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Influence of Lithology and Neogene Uplift on Seismic Velocities in Denmark: Implications for Depth Conversion of Maps[1]

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ABSTRACT

A new velocity parameter, the velocity anomaly (*dV*), is defined as the deviation between a velocity measurement and a linear velocity-depth function for a formation.

The variation of the velocity anomaly of the lowermost Jurassic claystones in northern Denmark is interpreted as the combined effect of three types of uplift movements, and the magnitude of these are estimated. It is concluded that the pre-Quaternary surface in northern Denmark is an erosional surface, created by up to 1000 m of Neogene uplift and subsequent erosion. The Upper Cretaceous Danian Chalk Group and the Tertiary sequence must thus have been deposited far beyond their present extensions. Consequences of Neogene uplift for the source rock potential of the area are discussed.

In the Danish Central Trough, North Sea, the velocity anomaly of the Lower Cretaceous is shown to reflect lithological variations related to differential subsidence and to characterize the geological regions better than the interval velocity.

Since the primary influence of depth on velocity is removed, the velocity anomaly is an expression of physical factors such as lithology, overpressure, and gas content as well as uplift and erosion. Contoured maps based on well data of the velocity anomaly of a formation may thus be used to estimate the velocity anomaly at arbitrary locations. The velocity anomaly map, the seismic time maps, and the linear velocity parameters for each layer constitute the input for velocity-anomaly depth conversion. This depth-conversion method results in depth maps with perfect well ties and geologically reasonable depth estimates away from wells.

INTRODUCTION

In well-explored regions of the world, such as the North Sea, data from a dense seismic grid and numerous wells are available. These data have been used for numerous regional time maps and geological models, whereas depth maps tend to be restricted to more limited areas, perhaps due to the lack of a general method for precise depth conversion. It is the objective of this paper to present a method for analyzing regional velocity data and for producing reliable depth maps from seismic maps, when velocity information is available from several wells.

Velocity is the key in the conversion from seismic traveltime to depth. But velocity is itself dependent on depth. The porosity of most sedimentary rocks is reduced as the depth of burial increases. The increased compaction in turn results in a higher degree of grain contact and, hence, in an increased acoustic velocity in the rock.

If the dependency of velocity on depth is approximated by a linear function, the velocity for a formation becomes expressed by depth of burial and the velocity parameters of the formation: the velocity gradient and the so-called surface velocity or, rather, velocity at the surface. This simple velocity model will, of course, differ from velocity data in general. The velocity anomaly (defined below) is the deviation between a velocity measurement and a regional linear-velocity function for a formation.

The velocity anomaly can be related to the concept of apparent uplift (Bulat and Stoker, 1987). Apparent uplift is defined as the difference between the estimated depth that the velocity-depth function predicts for an observed velocity and the actual depth. If a layer has been buried at greater depths than the actual level, the sedimentary rock will often keep the degree of compaction and seismic velocity corresponding to the highest pressure level in its past; an "excess" velocity may thus be an indication

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of uplift. Variations in other parameters may give rise to deviations from a regional velocity function; Bulat and Stoker (1987) mention variations in pore fluid, matrix velocities, shale content, geothermal gradient, diagenetic processes, and localized tectonic stress.

As the factors affecting the velocity anomaly are related to geological processes, it will be shown that it is useful to draw contoured maps of the velocity anomaly for a formation. The basic concept in the interpretation of these maps is that the velocity anomaly is an expression of lithology and the post-depositional history of the sedimentary rock. Changes in velocity anomaly may thus be interpreted in terms of a sand/shale or shale/carbonate ratio, uplift, overpressure, erosion, or other factors. This concept makes it meaningful to estimate the velocity anomaly at arbitrary locations by interpolation between anomalies determined in wells. Maps of the velocity anomaly can thus be used in the geological interpretation of an area and as tools for producing reliable depth maps, which will be shown through examples from the Danish area.

THE CONCEPT OF VELOCITY ANOMALY

The increase of velocity with depth due to compaction is the first thing to be considered when investigating the regional variations of velocity for most sedimentary rocks. The simpler way to take this depth dependency into account is by assuming linearity of velocity, V, with depth, z:

$$V(z) = V_0 + kz$$

where V_0 is the surface velocity (measured in m/s) and k the velocity gradient (measured in s^{-1}). On the basis of this linear approximation, we can develop an expression for the interval velocity, V_i, for a given layer (see Appendix):

$$V_i = 2\frac{\Delta z}{\Delta T} = \frac{2}{k\Delta T}(V_0 + kz_t)(e^{k\Delta T/2} - 1) \qquad (1)$$

where z_t is the depth to the top of the layer, Δz is the thickness, and ΔT the two-way time thickness of the layer. The assumption of linearity of velocity, V, thus leads to this nonlinear expression for the interval velocity, V_i, measured for the layer as a whole. The velocity parameters, V_0 and k, for a layer can be determined by nonlinear regression of well measurements of Δz, ΔT, and z_t. The determination of V_0 and k may be supported by involving log measurements of V and z in the analysis. This may be necessary when investigating a layer with little variation in depth of burial and thickness, in which case the interval velocity will be almost constant in all wells. Note that what is important in depth conversion is the increase of velocity with depth for equal lithologies. The velocity-depth trend of a sonic log is thus not always indicative of the velocity gradient for varying depth of burial for the whole layer.

If we want to convert traveltimes to depths, we have to solve equation 1 for Δz. But the analytical expression for V_i given by the equation represents a model of the earth that does not take the lateral variations of the seismic velocity into account. In general, a model calculation of Δz will differ from well data. The correction needed to calibrate the linear velocity model to well data will thus be defined as the change in the surface velocity, V_0, for a given layer, that will lead to the measured thickness of that layer in a given well. This calibration constant, referred to as the velocity anomaly, dV, is defined in each well for each layer with given linear velocity parameters V_0 and k, for measured values of thickness, $\Delta z'$ two-way time thickness, $\Delta T'$, and depth to top of the layer, z'_t:

$$\Delta z' = \frac{1}{k}(V_0 + dV + kz'_t)(e^{k\Delta T'/2} - 1) \qquad (2)$$

The velocity anomaly has the dimension m/sec and ranges typically between -500 and 500 m/sec. As will be shown later in this paper, lateral variations of dV are the combined expression of lateral variations in V_0 and k, caused by differences in both lithology and in the post-depositional history of the sediments. Presence of gas will of course lead to low dV values.

If the velocity anomaly is believed to be a result of uplift, z_{up}, then z'_t in equation 2 should be replaced by $z'_t + z_{up}$ and dV set equal to zero to find the depth of burial that would result in the observed velocity anomaly. After division, we get

$$dV = kz_{up}$$

The velocity anomaly would thus be half the size of the apparent uplift for many clastic sediments having $k = 0.5s^{-1}$. As dV is independent of layer thickness, it is a better measure of apparent uplift than differences in interval velocities.

One important condition to be met before uplift can be estimated is that the layer under consideration is homogenous, i.e., that no lateral lithological variations affect the interval velocity. In order to investigate whether the calculated uplift is only apparent or it is actually real, the uplift theory should be tested to see if it is physically and geologically possible. This can be done by analyzing the geographical variation and the timing of the apparent uplift and comparing it with possible geological

