The reservoir development of the Late Maastrichtian–Early Paleocene Ormen Lange gas field, Møre Basin, Mid-Norwegian Shelf

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Abstract: The giant Ormen Lange Gas Field is situated in the Møre Basin, Norwegian Sea, at water depths around 1000 m. The reservoir interval consists of a lower heterolithic unit mainly of Maastrichtian age (Jorsalfare Formation) and an upper sandstone unit of Danian age (Egga Member, Våle Formation), separated by a widely distributed mudstone unit ('Våle Tight'), and consists of both high and low density turbidites in association with fine-grained background sediments. The reservoir interval spans the Cretaceous–Tertiary boundary, without any major stratigraphic breaks. The main architectural elements of the reservoir are: channel-dominated deposits, channelized lobe deposits and non-channelized frontal splay or fan fringe deposits. The spatial distribution of these facies elements indicates a dynamic system with changing sediment supply through time. Core and log studies from the wells have been integrated with high-resolution biostratigraphy to develop a dynamic depositional model.

During latest Cretaceous and Paleocene times, a series of subtle sub-basins developed along the southeastern Møre Basin margin as a result of differential subsidence above Jurassic extensional faults. The sub-basins created a stepped slope topography that influenced the flow pattern of turbidity currents.

In a landward position, east of the Ormen Lange Field, the Cretaceous–Tertiary boundary is characterized by a major unconformity, implying erosion and sediment by-pass. In early Danian time, uplift and rotation of the provenance area to the east (Norwegian mainland) led to extensive erosion and redistribution of sandy sediments into the Møre Basin and caused deposition of the Egga Member. The turbidites filled the intraslope basins initially in a basinwards-stepping sense, but later sedimentation switched landward.

Onshore structural, provenance and geomorphological data, combined with offshore structural, seismic and sedimentological data, yield an integrated interpretation of the deep-water depositional system in the greater Ormen Lange area. Present fjords at Møre, controlled by the Møre–Trøndelag fault zone, were valley systems that fed sands to a point source updip of the Ormen Lange area. The point source was located at the transition/relay zone between a narrow shelf area in the south and a broad shelf area in the north. In addition to being a key transition zone for both Caledonian and Jurassic structures, this zone area is also the 'landing' point of the major oceanic Jan Mayen Fracture Zone.

Keywords: Ormen Lange Field, Paleocene, Maastrichtian, Møre Basin, basin floor fan, Egga Member, Våle Formation, K–T boundary

The giant gas field Ormen Lange, in the southeastern Møre Basin (Fig. 1) was discovered by Norsk Hydro in 1997. The field contains about $500 \times 10^9\,m^3$ of gas (GIIP). This paper reviews the depositional history of the Ormen Lange Field from source area to deep-water basin sink and revises and extends the work of Vågnes *et al.* (1998), Gjelberg *et al.* (2001), Martinsen & Gjelberg (2001) and Smith & Møller (2004). The new data include onshore geomorphological, structural and provenance data with offshore seismic and new well data. In addition, a new biozonation for all the wells in the field has been considered in the depositional model.

Attempts to gain information on primary sedimentary depositional and morphological features from seismic attribute maps were unsuccessful. Therefore, the stratigraphic and sedimentary information from the five wells drilled on the Ormen Lange structure are of essential importance in understanding the sedimentological development of the Ormen Lange reservoir. For all wells cores were cut through most of the reservoir, giving important samples for studies of reservoir parameters, mineralogy, biostratigraphy, fluid inclusions and palaeomagnetism.

Structural framework

Jurassic structural configuration

The Ormen Lange structure is one of several Cenozoic dome structures formed in the Norwegian–Greenland Sea (Blystad *et al.* 1995). The onset of doming and creation of the Ormen Lange structure closely coincided with the onset of seafloor spreading in the Norwegian–Greenland Sea (Doré *et al.* 1997), suggesting a change in the far-field plate-tectonic stress as a possible causal mechanism. Significant compression occurred during Oligocene–Miocene time and appears to have continued up to the present (Vågnes *et al.* 1998).

