work of Professor L.-K. Königsson and Dr. K. Palmqvist on the Great Alvar of Öland. I am greatly indebted to Professor Königsson for the loan of aerial photographs and for generous help during the field work.

Stratigraphical determinations of some of the structures were kindly made by Dr. K. Larsson. Discussions with Drs. B. Dahlman, H. Möller and J. Pousette, all of the Geological Survey of Sweden, have been of importance. The former chief geologist of the Terratest Company, Dr. O. Meier kindly placed at my disposal diaries and maps of the Mossberga dome and the Målartorp ridge.

I am very grateful to Dr. P. Hudleston, Professor A. Martinsson, Professor B. Collini and Professor H. Ramberg for reading and criticizing the manuscript. Dr. Hudleston also revised the English of the manuscript.

The maps and drawings were made by Mrs. A. Kaljusaar and Mr. H. Holmström of the Institute of Mineralogy and Geology, University of Uppsala. Financial support was given by The Swedish Board for Technical Development, grants 512910 and 5398.

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Printed by Ekenäs Tryckeri Ab, Ekenäs, Finland Aug. 31, 1971.