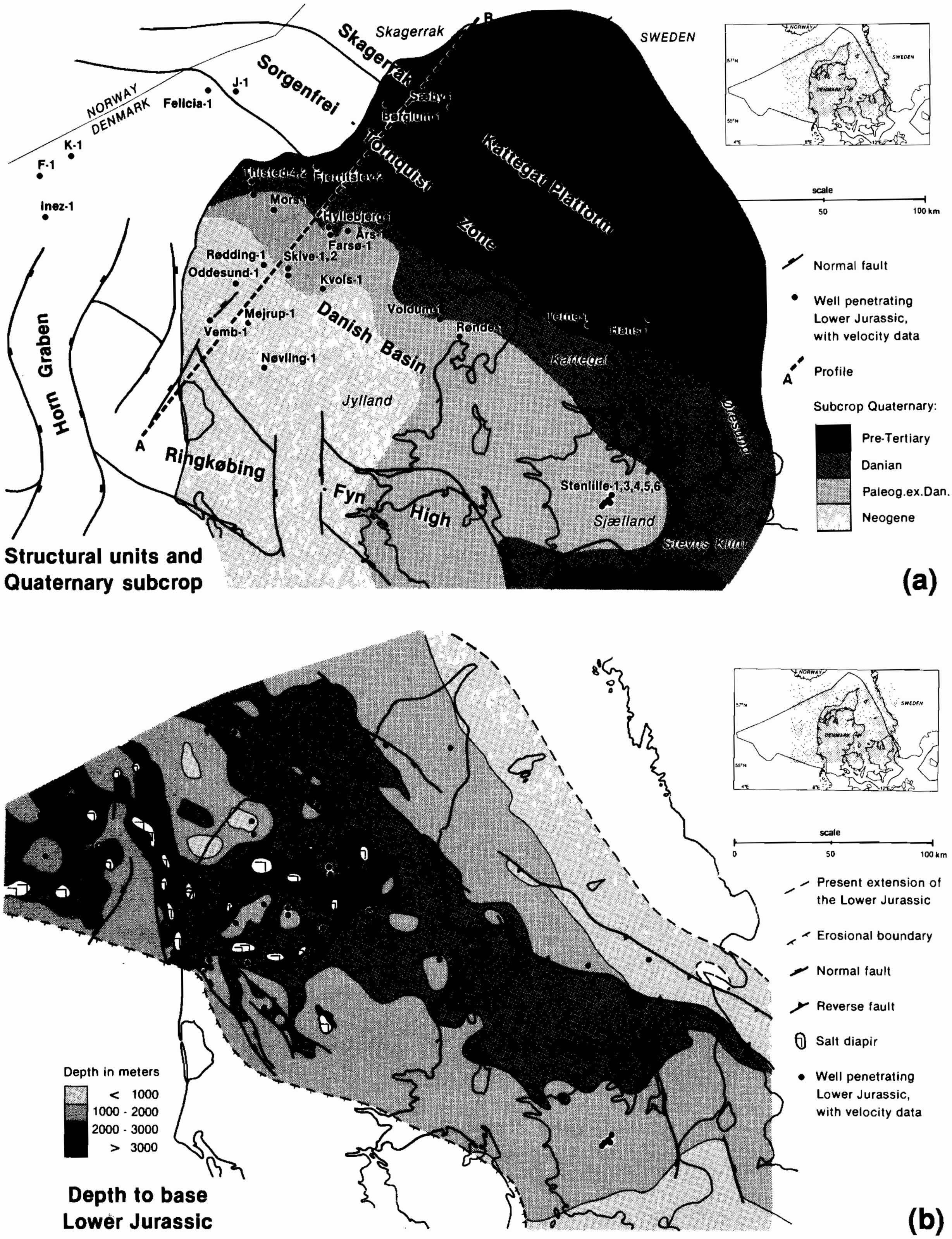
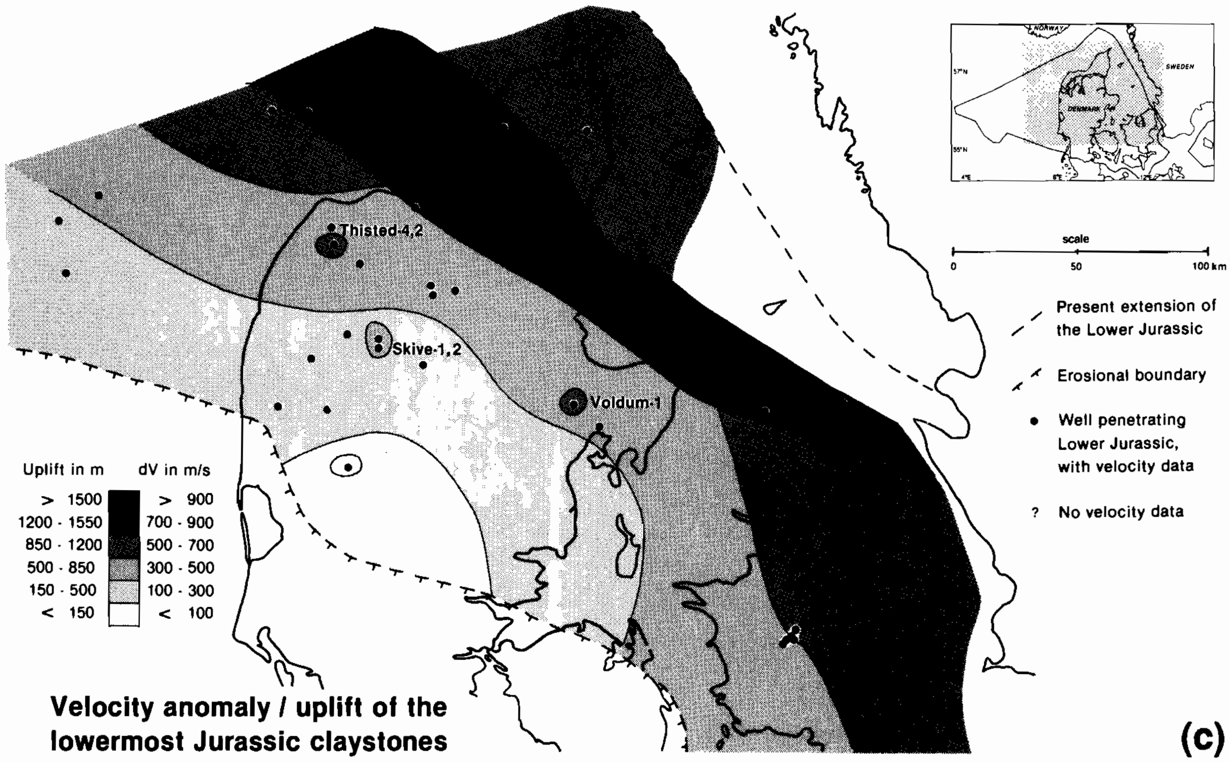
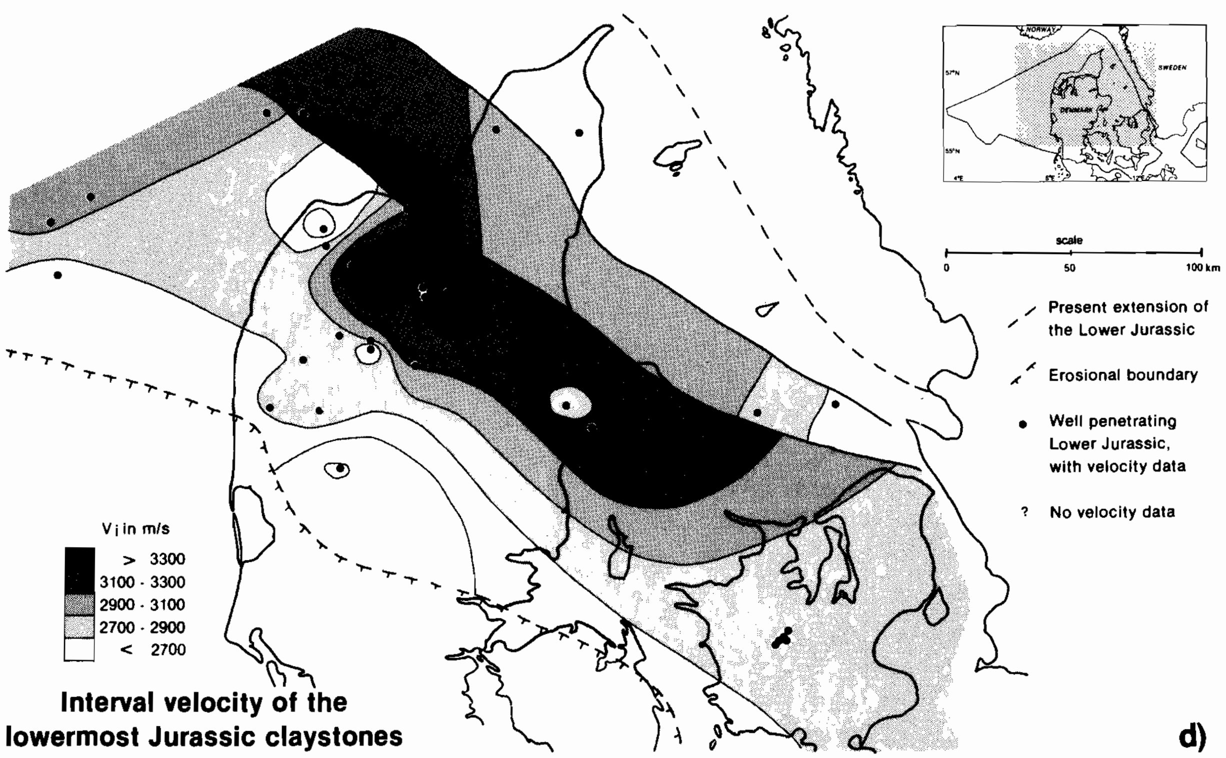


Figure 1—The Lower Jurassic in northern Denmark. (a) Structural units and Quaternary subcrop. (b) Depth to base Lower Jurassic, modified after Bjelm et al. (1977), Bertelsen (1978), Michelsen (1981), Pedersen (1985), and Japsen





and Langtofte (1991b). (c) Velocity anomaly and the corresponding uplift of the lower Jurassic F-Ia and F-Ib members, relative to $V = 1535 + 0.58z$. (d) Interval velocity of F-Ia and F-Ib members.

mechanisms. Furthermore, there should be agreement between uplift calculated for different formations believed to have undergone the same tectonic evolution.

Of crucial importance, if the absolute uplift is to be determined, is that the uplift be measured relative to a velocity-depth function corresponding to a normal compaction curve. This means that the function should be determined in an area where the formation under investigation has not undergone uplift nor is affected by overpressure.

DATA

Linear velocity parameters have been determined for several geological and seismic units within two geological provinces in Denmark. One analysis was based on 59 wells outside the Central Trough (Japsen, 1988), mainly in the Danish basin (see Figure 1), the other on 74 wells in the Danish Central Trough (central North Sea, see Figure 6). A further subdivision of the area outside the Central Trough would make the statistical basis too narrow. Each well has been subdivided into lithostratigraphic units (Nielsen and Japsen, 1991) and units corresponding to seismic boundaries. For each group, formation, member, or seismic unit, the depths and two-way traveltimes to top and base of the unit have been used as input to the analysis. The independent parameters V_0 and k in equation 2 have been estimated by the Marquardt nonlinear regression method (SAS Institute Inc., 1985) for each unit, given the values of z'_t, $\Delta z'$, and ΔT for each unit in all wells.