The development of the Ormen Lange structure was strongly influenced by reactivation of underlying late Jurassic and early Cretaceous fault blocks, resulting from Jurassic extension on mainly NE–SW orientated faults south of the Trøndelag Platform and the Halten Terrace (Blystad *et al.* 1995; Grunnaleite & Gabrielsen 1995; Gjelberg *et al.* 2001). Several N–S-trending faults also are present in the Klakk fault complex (Blystad *et al.* 1995)

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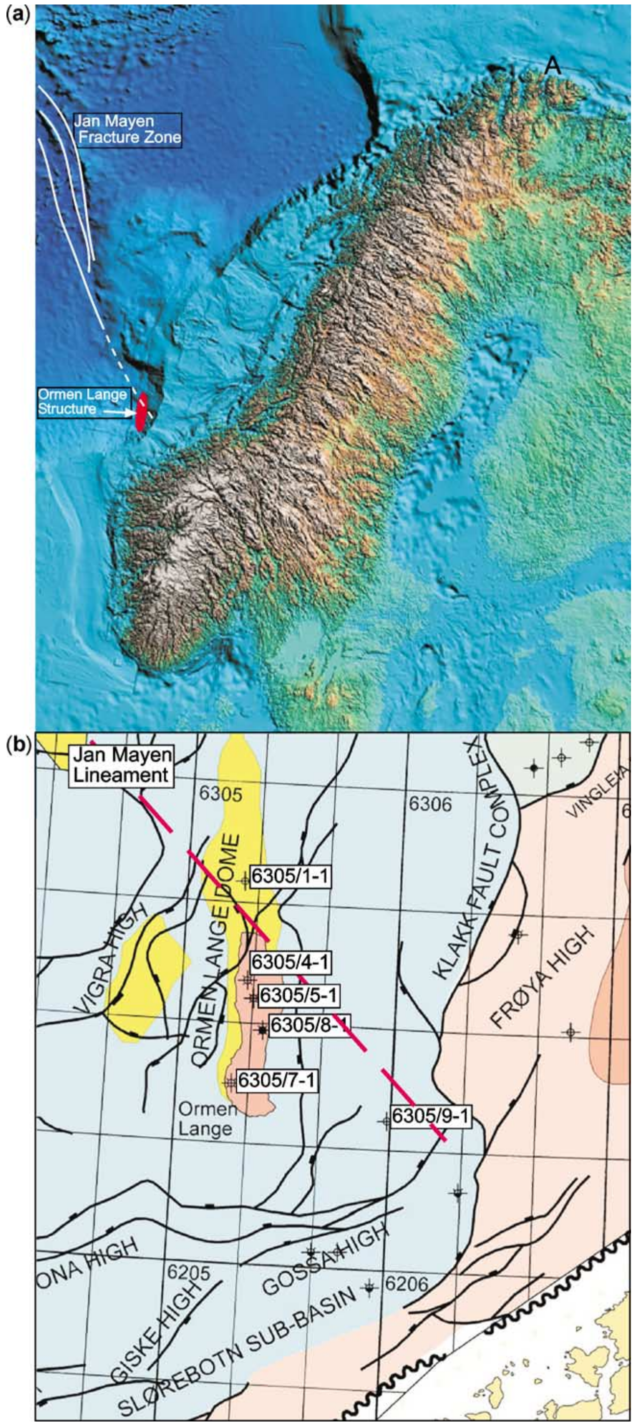

Fig. 1. (**a**) Location of the Ormen Lange structure in relation to the Jan Mayen Fracture Zone. (**b**) Jurassic Structural elements at the eastern margin of the Møre Basin, with the location of the Ormen Lange Field and the extension of the Jan Mayen Lineament towards the Norwegian mainland.

and align with the present western margin of the Frøya High. There is ample seismic evidence that several Jurassic faults were reactivated during the Cretaceous period (e.g. Brekke 2000). In addition, because the Cretaceous period was dominated by fine-grained sedimentation, albeit with high rates of deposition in the Late Cretaceous, the relief created was sustained for long periods of time after the Jurassic because of differential subsidence and compaction. Major structures, such as the Slørebotn Sub-basin, the Gossa, Vigra, Ona and Frøya highs (Fig. 1) and the Ormen Lange anticline are, thus, likely to have acted as basinal highs and lows, trapping or controlling the distribution of latest Cretaceous and Palaeogene deep-water sediments. The primary distribution of thick Paleocene sands within the Slørebotn Sub-basin (Fig. 2) is clear evidence of the role of the Jurassic structures in controlling Paleocene deep-water sedimentation with important implications for the distribution of reservoir sandstone within the Ormen Lange licenses.

Cenozoic uplift

Recent work both in the northern North Sea and in the mid-Norway area suggests that several phases of uplift of Fennoscandia took place through the Cenozoic (Riis & Fjeldskaar 1992; Martinsen *et al*. 1999; Brekke *et al*. 2001; Gjelberg *et al*. 2001). Plio-Pleistocene uplift is best documented, but the presence of several unconformities of regional extent and a significantly tilted succession on seismic lines supports that at least five phases of uplift may have occurred (Martinsen *et al*. 1999): Latest Maastrichtian and earliest Paleocene; Early Eocene; Early Oligocene; Early–mid Miocene; and Plio-Pleistocene.

While the last phase is closely linked to Plio-Pleistocene glaciations (Riis & Fjeldskaar 1992), the preceding phases are probably tied to regional tectonics (Martinsen *et al*. 1999). The latest Maastrichtian–early Paleocene phase is of interest to the Ormen Lange geological story because it may have been the primary cause for the supply of clastic material from the Norwegian mainland to the deep-water areas in the Møre Basin (Gjelberg *et al*. 2001). Sands were delivered into a basin, which throughout most of the Cretaceous was dominated by deposition of fine-grained material (Dalland *et al*. 1988; Vergara *et al*. 2001). The mechanism for this uplift phase was probably shoulder uplift of the basin margin, taking place prior to Atlantic break-up.

Shelf and onshore extension of oceanic fracture zones

Several NW–SE oceanic fracture zones occur within the Møre and Vøring basins (see Brekke (2000) and references therein). One of these zones, the Jan Mayen Lineament, is of importance to the Ormen Lange area (Fig. 1). The postulated onshore extension of this lineament coincides with the southern extension of the Klakk fault complex, the southern end of the Frøya High and the northeastern end of the Gossa and Ona highs. As will be shown below, this area coincides with the postulated position of the feeder system to the greater Ormen Lange deep-water depositional system. Thus, it is assumed that the presence of the Jan Mayen Lineament had some influence on the delivery of latest Maastrichtian and earliest Paleocene sand to the deeper-water areas in the Møre Basin prior to Atlantic break-up.

Onshore geomorphology

Present-day fjords onshore mid-Norway show a systematic semi-rectangular pattern. This pattern corresponds with the long-lived structures of the Møre–Trøndelag fault zone (Fig. 1). The NE–SW-orientated fjord segments are orientated slightly obliquely with respect to the present-day coastline and several appear to converge towards the area near the mouth of the Romsdalsfjord. This convergence area is located immediately inboard of the structural relay zone from which the Ormen Lange deep-water depositional system was sourced (see below). It is assumed that because of the long-lived structural context, the fjords occur in former palaeovalleys along which sediments were transported to the offshore area upslope of the Ormen Lange Field. Their orientation caused a sediment sink near the mouth of the present-day Romsdalsfjord, from which the sediments were remobilized and transported onwards to deep-water areas in the Møre Basin. This assumption is supported by the fact that the updip extension of the mapped sand thickness at Ormen Lange can be traced directly into this area and that no other deep-water fan point source has been identified in the area.