The data were given different weights in the analysis. Data points were excluded from the regression analysis if the well is drilled on a salt diapir, if a unit is thin ($\Delta z < 20$ m or if $\Delta T < 10$ ms), or if the data point was considered to be too far from the general velocity-depth trend. Data points representing only the upper part of a layer were also included in the analysis. If the unit, however, was pierced by a neighboring well or if the thickness of the drilled part of the unit was thinner than the measured maximum thickness for nearby wells, the data point was excluded. In order to approach a geographically even distribution of the data points in the regression analysis, averages of interval velocities were calculated for the same unit for wells within quadrants of 30' × 15' if the depth to the midpoints of the units was within a distance of 200 m. Finally the data points were weighted in the regression analysis by their thickness for $\Delta z < 100$ m and by 100 for $\Delta z \geq 100$ m.

No attempt, however, was made to establish velocity-depth functions corresponding to normal compaction curves by discriminating data points believed to have undergone uplift, other than for wells clearly situated on top of salt diapirs. The velocity functions thus established represent regional averages of subsidence and uplift.

Given the linear velocity parameters for each unit, equation 2 was applied to calculate the velocity anomalies for every unit in every well. Finally, contoured maps of the velocity anomaly were prepared for each of the units.

ESTIMATION OF LATE CRETACEOUS TO TERTIARY UPLIFT IN NORTHERN DENMARK

Structural Outline

The northern part of Denmark has had a complex structural development due to large-scale tectonic movements and salt tectonism. In the following, the amount of uplift of the area is quantified based on velocity measurements.

Following the EUGENO-S Working Group (1988), the structural units of northern Denmark are the following (Figure 1a): (1) The Skagerrak-Kattegat platform, situated southwest of the Baltic shield, with a partial cover of relatively thin and undisturbed Mesozoic sediments erosionally truncated by a Quaternary cover. (2) The Sorgenfrei-Tornquist zone, characterized by Late Cretaceous–Paleogene inversion tectonics. (3) The Danish basin, the Danish part of the Norwegian-Danish basin, where the basinal development began with evaporitic precipitation during the Late Permian. During the Mesozoic and Cenozoic, sedimentation was mainly clastic, but with Late Cretaceous and Danian carbonates. The structural development of the basin was characterized by repeated changes in subsidence pattern and by halokinetic processes. (4) The Ringkøbing-Fyn high, a basement high with a reduced Mesozoic-Cenozoic cover.

Northern Denmark has been affected by three types of uplift movements since the Early Cretaceous. Two of the uplifts are well known, but the Neogene uplift, suggested by the present study, has attracted less attention. These uplifts are (1) Late Cretaceous–Paleogene inversion movements in the Sorgenfrei-Tornquist zone, (2) halokinetic movements in the Danish basin, and (3) Neogene uplift affecting the Skagerrak-Kattegat platform, the Sorgenfrei-Tornquist zone, the Danish basin, and parts of the Ringkøbing-Fyn high.

The regional extent of the Neogene uplift and the accompanying erosion makes it difficult to recognize, and only few speculations about the uplift are found in the literature. Larsen and Dinesen (1959) assumed that considerable parts of Fennoscandia (including not only basement, but also sedimentary formations) were eroded in the Neogene. On the basis of various geological evidence, Spjeldnæs (1975) argued that a major late Tertiary uplift movement (starting in the Oligocene) had affected Denmark. According to Spjeldnæs (1975), this movement possibly reflected the uplift of the Atlantic margin of the Fennoscandian shield. Following Nielsen et al. (1986), Neogene uplift could be located in the Fennoscandian border zone or Tornquist line (i.e., the Skagerrak-Kattegat platform and the Sorgenfrei-Tornquist zone) as a result of a late Alpine tectonic episode.

Jensen and Schmidt (in press) argue for the theory of Neogene uplift of the eastern North Sea basin. The onset of Neogene erosion is observed on seismic sections northeast of the Central Trough (central North Sea, Figure 6) and an increasing amount of erosion is noted towards the Norwegian coast. Jensen and Schmidt (in press) believe this pre-Quaternary erosional unconformity to be related to the uplift of the Norwegian mountains. On the basis of vitrinite reflectance trends and shale compaction methods, they estimated the amount of Neogene uplift south of Norway; in northwestern Denmark, they estimated the uplift to be from 1000 to more than 1500 m, increasing toward the north. Jensen and Schmidt (in press), however, did not distinguish between the Late Cretaceous–Paleogene and the Neogene uplift movements.

Michelsen and Nielsen (in press) estimated the uplift in northern Denmark on the bias of sonic readings for the two lower members of the Lower Jurassic Fjerritslev Formation for seven wells. They noted an increasing amount of post–Early Jurassic uplift moving from the central North Sea towards the Skagerrak-Kattegat platform, and estimate values of Late Cenozoic uplift ranging from 600 to 1100 m. Well data are compared with a constructed velocity-depth trend assumed to represent undisturbed subsidence; the trend is drawn as a straight line on semilog paper between the sonic readings for the O-1 well in the Danish Central Trough at a depth of 3000 m and a surface velocity value of 200 µs/ft (1524 m/s) (Magara, 1976). Michelsen and Nielsen (in press) noted that a precondition for this use of the O-1 well is that no significant overpressure is present. This, however, is not the case. Mud weight data indicate that the excess head at the base of the Jurassic in the O-1 well is a 1700-m water column (2500 psi). This level is in good agreement with the overpressure generally found in this depth in the Central Trough (Thorne and Watts, 1989).

Using the method described below, the apparent uplift for the Lower Jurassic in the O-1 well is calculated to be −420 m. The negative value may be explained by undercompaction caused by the rapid Neogene to Quaternary subsidence in the central North Sea found by Nielsen et al. (1986). Values of the total uplift found by Michelsen and Nielsen (in press) are thus 200–400 m above the values estimated in this paper.

Interval Velocity of the Lowermost Jurassic Claystones

To investigate the uplift of the region, the interval velocities of the two lower members of the Lower Jurassic Fjerritslev Formation were studied. The F-Ia Member consists of claystone, interbedded with siltstone or limestone, whereas the overlaying F-Ib Member is dominated by claystone (Michelsen, 1989). The two members are uniformly developed over the entire area (Michelsen and Nielsen, in press). The velocity anomaly for the lowermost Jurassic claystones (calculated as the deviations from a velocity-depth trend for normal compaction) can thus be used to estimate uplift.

The interval velocity was investigated over the two members as a whole to assure a substantial thickness over which the velocity could be determined. The F-Ia and F-Ib members are encountered in numerous wells, located within the above mentioned structural units apart from the Ringkøbing-Fyn high; the average thickness is 200 m, ranging from 12 to 411 m. Of these wells, 30 penetrate the Lower Jurassic and have measurements of the interval velocity (see Table 1).

The interval velocities found for the F-1a and F-1b members show a clear tendency to increase with depth (Figure 2), but the data points from the Sorgenfrei-Tornquist zone indicate uplift relative to the general trend given by line (a). This is in good agreement with the inversion movement of the Sorgenfrei-Tornquist zone during the Late Cretaceous–Paleogene. Compaction line (a) is determined in the standardized manner described previously: data points considered to be too far from the general velocity-depth trend and data from wells drilled over salt diapirs were omitted from the regression analysis. But the trend given by line (a) will of course be biased by a broad-scale Neogene uplift that would have caused even the presumably subsiding parts of the Danish basin to be uplifted.

The Reference Compaction Line

To estimate the bias of the data, the apparent uplift was determined relative to line (b) in Figure 2 ($V = 1535 + 0.58z$), established by Scherbaum (1982) for undisturbed subsidence for the Lower Jurassic of northwest Germany, south of the Ringkøbing-Fyn high. The lithofacies of these Lower Jurassic sediments is not indicated by Scherbaum (1982), but according to Bertelsen (1978) northwest Germany and northern Denmark were part of one major depositional province during the Late Triassic and Early Jurassic. A striking similarity is found in facies and faunal distributions on both sides of the Ringkøbing-Fyn high. Bertelsen (1978) argued that the present extensional limit of the Lower Jurassic towards the

Table 1. Well Data for the Two Lower Members of the Lower Jurassic Fjerritslev Formation (F-Ia and F-Ib Members) in Denmark Outside the Central Trough*

Well	Company	z_{up}** (m)	dV** (m/s)	V_i† (m/s)	z_t (m)	Δz (m)	ΔT (m/s)	Comments††
Børglum-1	Dapco	1273	738	3058	1323	26	17	NCSL
F-1	DUC	500	290	2950	1839	202	137	Gentle salt pillow
Farsø-1	DONG	481	279	3296	2395	323	196	
Felicia-1	Statoil	1018	590	2922	1244	262	179	
Fjerritslev-1	Dapco	1408	816	2867	867	43	30	NCSL, TD, pronounced salt pillow
Fjerritslev-2	Dapco	1531	888	3556	1849	208	117	NCSL
Hans-1	DUC	1733	1005	2660	141	133	100	
Hyllebjerg-1	DUC	575	334	3266	2265	289	177	
Inez-1	DUC	335	194	2647	1532	101	76	Salt pillow
J-1	DUC	1493	866	3296	1393	303	184	Gentle salt pillow
K-1	DUC	625	362	2906	1536	411	283	
Kvols-1	DUC	448	260	3123	2177	228	146	Gentle salt pillow
Mejrup-1	Phillips	327	190	2879	1808	368	256	
Mors-1	DUC	785	456	3510	2492	258	147	
Nøvling-1	DUC	-431	-250	2236	1503	275	246	NCSL, abnormal pressure?
Oddesund-1	DUC	229	133	2698	1661	232	172	
Rødding-1	DUC	425	247	2833	1712	204	144	
Rønde-1	DUC	513	297	3264	2367	204	125	
Skive-1	DUC	595	345	3026	1860	233	154	Down-flank of salt pillow
Skive-2	BP	682	396	2639	1174	95	72	Salt pillow
Stenlille-1	DONG	940 848§	545 492§	2869 2771§	1257	208	145	Gentle salt pillow
Stenlille-3	DONG	1007 839§	584 487§	2903 2746§	1247	209	144	TD, gentle salt pillow
Stenlille-4	DONG	1165 839§	676 487§	3000 2746§	1248	228	152	Gentle salt pillow
Stenlille-5	DONG	1143 848§	663 492§	3003 2771§	1282	213	142	Gentle salt pillow
Stenlille-6	DONG	1188 914§	689 530§	3046 2822§	1304	227	149	Gentle salt pillow
Sæby-1	Britoil	997	578	2671	911	102	76	
Terne-1	Amoco	1405	815	2870	863	66	46	NCSL
Thisted-2	DONG	895	519	2676	1028	91	68	Salt pillow
Thisted-4	Amoco	662	384	2314	622	118	102	Salt pillow
Vemb-1	Dapco	309	179	2708	1649	130	96	NCSL
Voldum-1	DUC	942	546	3023	1527	195	129	Salt pillow
Års-1	DONG	558	324	3617	2905	255	141	