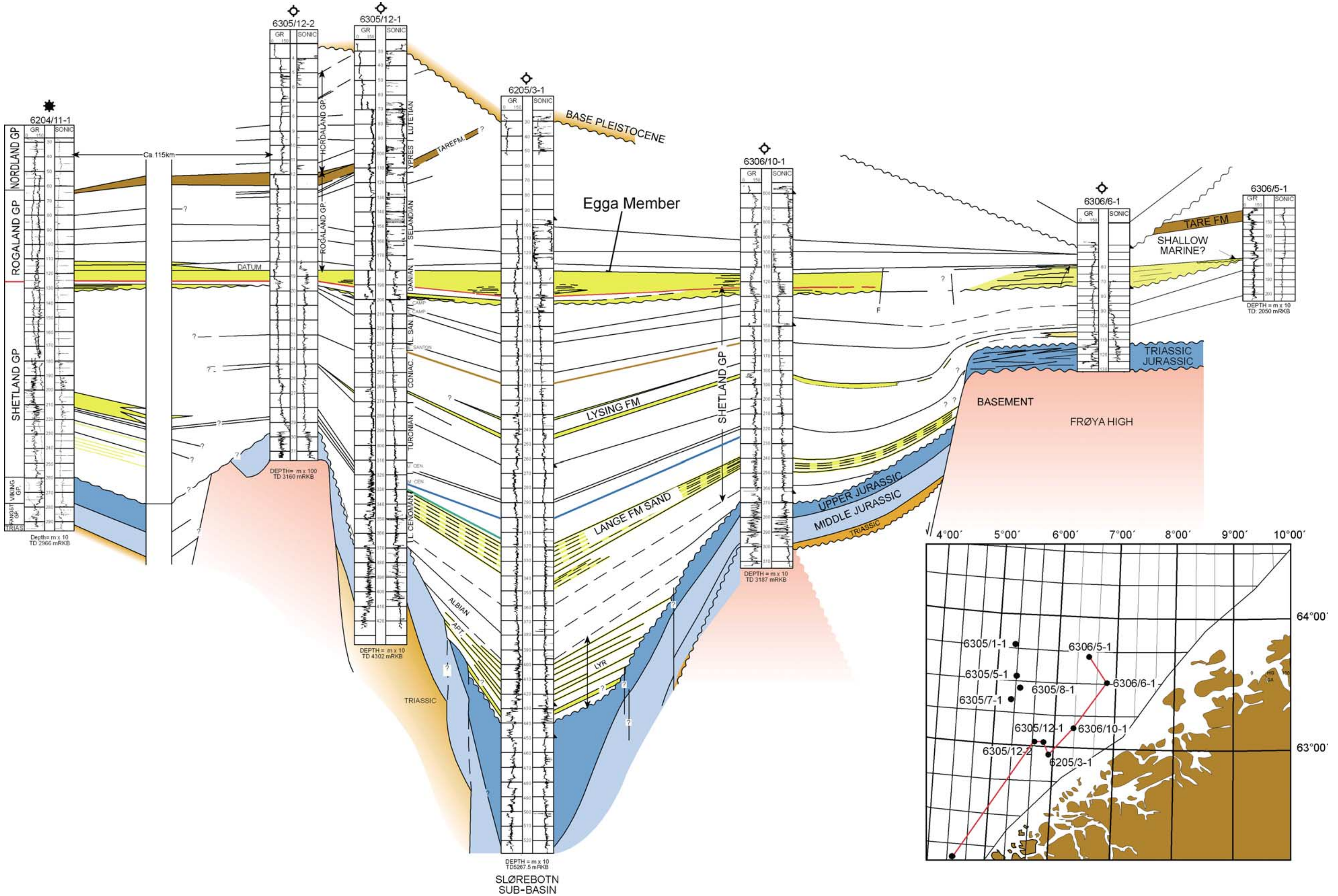

Fig. 2. Correlation of the Cretaceous and Tertiary succession along the eastern margin of the Møre Basin. The Egga Member is located above a major unconformity spanning the K–T boundary. The member consists mainly of relatively shallow-water, delta front turbidites.

Basin physiography

Shelf areas

The broad Trøndelag Platform and the Halten Terrace are main structural features to the north of the southeastern Møre Basin area. The relative distance to the Cretaceous–Cenozoic deep-water areas across these structural elements is comparatively much larger than from the Norwegian mainland across the Slørebotn Sub-basin (sustained by interpretations of sedimentary successions; see Dalland *et al*. (1988) and several papers in Martinsen & Dreyer (2001)). It is tempting to interpret this difference as a reflection of the presence of a wide Cretaceous–Cenozoic shelfal area in the Trøndelag Platform/Halten Terrace area, while a much narrower shelf existed at the eastern margin of the Møre Basin, probably reflecting change in plate polarity (i.e. change from upper to lower plate configuration) across the Jan Mayen Fracture Zone. This had important influence on the effectiveness with which sands were delivered to the deeper-water areas in the Møre Basin.

Sub-basin areas

Based on the present-day position of main structural features (Fig. 1) and the time isochore distribution of latest Cretaceous and early Paleocene sediments (Fig. 3), it is argued that the greater Ormen Lange area seafloor topography consisted of a series of structurally bounded lows or ‘minibasins’ (Fig. 3). The assumption that Jurassic structures had some influence on the distribution ofz sediments also in the latest Cretaceous and early Paleocene (see above) introduces ideas that several minor sub-basins or ‘minibasins’ could have existed between the Norwegian mainland and the Møre Basin floor. Within the Cretaceous and early Paleocene sub-basins at the eastern margin of the Møre Basin, the sediment influx was probably so high that the rate of sedimentation most of the time outpaced rates of structural movement, leaving sub-basinal areas without significant seafloor depressions. However, in periods with high tectonic activity and relative low sediment input, shallow seafloor depressions may have developed above these basins.

There are four smaller basins.

- The Slørebotn Sub-basin (Blystad *et al.* 1995) is located between the Gossa High, the transition to the Frøya High (Fig. 3) and the present truncation line underneath the sub-Pleistocene unconformity. This basin has, by far, the thickest Paleocene succession (at least 200 ms in places). Gjelberg *et al.* (2001) have discussed the Tertiary fill of the sub-basin, whereas Jongepier *et al.* (1996) have discussed the Mesozoic fill.
- The ‘Frøya High sub-basin’ (informally named here) is located along the eastern margin of the mapped area.
- The ‘Gossa sub-basin’ (informally named here) (Fig. 3) encompasses the eastern part of the Ormen Lange licenses and is limited to the east by the thickness reduction at the border to the Slørebotn Sub-basin and to the west by the constriction in the thickness anomaly at the border to the Ormen Lange system itself. This basin has a much smaller Paleocene sediment thickness than the Slørebotn Sub-basin.