*From Nielsen and Japsen (1991). In the Stenlille wells (Figure 1a) are pronounced intervals of siltstone or sandstone reported for the F-Ia Member (Pedersen, 1985), resulting in higher interval velocities of the F-Ia and F-1b members. To attempt a comparison with the relatively clean lower Jurassic claystone in the rest of the area, only the velocities for a shale section near the top of the F-1b Member were used for the Stenlille wells. The Fjerritslev-1 well was not included in the analysis as it is drilled on a pronounced salt structure. The interval velocities of the F-Ia and F-Ib members for the Nøvling-1 well (2236 m/s) and Thisted-4 well (2314 m/s) are below the level where the assumption of linearity of velocity with depth is valid. The Nøvling-1 well plots below line (b) in Figure 2, whereas the Thisted-4 plots above. Assuming an increased velocity-depth gradient for velocities below 2500 m/s would make the velocity anomaly (and the apparent uplift) for both wells increase.

**Apparent uplift, z_{up}, and velocity anomaly, dV, calculated relative $V = 1535 + 0.58z$ (Scherbaum, 1982).

†The accuracy of the velocity measurements is reduced for thin sections and for sections where no calibrated sonic log is available (cf. Børglum-1 and Terne-1 wells).

††TD = well does not penetrate the Jurassic, NCSL = well has no calibrated sonic log, only check-shot survey and uncalibrated sonic log.

§Log readings for pure shale.

Ringkøbing-Fyn high is an erosional boundary caused by the middle Kimmerian uplift phase by the end of the Early Jurassic. A normal compaction velocity-depth trend established for the Lower Jurassic in northwest Germany thus provides an optimum reference for velocity measurements on the Lower Jurassic shales found north of the Ringkøbing-Fyn high.

Scherbaum (1982) used data from several hundred wells to establish linear velocity-depth func-

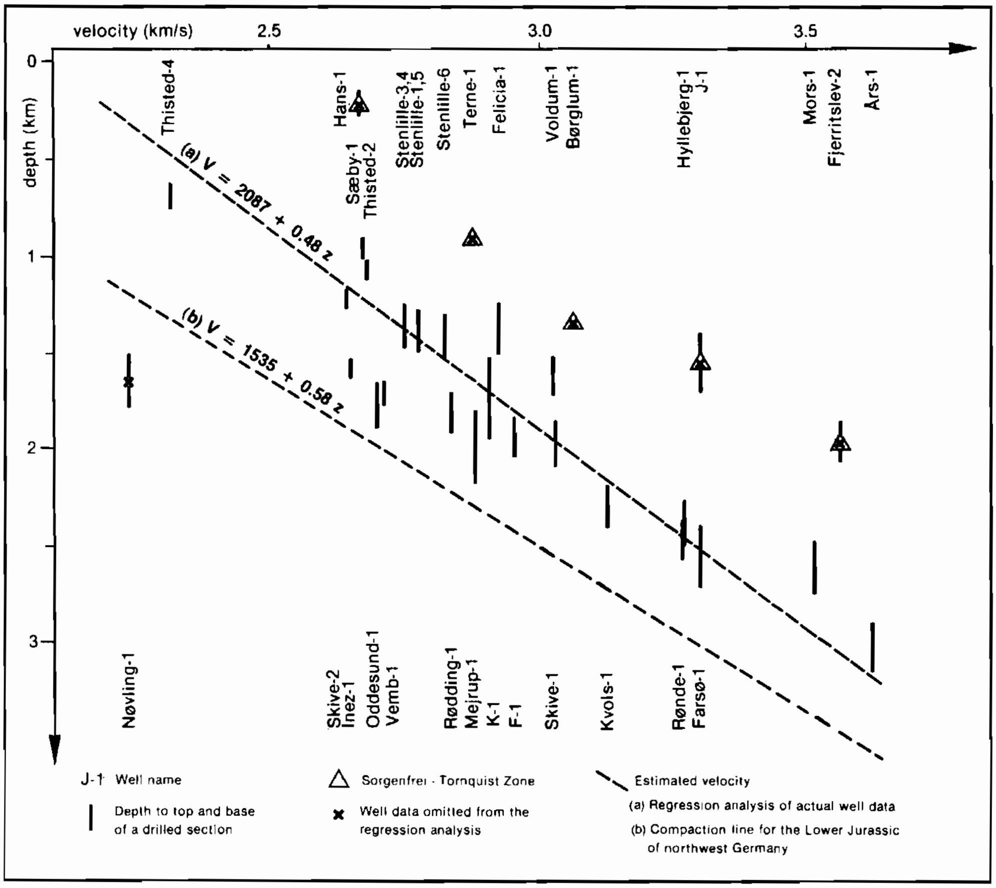


Figure 2—Interval velocities of the lower Jurassic F-Ia and F-Ib members in northern Denmark as measured in 30 wells penetrating the Lower Jurassic.

tions for several geological units. By choosing the smallest recorded velocities at each depth for each unit as reference for the determination of the linear functions, the data points were thought to have the highest probability of not being influenced by uplift movements. The actual number of wells available for the determination of line (b) in Figure 2 is not given, but Scherbaum (1982) stated that there is a high degree of uncertainty in the determination of the parameters for the velocity-depth trend for the Lower Jurassic. Scherbaum noted that the velocity-depth trend can be treated as linear at greater depths (i.e., below 1000 m, corresponding to velocities above 2500 m/s; at shallow depths velocity has an increased depth gradient). In the present study, all but two measurements of interval velocity are above 2500 m/s, and the assumption of linearity is thus justified (see Figure 2; Table 1).

Yet another compaction trend for Lower and Middle Jurassic shales in northwest Germany (Pompeckj swell), believed to be at their deepest depth of burial, is shown by Jankowsky (1962; 1986, personal communication). Within the depth range from 750 to 3500 m, this trend can be approximated by the linear function $V = 1800 + 0.5z$ with a maximum error of 5%. This linear function is in good agreement with that given by Scherbaum (1982). At a depth of 3000 m, the estimate of V is 25 m/s above line (b) in Figure 2; at a depth of 1500 m, the difference is 145 m/s.

The Neogene Uplift

The lateral variation of the interval velocity for the F-Ia and F-Ib members reflects the present-day depth to base Lower Jurassic (Figure 1b, d): high velocities are found in the deep parts of the Danish basin, whereas low velocities are found in the shallow parts of the basin, on top of structural highs over salt pillows, on the Skagerrak-Kattegat platform, and in the eastern part of the Sorgenfrei-Tornquist zone.

If the depth dependency of the interval velocity is removed by calculating the velocity anomaly relative to velocity function (b) in Figure 2, a different picture emerges (Figure 1c). The velocity anomaly increases to the north in the Danish basin and lines up with the level found on the Skagerrak-Kattegat platform (the Sæby-1 well). In between, the Sorgenfrei-Tornquist zone forms a distinct unit of high velocity anomalies, with the higher values found toward the east. The velocity anomalies of both preand post–Early Jurassic formations are likewise found to increase to the north and east (Japsen, 1988). This fact substantiates the theory that the variation of the velocity anomaly of the Lower Jurassic is caused by uplift rather than by lithology. Due to the lateral homogeneity of the mapped unit, the calculated apparent uplift may be interpreted as actual uplift, and the main pattern may be explained by two tectonic events: Late Cretaceous–Paleogene inversion of the Sorgenfrei-Tornquist zone and broad-scale uplift of the entire area in the Neogene.

The increasing amount of uplift and erosion to the north and east is in agreement with the ages of the Quaternary subcrop sediments found in these directions (Figure 1a). Only the pre-Quaternary unconformity has an areal extent similar to the uplift pattern shown in Figure 1c; all other post–Early Jurassic unconformities do not affect the entire area (EUGENO-S Working Group, 1988, where the pre-Quaternary unconformity, however, is not mentioned). The middle Kimmerian unconformity at the transition from the Lower to Middle Jurassic is observed over most of the area, but with increasing intensity towards the Ringkøbing-Fyn high, in contrast to the uplift seen along profile AB (Figure 1a) (Ziegler, 1990). Furthermore, in the Danish basin, only uplift that occurred after the Early Cretaceous affected the measured uplift. Prior to the deposition of the thick Late Cretaceous and Danian Chalk Group (cf. Nielsen and Japsen, 1991), the depth of burial of the Lower Jurassic was small, and maximum compaction of these sediments had thus not taken place.

Several observations (see Spjeldnæs, 1975) indicate that the pre-Quaternary erosional surface was created by Neogene uplift, or possibly by uplift that began in the Oligocene. In the Danish Central Trough, grain size is found to increase abruptly at the transition from the Paleogene to the Neogene, and the amount of kaolinite increases, indicating a supply of terrigenous weathering products presumably from the Baltic area and/or from the land areas of Scotland (Nielsen, 1979). In southern Jylland (Figure 1a), Paleogene sediments were eroded during the Late Oligocene to Miocene (Rasmussen and Larsen, 1989). Furthermore, a general westward withdrawal of the coastline over Denmark is observed during the Tertiary (EUGENO-S Working Group, 1988).

The uplift movement seems to have ceased before the Quaternary. The Quaternary thickens to the northeast along profile AB (Figure 3c), thus indicating ongoing subsidence (the Quaternary of northern Jylland is primarily marine). This conclusion is supported by the observation made by Fredericia and Knudsen (1990) that subsidence and tectonic movement may have been active during the Late Quaternary in northernmost Jylland.

The Holocene uplift due to isostatic rebound after the glaciations, registered over the entire Fennoscandia (see Mörner, 1991), is thus unrelated to the Neogene uplift. A possible cause for the overcompaction of the Lower Jurassic claystones could be the load of the Quaternary ice shields, but basin modeling studies of the effect of ice loads on compaction, show that the duration of the glaciations is much too small to have any effect (T. Bidstrup, 1991, personal communication).

It must be concluded that the pre-Quaternary surface in northern and eastern Denmark is an erosional surface created by Neogene uplift and subsequent erosion.

The linear increase of uplift to the north along profile AB is remarkable and suggests a rather rigid Neogene uplift movement of the Danish basin, the Sorgenfrei-Tornquist zone, and the Skagerrak-Kattegat platform as a whole (Figure 3a). Extrapolation along the profile yields zero uplift over the Ringkøbing-Fyn high west of Jylland, and 1 km uplift just north of the Sæby-1 well. This corresponds to an uplift gradient of 450 m/100 km along the profile. Direct evidence of the Neogene uplift seems to be preserved in the regional dip of the base of the Tertiary in the Danish basin (Figure 3c). On the geoprofile, the base of the Tertiary has a very well-defined regional dip to the south with a gradient of about 700 m/100 km. This is more than the dip estimated from Jurassic velocity data, but it is in good agreement with the theory of Neogene uplift if uniform Paleogene subsidence in the Danish basin is assumed along the profile.

A consequence of the Neogene uplift is that estimates of the maturity of the Lower Jurassic source rocks in the area (Thomsen et al., 1987) based on the present-day depth of burial of the Lower Jurassic will tend to be too low. From Figure 1b, c, it is seen that most of the areas where the base of the Lower Jurassic presently is below 3000 m have been exposed to from 500 to 850 m of Neogene uplift. The subsidence history of these potential source rock kitchens will thus be changed significantly if Neogene uplift of this magnitude is taken into account.

The contours of 500 and 850 m uplift run along the Norwegian coast in the western part of the study area (Figure 1c), but to the east the contours bend towards south, following the orientation of the Swedish coast. This indicates that Neogene uplift has affected not only Norway and Denmark, but also southern Sweden. The orientation of these uplift contours is in perfect agreement with the trends on the Quaternary subcrop map (Figure 1a). If the uplift trend, seen in Figure 1c, is extrapolated eastward toward Sweden, the interesting result emerges that 1 km of Paleogene sediments must be missing at Stevns Klint (Figure 1a), the type locality for the

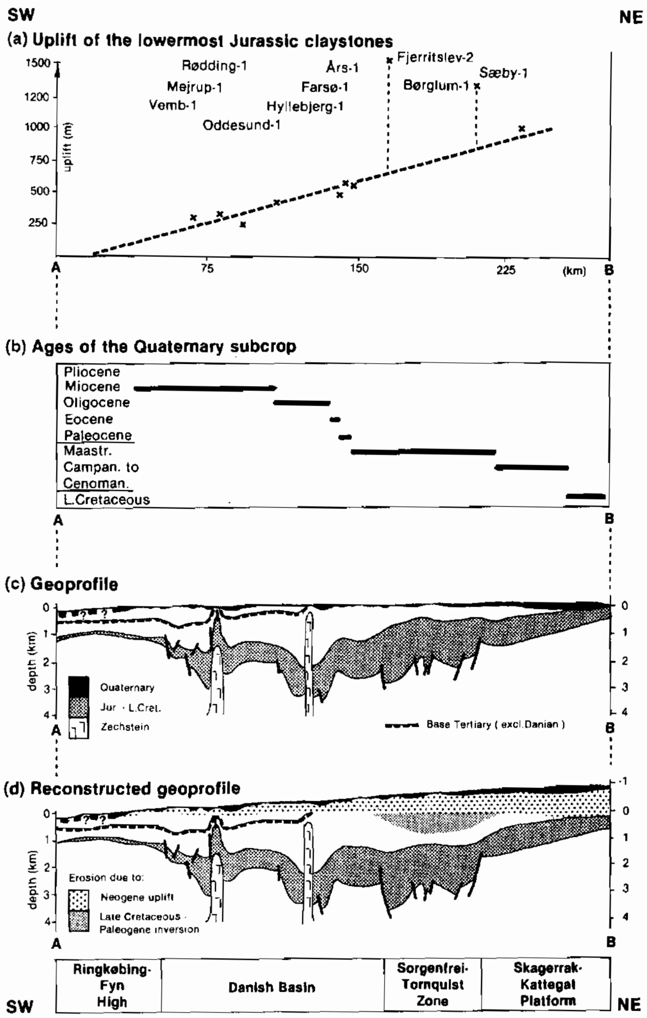


Figure 3—Uplift and geology along profile AB (see Figure 1a for location). (a) Uplift of the lower Jurassic F-Ia and F-Ib members, based on wells within 20 km from the cross section (excluding the wells on the Skive structure). (b) Age of the Quaternary subcrop (excluding local anomalies due to salt structures). Modified after Th. Sorgenfrei (Rasmussen, 1970). (c) Geoprofile to the base of the Jurassic–Lower Cretaceous. Compiled from Ter-Borch (1990), Japsen and Langtofte (1991a, b), and K. Binzer and J. Stockmarr (1991, personal communication). (d) Reconstructed geoprofile to base of the Jurassic–Lower Cretaceous, where the estimated volumes removed after uplift have been inserted.

Danian and where the Cretaceous-Tertiary boundary crops out just above sea level. In southern Sweden, more than 1000 m of sedimentary and basement rocks are likely to have been eroded as a consequence of Neogene uplift.

Estimates of pre-Quaternary erosion based on vitrinite reflectance data indicate uplift of the same magnitude as that estimated by the present study. In central Jylland, vitrinite reflectance ranges from 0.25 to 0.29 at depths close to sea level for Miocene brown coal seams (15 km southeast of the Nøvling-1 well) (E. Thomsen in Koch, 1989). These vitrinite reflectance values and the presence of brown coal indicate that a sedimentary cover on the order of 200 m must be missing (E. Thomsen, 1991, personal communication). Farther north, erosion of at least 500 m is indicated by the vitrinite reflectance-depth trend for Mesozoic sediments in 14 wells located mainly in the central part of the Danish basin, but also in the Sorgenfrei-Tornquist zone and the Skagerrak-Kattegat platform (Thomsen et al., 1987; E. Thomsen, 1991, personal communication). In the Kattegat, within the Sorgenfrei-Tornquist zone, erosion at the location of the Hans-1 well is estimated to have been at least 1500 m on the basis of vitrinite reflectance data for Lower Jurassic coal. In the northern Øresund (Figure 1a), within the Sorgenfrei-Tornquist zone, vitrinite reflectance ranging from 0.34 to 0.44 for Lower and Middle Jurassic coal found at shallow depths indicates erosion of at least 1000 m (Thomsen, 1980).

The Late Cretaceous–Paleogene Uplift

When assessing the part of the total uplift caused by the Late Cretaceous–Paleogene inversion in the Sorgenfrei-Tornquist zone, it must be recalled that the measured uplift is the difference between the deepest burial of the sediment and its present position. Neither uplift movements prior to the time of deepest burial nor subsidence in between subsequent phases of uplift, are thus "recorded."

In the Fjerritslev-2 well, the estimated uplift is 1.5 km for the Lower Jurassic. The base of the Jurassic must therefore have been buried 1.5 km deeper before the onset of the two uplift movements affecting the Sorgenfrei-Tornquist zone (i.e., at the beginning of the Late Cretaceous). From Figure 3a, it is seen that about half of this uplift, 750 m, can be explained by Neogene uplift. In the Fjerritslev-2 well, no Tertiary sediments are preserved, but some of the material removed by the Neogene uplift is likely to have been Paleogene sediments. In Figure 3c, it is seen that the Tertiary pinches out just south of the Fjerritslev-2 well and that its thickness soon reaches 500 m farther south. A cautious guess of the thickness of the eroded Paleogene in the Fjerritslev-2 well is about 250 m. Thus, 500 m of Late Cretaceous-Danian Chalk Group was eroded by the Neogene uplift and 1000 m by the Late Cretaceous-Paleogene inversion (Table 2).