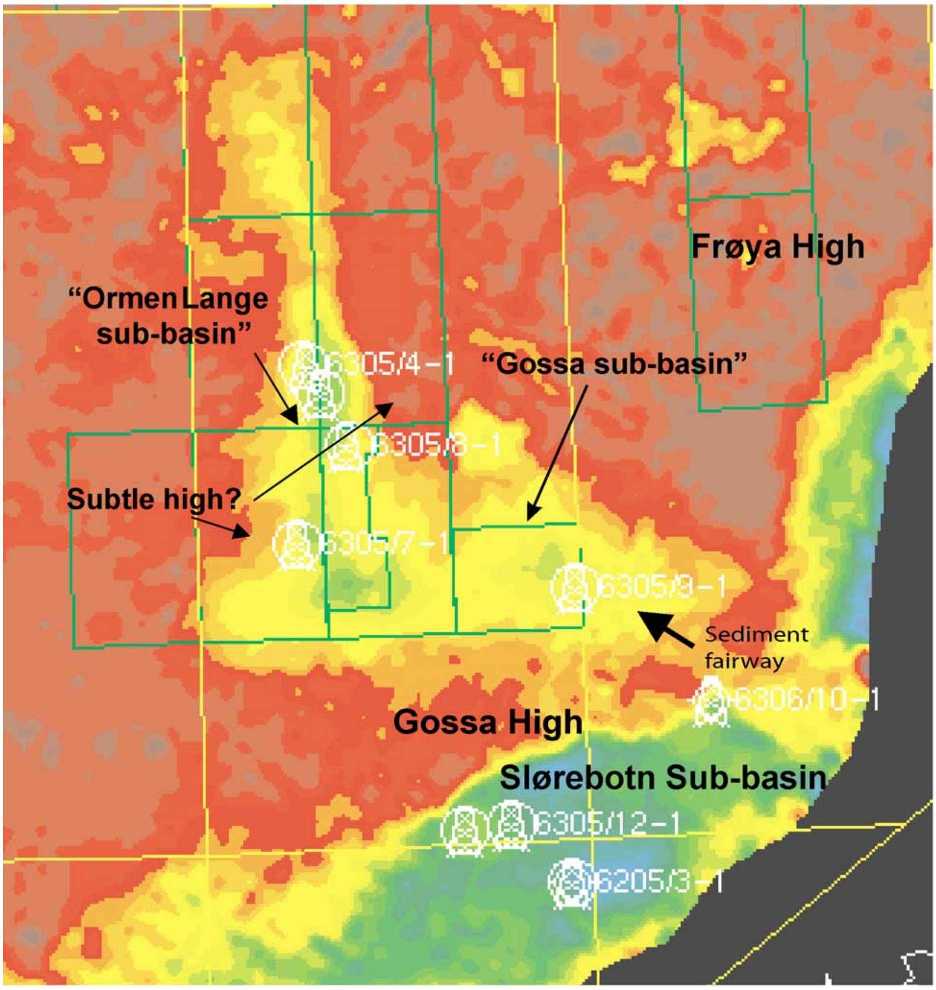

Fig. 3. Time isochore map for the Upper Maastrichtian/Lower Paleocene succession at the eastern margin of the Møre Basin with definition of ‘sub-basinal’ areas.

The structural definition of this proposed sub-basin is poor, particularly at its downdip end at the passage to the Ormen Lange area where the thickness seems to be retained but the plan view extent of the Paleocene system is clearly reduced.

- The Ormen Lange 'sub-basin' (Fig. 3) (informally named here) is defined at its southeastern end by the abrupt eastward turn of the isochore anomaly from the main N–S trend of the Ormen Lange system and at its eastern end by a sudden thinning, probably caused by reactivation of an underlying Jurassic fault. This sub-basin extended to the north, towards the location of the 6305/1-1 well where Paleocene sandstones, albeit with reduced reservoir quality also occur.

Admittedly, the delineation of the four sub-basins is defined by sediment distribution (Fig. 3), but coincidence of basin boundaries with underlying Jurassic structures is compelling.

The present-day subsurface structural morphology of the greater Ormen Lange deep-water system area is considered to be significantly influenced by post-Paleocene movement and uplift/ subsidence along major structures of Fennoscandia. The structural picture in the area is mainly inherited from the Jurassic (see above), but later uplift obviously created structures such as the Ormen Lange Dome (e.g. Brekke (2000) and references therein). These later movements, thus, resulted in the observation that previously deep-water sands, which obviously were originally deposited in lows, are now on structural highs.