In the Kattegat, lack of well control makes an estimate of Neogene uplift difficult, but 900 m at the location of the Terne-1 well and 1200 m at the loca-

Table 2. Estimated Values for the Uplift Caused by the Late Cretaceous–Paleogene Inversion in the Sorgenfrei-Tornquist Zone

Well	Apparent Uplift (m)	=	Late Cretaceous–Paleogene Uplift	+	Neogene Uplift	–	Paleogene Subsidence (Estimate)	+	Late Cretaceous–Tertiary Salt Uplift
Børglum-1	1300		700		850		250		0
Hans-1	1700		750		1200		250		0
Terne-1	1400		750		900		250		0
Fjerritslev-2	1500		1000		750		250		0
J-1	1500		550		1000		250		200

tion of the Hans-1 well are found by extrapolating the trend found in the Danish basin (Figure 1c). By analogy with this argument, the uplift caused by the Late Cretaceous–Paleogene inversion movement becomes 750 m for these two wells. In the Skagerrak, the J-1 well is drilled on a gentle salt pillow, where an estimate of the salt-induced uplift since the Early Cretaceous is 200 m (cf. Japsen and Langtofte, 1991a). The Late Cretaceous–Paleogene inversion movement at the J-1 well is thus reduced by 200 m and becomes 550 m (Table 2). This should be compared with the observation by Liboriussen et al. (1987) that the intensity of the Late Cretaceous–Paleogene inversion diminishes northwestward along the Sorgenfrei-Tornquist zone.

To visualize the effect of the two uplifts, a cross section has been constructed where the estimated volumes removed after uplift are inserted (Figure 3d). The structure at the base of the Jurassic–Lower Cretaceous and at the base of the Upper Cretaceous in the Sorgenfrei-Tornquist zone in Figure 3d thus reflects the situation in the early Late Cretaceous. At that time, the Sorgenfrei-Tornquist zone was a depocenter, and the base of the Jurassic appeared as a pronounced graben structure. On the basis of this model, Upper Cretaceous as well as Paleogene sediments are likely to have been deposited far beyond their present extent. Moreover, the depocenter of the Chalk Group must have covered the Sorgenfrei-Tornquist zone and not only the Danish basin.

Salt Movements

Positive velocity anomalies are found over the Thisted, Voldum, and Skive salt pillows (Figure 1c). This can be explained by a local component of additional uplift due to vertical salt movement. The additional amount of uplift over these salt domes is a couple of hundred meters, which is in good agreement with the height of the structural closure at the base of the Upper Cretaceous level of these structures (Japsen and Langtofte, 1991a). Several other wells are also drilled on salt-induced anticlines (see Table 1), but the lack of nearby well control makes it impossible to determine whether updoming has caused local variations in the velocity anomaly of the Lower Jurassic.

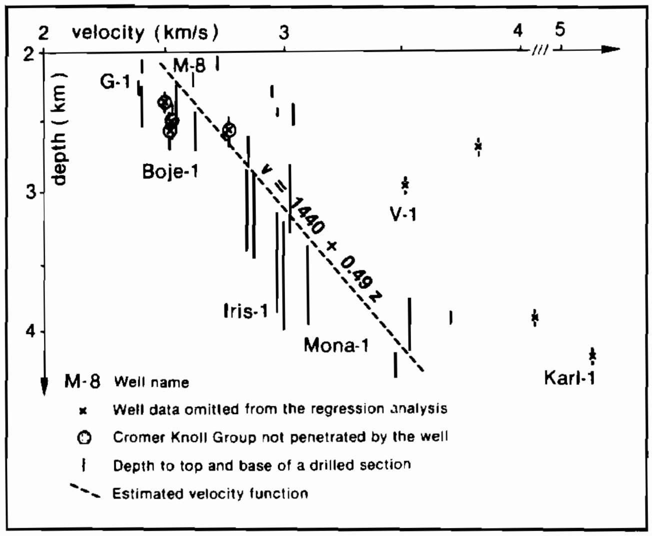


Figure 4—Interval velocities of the Lower Cretaceous (Cromer Knoll Group) in the Danish Central Trough measured in 27 wells where the penetrated thickness exceeds 20 m.

LITHOLOGICAL VARIATIONS OF THE LOWER CRETACEOUS IN THE DANISH CENTRAL TROUGH

The interval velocity for the Cromer Knoll Group (corresponding to most of the Lower Cretaceous, see Figure 5) in the Danish Central Trough shows a pronounced increase with depth (Figure 4). The estimated velocity function for the Lower Cretaceous is $V = 1436 + 0.49z$ (the well data are found in Nielsen and Japsen, 1991).

Chronostratigraphic age		Lithostratigraphic units	
Lower Cretaceous	Albian	Cromer Knoll Grp.	Rødby Formation
	Aptian		Sola Formation
	Barremian		Tuxen Formation
	Hauterivian		Valhall Formation
	Valanginian		Valhall Formation / Vyl Fm.
	Ryazanian	Farsund Form.	Vyl Fm.

Figure 5—Lithostratigraphic subdivision of the Lower Cretaceous of the Danish Central Trough (Jensen et al., 1986).

This velocity-depth trend is parallel to the velocity function of the Lower Cretaceous found in Danish wells outside the Central trough (mostly in northern Denmark): $V = 2035 + 0.51z$. The difference between the surface velocities of 600 m/s corresponds to a relative negative uplift in the Central Trough: $z_{up} = dV/k = -600 \times 2 = -1200$ m. As shown previously, the subsurface in northern Denmark is affected by pronounced uplift, whereas a negative uplift of −420 m was found for the Lower Jurassic in the O-1 well in the Central Trough. Even though normal compaction trends have not been established for the Lower Cretaceous, the difference in estimated uplift between the two regions may be interpreted as the sum of the effects of Neogene uplift in northern Denmark and undercompaction due to rapid Neogene to Quaternary subsidence in the Central Trough (cf. Nielsen et al., 1986).

The lateral variation of the velocity anomalies for the Cromer Knoll Group, shown in Figure 6c, follows a distinct pattern closely related to the structural elements and the sediment distribution of the Lower Cretaceous deposits shown in Figure 6a. The velocity anomaly contours shown in Figure 6c represent a computer interpolation of the calculated dV values for each well, including truncations and major faults as discontinuities in the interpolation. The variation of the velocity anomaly clearly reflects the geological conditions during the Early Cretaceous in the Central Trough (Michelsen et al., 1987). The sediments from this period can be divided into the younger, mainly carbonate-rich deposits of the Rødby, Sola, and Tuxen formations and the older, shaly sediments of the Valhall Formation. The younger formations have a fairly uniform thickness, whereas the thickness of the Valhall Formation varies significantly due to differential subsidence. Carbonate will thus dominate where the Cromer Knoll Group is thin. The velocity anomaly is negative in the areas of pronounced subsidence during the Early Cretaceous (the Tail End, Gertrud, and Feda grabens) where the Valhall Formation is thick. Positive anomalies are found on the more structurally stable areas, where the carbonate-rich deposits of the upper Lower Cretaceous (particularly the Tuxen Formation) dominate: Gert Ridge, Heno Plateau, and the Salt Dome province.

Clear examples of this interpretation of the lateral changes of dV are in the Mona-1 and Karl-1 wells, which are on either side of the fault that defines the northeastern edge of the Gertrud graben. The positive dV value in the Karl-1 well reflects the dominance of the calcareous Tuxen Formation, whereas the Mona-1 well is situated within the Gertrud Graben, where the Valhall Formation dominates the Lower Cretaceous sediments. In the Salt Dome province, the mainly positive velocity anomalies reflect the generally stable evolution of this area throughout the Early Cretaceous. The low dV values around the G-1 well may be explained by erosion of the calcareous uppermost parts of the Cromer Knoll Group due to later uplift of this area. The high values of dV determined in the V-1 well are caused by the high content of coarse material adjacent to the Ringkøbing-Fyn high.

The depth to the base of the Upper Cretaceous increases from about 2 km in the southern Central Trough to more than 4 km in the northwestern part (Figure 6b). This increase in depth strongly affects the interval velocities for the Cromer Knoll Group shown in Figure 6d. On this map, the geological provinces of the area are nearly indistinguishable. The lateral variations of the interval velocity are seen to be much stronger than those of the velocity anomaly because the depth dependency of the interval velocity has not been removed. As an example of this, a comparison of Figure 6c and d reveals that the interval velocities for the M-8 and the Boje-1 wells are about equal, whereas their depth-independent velocity anomalies differ, indicating that the wells belong to two separate regions (the Salt Dome province and the Tail End graben). The Iris-1, on the contrary, belongs to the Tail End graben, as does Boje-1, and hence the two wells have about equal dV values although the difference in the interval velocities for the Lower Cretaceous in the two wells are distinct. The velocity anomalies for the Lower Cretaceous in the northern Central Trough, however, are also likely to be generally lower than in the southern part of the trough. This is to be expected in consequence of the increased undercompaction of the pre-Tertiary sediments in the northern part of the trough, which is related to the rapid subsidence in the late Cenozoic (cf. Nielsen et al., 1986; Nielsen and Japsen, 1991).

VELOCITY-ANOMALY DEPTH CONVERSION

The previous discussion shows that the velocity anomaly, dV, is closely linked to the lithology and

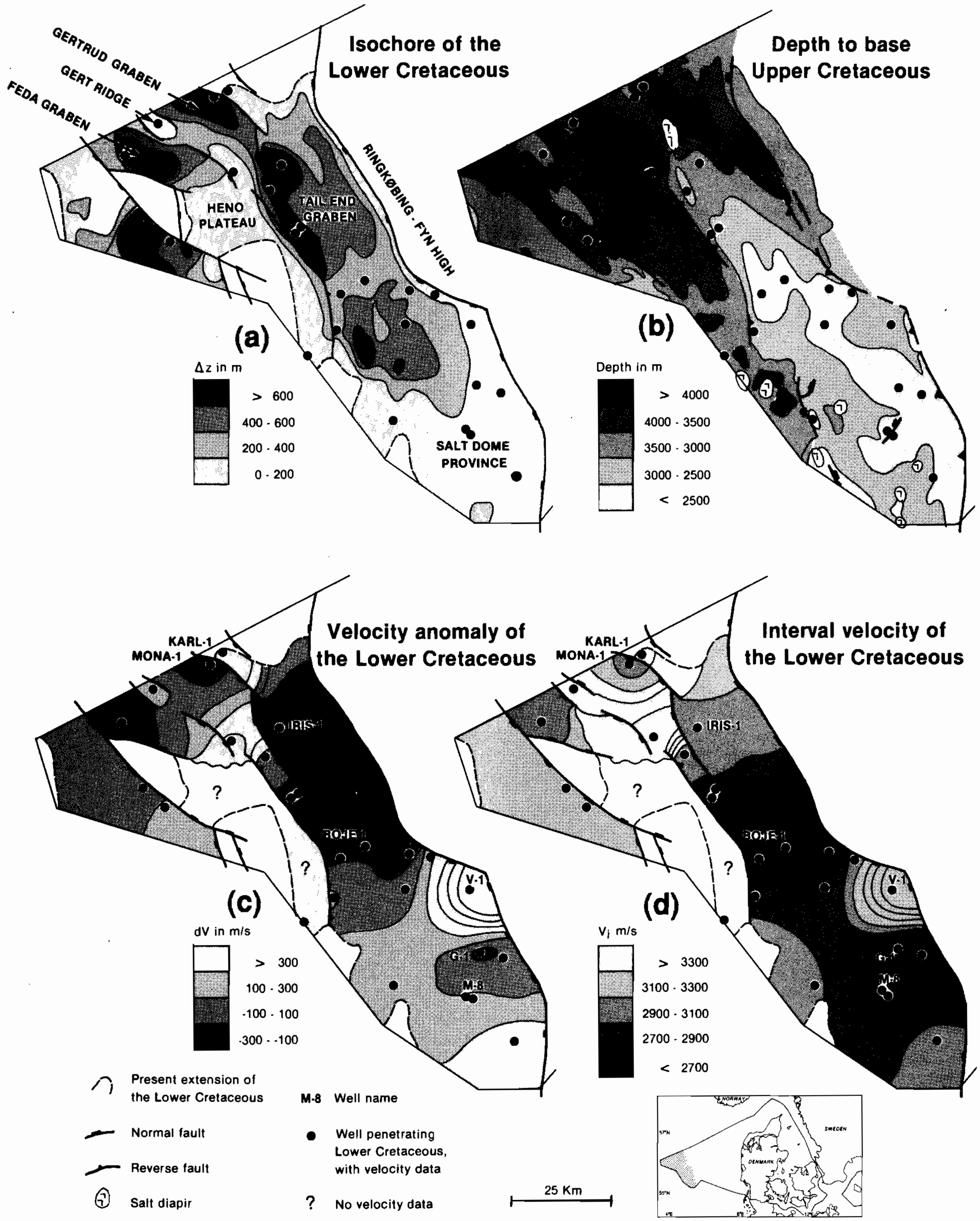


Figure 6—The Lower Cretaceous (Cromer Knoll Group) of the Danish Central Trough. (a) Isochore of the Lower Cretaceous (after Vejbæk, 1986). (b) Depth to base Upper Cretaceous. (c) Velocity anomaly of the Lower Cretaceous. (d) Interval velocity of the Lower Cretaceous.

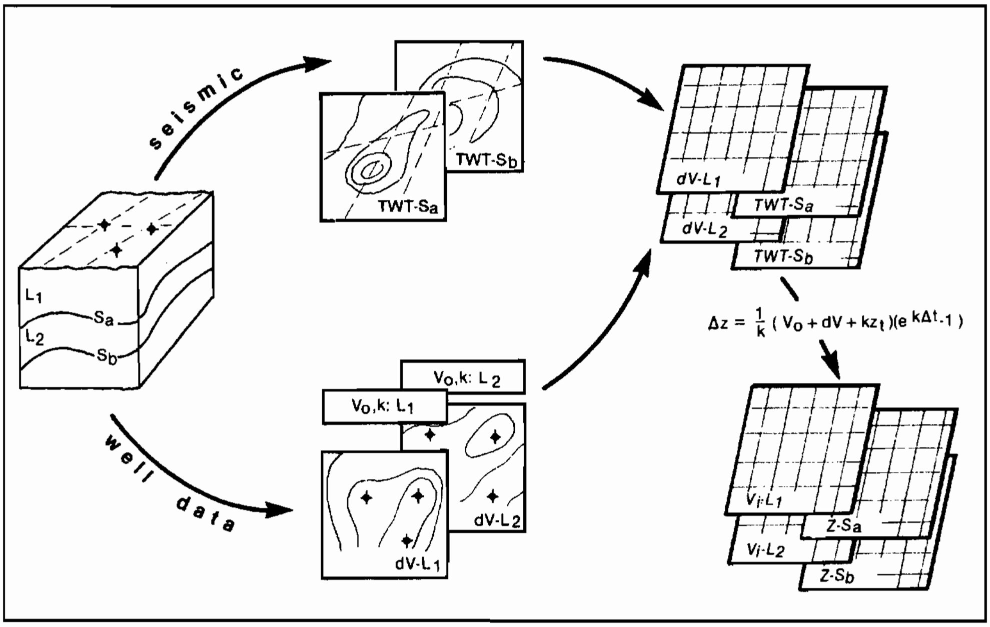


Figure 7—Flow chart showing the steps in velocity-anomaly depth conversion. The two layers, L_1 and L_2, are defined by the surfaces, S_a and S_b.

post-depositional history of a geological formation. It is thus reasonable to make predictions of dV by interpolating between well-derived values of the velocity anomaly. Contoured dV maps can then in turn be used to predict interval velocities and depths.

A computer program was designed for depth conversion of digitized seismic maps using the concept of velocity anomaly. Apart from the grids representing the two-way time maps to the tops of each of the layers in the model, the input to the program are the grids representing the velocity anomalies and the velocity parameters for every layer (see Figure 7). The depth conversion is then carried out starting with the uppermost layer by applying equation 2 and then stepwise summing up the calculated thicknesses at every grid point to achieve grids of the depths to the top of each layer as well as grids of the interval velocity for each layer, calculated as $V_i = 2\Delta z/\Delta T$.

The calculation of the depths is in itself straightforward; the more complicated part of the procedure is preparation of the time grids needed as input, particularly if the surfaces are heavily faulted. The design of the velocity model for every layer can also be a time-consuming process, involving careful evaluation of the basic well data, assigning different weights to the data points in the regression analysis to achieve geologically and physically reasonable values of the velocity parameters, V_0 and k, over the area of interest, and to assure correspondence between seismic maps and well data. Furthermore, the thorough comparison carried out by the computer of the maps defining the top and base of a layer demands accuracy in the positioning of every contour and every fault.

Regional two-way time maps of base Upper Cretaceous, top Triassic, and base Mesozoic in the Danish basin covering 55,000 km² have been depth converted applying velocity-anomaly depth conversion (Britze and Japsen, 1991; Japsen and Langtofte, 1991a, 1991b). The central part of the base Lower Jurassic depth map, shown in Figure 1b, is drawn from the map of top Triassic (Japsen and Langtofte, 1991b). The estimated linear velocity functions used for the depth conversion were $V = 2435 + 1.07z$ for the Upper Cretaceous–Danian Chalk Group, $V = 2085 + 0.52z$ for the Jurassic–Lower Cretaceous, and $V = 2625 + 0.53z$ for the Triassic. Through velocity-anomaly depth conversion, the interval velocity for the Chalk Group in the mapped area was found to range from 2400 to 4000 m/s, for the Jurassic–Lower Cretaceous from 2100 to 4000 m/s, and for the Triassic from 2800 to 5300 m/s. The thickness of these units is found to reach 2500, 3500 and 7300 m, respectively. The maximum thickness of the complete Mesozoic sequence is 9500 m. The effect of uplift was not taken explicitly into account, but the distance between the wells in most of the area is small relative to the regional scale of the uplift. The effect of the uplift on the velocities is thus expressed in the velocity anomaly maps, and in thus implicitly in the depth conversion.

An inherent advantage of the proposed method of depth conversion is that the computed depths, thicknesses, and interval velocities at the location of the wells used as input for the model will a priori be identical to well data because the velocity anomaly is defined as the calibration constant that makes the computed thicknesses identical to well data. Furthermore, the depth conversion can be modified by changing the velocity anomaly field with discontinuities such as major faults that affect the distribution of a certain unit. Such a modification, for instance, would be desirable in the case of rocks above a salt diapir, since uplifted sediments have high interval velocities relative to their present depth of burial (positive dV). If a geological model predicts lithological changes that will affect the interval velocity (such as a shelf break), estimated values of dV can be added to the map before contouring.

DISCUSSION

A linear velocity-depth function was chosen rather than a higher order approximation (Kunz, 1962) partly because of the simplicity of the solution, but also because the regression parameters based on the linear assumption are strongly influenced by data from the more deeply buried parts of a geological unit. When the velocity-depth function is assumed to be linear on a log-log plot (Faust, 1951, 1953; Acheson, 1963, 1981), the regression parameters are more influenced by the shallow data, because the input parameters are log (z) and log (V_i) rather than z and V_i themselves. In well explored basins, crests of structures are generally drilled, whereas there is uncertainty about the interval velocity downflank. It is thus desirable to emphasize the deeper rather than the more shallow data points when defining the velocity gradient. In the proposed method of velocity-anomaly depth conversion, well locations are used as fix points, thus assuring correct depths and interval velocities at these locations.