Maastrichtian and Paleocene stratigraphical framework

The reservoir interval of the Ormen Lange Field comprises sand both of Maastrichtian (Springar/Jorsalfare Formation) and Danian age, informally called the Egga Member (Gjelberg *et al.* 2001). The Egga Member comprises the lower part of the Våle Formation (Danian age). In addition to the Våle Formation, the Paleocene of the Møre Basin also comprises the Lista Formation (Selandian) and the Sele Formation of latest Paleocene–earliest Eocene age (Fig. 4). In the Møre Basin it is generally difficult to distinguish between the Sele and Lista formations, based on the original definition of Isaksen & Tonstad (1989). They differentiated between the two formations on the basis of a slightly higher tuff content in the Sele Formation, compared to the Lista Formation below, and an abrupt decrease of interval transit time downwards across the formation boundary. It is, however, easier to apply the Mid-Norwegian shelf stratigraphy of Dalland *et al.* (1988) for this area, who defined only one formation for the entire Paleocene – the Tang Formation – which consists mainly of dark grey to brown claystone, with minor sandstone and limestone. The main change in the Tertiary stratigraphy between the Northern North Sea area and the Trøndelag Platform area occurs close to the Jan Mayen Lineament (Fig. 1) and it is, therefore, reasonable to use this as the geographical border for the application of the two stratigraphical schemes.

The most prominent feature of the Maastrichtian and Paleocene stratigraphy in the Møre Basin is the base Tertiary unconformity that is well developed and present in all wells along the margin (Figs 2 and 4). Most of the Maastrichtian and the lower Paleocene stratigraphy is absent in these wells. The origin and nature of this unconformity is not well understood; however, it is tentatively suggested that it may represent an erosion surface related to a significant relative sea-level fall during the late Maastrichtian and early Paleocene, where both the basin margin and platform areas may have been exposed. This relative sea-level fall was probably related to regional epeirogenic uplift of the mainland (Riis 1996; Gjelberg *et al.* 2001). In the Møre Basin, on the other hand, there is a continuous stratigraphical development throughout the Late Cretaceous–Early Tertiary (Fig. 4).

The Egga Member has been introduced as an informal name for the sandstones of Paleocene age in the wells in the Slørebotn Sub-basin. Within the same sub-basin there are little or no sandstones of Maastrichtian age because there is a major stratigraphic break between the Campanian and the Paleocene (Fig. 4). This implies that the Maastrichtian sandstones of the Springar (Jorsalfare) Fm. in the Møre Basin are, in principle, unnamed.

In the Ormen Lange licenses, a Northern North Sea stratigraphical nomenclature has been used (cf. Isaksen & Tonstad 1989). This is formally wrong, but since the nomenclature has been used in reservoir zonation schemes, the task of changing the scheme is cumbersome. In the present paper, the established North Sea stratigraphical nomenclature will be used, but with reference to the Mid-Norway stratigraphy (*sensu* Dalland *et al.* 1988).

Depositional patterns

The Egga Member at the eastern margin of the Møre Basin

The Egga Member has been known as a prominent sandstone interval in the Slørebotn Sub-basin area when it first was penetrated in well 6205/3-1 (Fig. 2). During 1989–1994 several wells were drilled along the eastern margin of the Møre Basin, from the Selje High in the south to the Frøya High in the north, and the Egga Member is present in all these wells. It consists of thick, amalgamated medium- to coarse-grained turbiditic sandstones, with a maximum total thickness close to 150 m (Fig. 2). The base of the Egga Member in the Slørebotn Sub-basin area and in other areas close to the eastern margin of the Møre Basin is a significant unconformity with the Danian succession, usually overlying strata of Campanian age. In the Halten Terrace and Trøndelag Platform, the base Tertiary unconformity is also well developed, as an angular unconformity in the more proximal regions to the east. None of the wells drilled on the Halten Terrace and Trøndelag Platform areas have proven sand above the unconformity and therefore, differ considerably from the development in the Møre Basin.

The Egga Member in the 'Ormen Lange sub-basin'

The K–T unconformity was initiated during the Maastrichtian, probably with shelf erosion and sediment bypass. The coarse clastic sediments that bypassed the slope areas were redeposited as turbidites further out in the Møre Basin and represent the reservoir in the Ormen Lange Field, consisting of alternating turbiditic sandstones and mudstones filling in the Paleocene 'Ormen Lange sub-basin'. This development suggests that the generation of turbidity currents occurred occasionally during long periods of quiescence, but without significant stratigraphic breaks throughout the upper Maastrichtian–lower Paleocene succession. Direct hydrocarbon indicators, such as flat spots, amplitude, AVO and frequency anomalies, are well expressed in the southern and central part of the structure (Fig. 5), but become poorly defined northwards. This poor definition is principally because of poor reservoir quality but also because of poor seismic data caused by seafloor irregularities, shallow gas and a strongly intersected reservoir related to polygonal faulting (Fig. 6). Seismic attribute maps, on the other hand, suggest that sandy lithologies related to the Ormen Lange reservoir may be present also in the area just north of well 6305/1-1 (Fig. 7). Isochore maps from the Egga Member (based on detailed seismic mapping) suggest that the member is thickest and best developed in the southeastern part of the structure, where it shows a radial pattern of thickening and thinning reflecting primary depositional features. It becomes thinner northwards into the saddle area just south of well 6305/8-1, whereas it increases in thickness in the area where well 6305/5-1 is located (Fig. 7).