The assumption of linearity of velocity with depth of course is only valid over a certain depth range. In particular, unconsolidated sediments have a high increase of velocity with depth. But the velocity gradient is stabilized at a lower level below a certain depth of burial, from where the assumption of linearity normally will be valid over a larger range of depths. An inherent drawback in assuming a linear increase of velocity with depth is that predicted velocities may attain unrealistic values at great depths. This can be easily overcome, however, by changing the model in a way so that velocities above a certain value are reduced.

An alternative way of incorporating lateral variation of interval velocities in the linear velocity model is by fitting individual values of V_0 and k to the sonic log for every layer in every well. Such a procedure involves analysis of the entire sonic log for every layer in every well, and not just determination of transit times to the top and bottom of the layer and a single regression analysis for a number of wells as in the proposed method. The concept of velocity anomaly allows for the lateral variations of the interval velocity to be reflected in a single parameter, dV, whose variations can be depicted in only one map. The concept also lets each layer be characterized by a single pair of velocity parameters, V_0 and k. A velocity-depth model based on explicit lateral variations of V_0 and k would thus be more laborious to establish and less clear compared to a model based on variations in only one parameter, dV. Furthermore, such a model may turn out to be deceiving as the log-gradient differs from the velocity gradient in response to varying depth of burial for the layer as a whole.

To calibrate the depth conversion, the application of maps of $dV + V_0$ or maps of dV will yield the same result. But there are advantages in considering the velocity variations of a geological unit to be divided between a regional function and anomalous deviations from this function. In this way, each unit becomes characterized by a separate linear velocity function defined by V_0 and k, making it possible to compare velocity functions for different layers or to estimate uplift. Furthermore, the mean value of dV is close to zero for a normally compacted geological unit, whereas V_0 varies for different units. From the sign of a dV value, it is immediately clear whether the velocity of a layer is relatively high or low compared to the depth of burial and the regional velocity function of the layer.

A possible definition of the velocity anomaly is as a dimensionless coefficient, q, to the velocity gradient in equation 2. The actual definition of dV was chosen because it has the dimension of meters per second and thus has an intuitive relationship to velocity. Furthermore, a velocity coefficient, q, would cause V_i to increase with a higher gradient than k for $q > 1$.

A commonly used method of depth conversion is to apply $\Delta z = V_i \, \Delta T/2$ by simply contouring maps of V_i (Owen and McCormac, 1987) and then multiplying maps of V_i and $\Delta T/2$ for each layer to achieve an estimate of the thickness. The method, of course, implies the same advantage of correspondence between model-calculations and well data as the proposed method. For rough estimates of depths or if the velocity gradient of a layer is negligible, the method is quite satisfactory. However, well data for many lithological units clearly demonstrate that interval velocity does increase with depth. The dependency of velocity on depth is accounted for in the proposed method by the simple assumption of linearity of velocity with depth. Contouring of marked differences in interval velocity, which easily can be explained by different burial depths, is thus avoided.

Still another method of depth conversion of seismic maps is based on the assumption of linearity of velocity with depth as is the method of velocity anomalies, but without explicit consideration of lateral variations of the velocity field (Ter-Borch, 1990).

Depth conversion is carried out by application of equation 2 with dV set equal to zero, thus keeping both V_0 and k constant over the entire area. In this way, the preparation of the depth map is facilitated by the omission of the velocity anomaly map, but in general there will be misties between the calculated depth map and the well data. These misties are removed by manually editing the contours around wells, whereas application of velocity-anomaly depth conversion allows for lateral velocity variations and results in perfect ties between the calculated maps and well data.

CONCLUSION

Assuming linearity of velocity with depth, the thickness of a layer is a function of the depth to the top of the layer, its time thickness, and the linear velocity parameters, V_0 and k. The velocity anomaly, dV, is defined as the correction needed to calibrate the linear velocity model to well data, expressed as the change in the surface velocity, V_0, that for a given layer will make the calculated thickness fit well data. The influence of depth of burial is greater on the interval velocity than on the velocity anomaly, which therefore reflects differences in lithology and the postdepositional history of the sedimentary rock. In the special case of a laterally homogenous geological unit, the velocity anomaly is proportional to the apparent uplift. Examples show that the velocity anomaly, dV, is related to lithology, overpressure, erosion, and uplift. The velocity anomaly characterizes a geological region better than interval velocity.

This geological interpretation of dV makes it reasonable to estimate dV at arbitrary locations by interpolation between values of dV determined in wells. Using a general geological model, the velocity anomaly field may be further modified by introduction of additional points of guessed dV values and of important faults or paleogeographical lineaments.

Seismic two-way time maps over an area can be depth converted one by one, once the linear velocity parameters and the velocity anomaly map of each layer are established. Velocity-anomaly depth conversion provides a consistent routine depth-conversion method for digital time maps taking into account both lateral and vertical velocity variations. Through this method, the predicted velocities and depths of the deeper, undrilled parts of basins can be constrained by the knowledge of the statistical distribution of the interval velocities of the layers in an area and by possible additional geological information, such as fault patterns and sedimentation models. The resulting depth maps will be in exact agreement with well data, and the computed depths in between the wells will reflect both the geological and the geophysical knowledge of the entire area under investigation.

APPENDIX

Consider a multilayered-earth model, where the velocity increases linearly with depth within each layer. The depth to any interface can be calculated, given the linear velocity parameters, V_0 and k, and the one-way time thickness, Δt, of each layer: Starting from the earth's surface, the thickness of each layer is calculated using the depth to the base of the layer above it as a boundary condition.

The thickness of the first layer, characterized by the linear velocity function $V_1 = V_{01} + k_1 z = dz/dt$, is obtained by integrating over a homogenous halfspace from the earth's surface, where $z = 0$ and $t = 0$, to the base of the first layer, where $z = z_{b1} = \Delta z_1$ and $t = t_{b1} = \Delta t_1$. This gives

$$\Delta t_1 = \int_0^{z_{b1}} dt = \int_0^{z_{b1}} \frac{1}{V_1} dz = \frac{1}{k_1} \ln\left(\frac{1}{V_{01}} \times k_1 z_{b1} + 1 \right)$$

This can be solved for the unknown depth to the base of the first layer, z_{b1}, giving,

$$z_{b1} = \frac{V_{01}}{k_1} (e^{k_1 \Delta t_1} - 1)$$

The derivation of this function was described by Slotnick (1936). The second layer in the model is characterized by the linear velocity function $V_2 = V_{02} + k_2 z$, and the depths to the top and base are $z_{t2} = z_{b1}$ and z_{b2}, respectively, and $t_{t2} = t_{b1}$ and t_{b2} are the corresponding one-way traveltimes. We obtain

$$\Delta t_2 = t_{b2} - t_{b1} = \int_{z_{t2}}^{z_{b2}} \frac{1}{V_2} dz = \frac{1}{k_2} \ln\left(\frac{k_2}{V_{02}} z_{b2} + 1 \right) - \frac{1}{k_2} \ln\left(\frac{k_2}{V_{02}} z_{t2} + 1 \right)$$

where Δt_2 is known and z_{t2} was calculated as z_{b1} for the first layer. To obtain the unknown depth of the base of the second layer:

$$z_{b2} = \frac{V_{02}}{k_2} \left[\exp\left(k_2 \Delta t_2 + \ln\left\{ \frac{k_2}{V_{02}} z_{t2} + 1 \right\} \right) - 1 \right]$$

$$= \frac{V_{02}}{k_2} \left(e^{k_2 \Delta t_2} - 1 \right) + z_{t2} e^{k_2 \Delta t_2}$$

where z_{t2} equals z_{b1}, the above calculated thickness of the first layer. In this way, the depth to any surface can be found by stepwise calculating the thickness of each layer starting from the earth's surface and thus calculating the depth to the top of the next layer.

In general, the thickness, Δz, of a layer, characterized by the linear velocity parameters, V_0 and k, the depths to the top and the base of the layer, z_t and z_b, and the one-way time thickness Δ_t, can be expressed as,

$$\Delta z = z_b - z_t = \frac{V_0}{k} (e^{k\Delta t} - 1) + z_t e^{k\Delta t} - z_t$$

or

$$\Delta z = \left(\frac{V_0}{k} + z_t \right) (e^{k\Delta t} - 1)$$

and the interval velocity of the layer will be,

$$V_i = \frac{\Delta z}{\Delta t} = \frac{1}{k\Delta t}(V_0 + kz_t)(e^{k\Delta t} - 1)$$

If one-way time, t, is substituted by two-way time, $T = 2t$, we get equation (1):

$$V_i = \frac{2}{k\Delta t}(V_0 + kz_t)(e^{k\Delta t/2} - 1)$$

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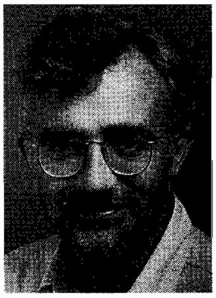

Peter Japsen graduated form the University of Copenhagen, Denmark, with a Master's degree in geophysics and physics in 1980. Since then, he has been with the Geological Survey of Denmark, where he has been working on a variety of geophysical problems related to the exploration boom in Denmark in the 1980s. Quantitative basin modeling studies made the production of reliable depth maps mandatory, which triggered his interest in depth conversion and the geological analysis of seismic modeling.