The reservoir interval of the Ormen Lange Field comprises the Jorsalfare (Springar) sandstones (Maastrichtian) in the lower part, overlain by the Våle Heterolithic and Våle Tight intervals (Danian), accompanied by the Egga Reservoir Unit (Danian), which represents the main reservoir interval. In addition a thin zone called the Egga Tight has been defined in both wells 6305/7-1

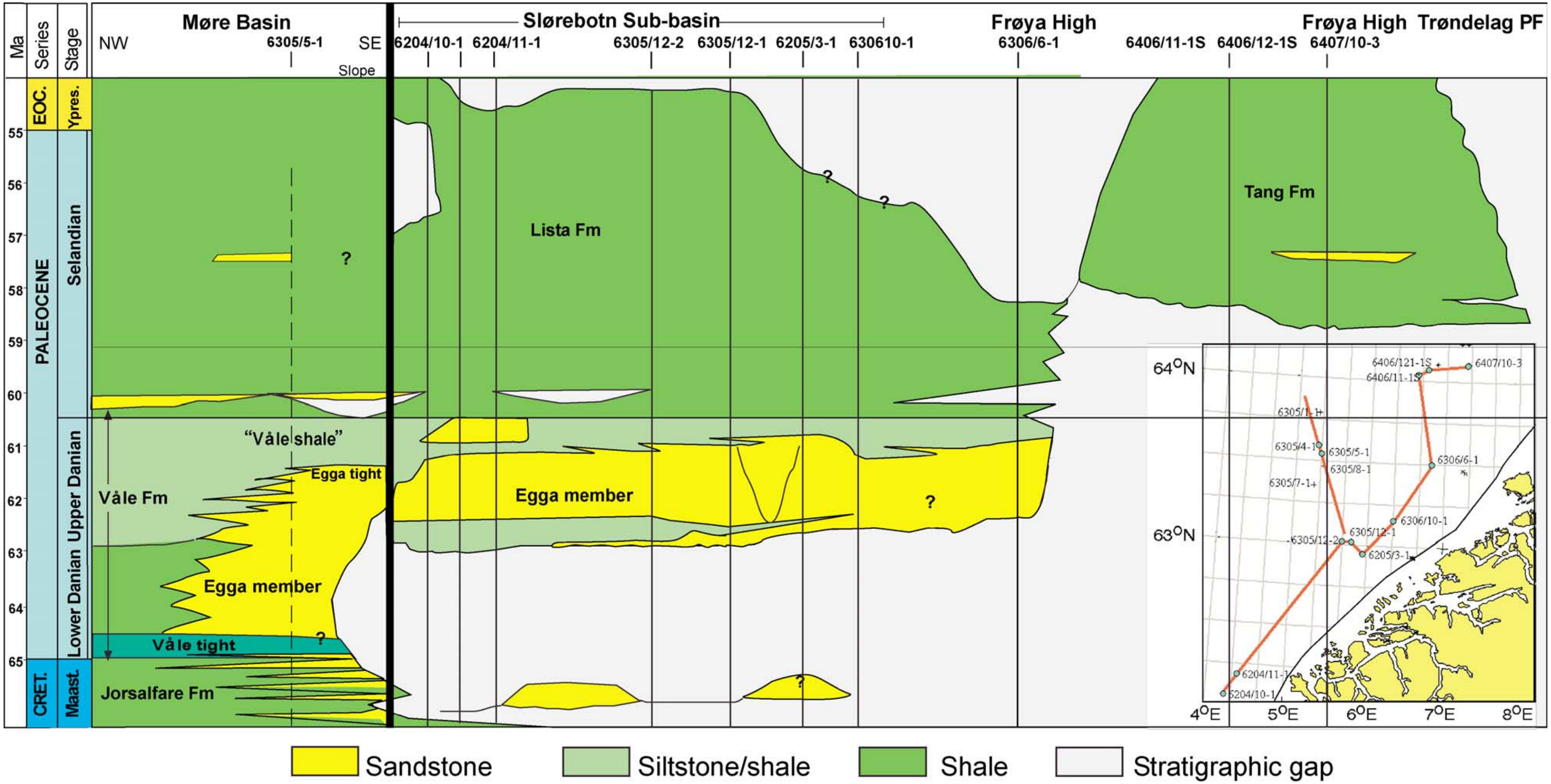

Fig. 4. Uppermost Cretaceous–Paleocene stratigraphy at the eastern margin of the Møre Basin.

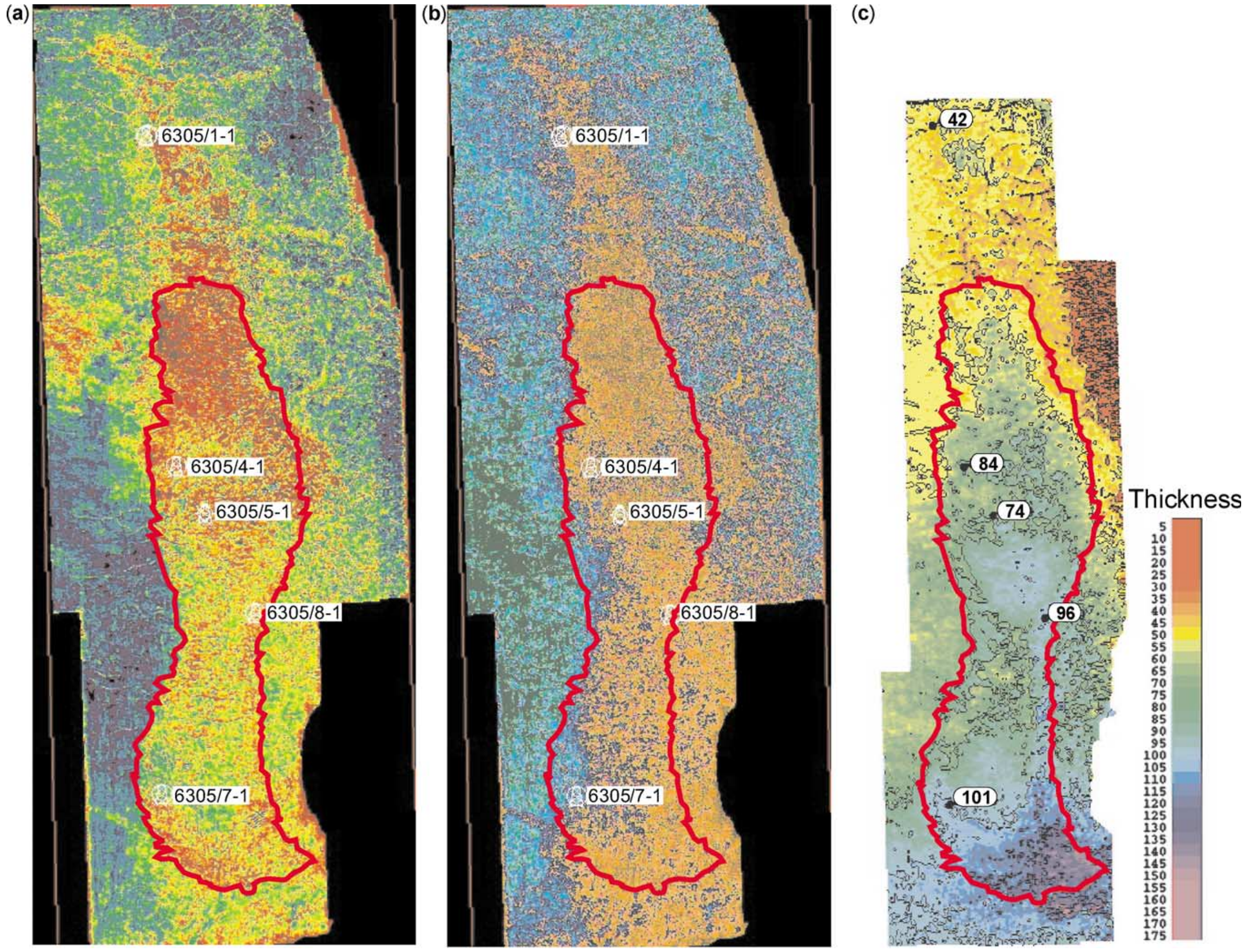

Fig. 5. Seismic attribute maps and isochore maps from the Ormen Lange Field: (**a**) horizon-based seismic amplitude map for the top Våle Formation; (**b**) volume-based seismic amplitude map for the top Våle Formation reflector (offset − 12 ms, window 24 ms); (**c**) isochore map for the Lower Paleocene succession of the 'Ormen Lange sub-basin'.

and 6305/5-1 on top of the Egga Reservoir Unit. Both the Våle Heterolithic/Våle Tight and the Egga Member are parts of the Våle Formation (Fig. 8). In addition, a 10–15 m thick shale and siltstone unit on top of the Egga Member is also included in the Våle Formation (Våle shale).

Facies and facies associations of the Ormen Lange reservoir unit

Seven process-related facies have been identified within the cored succession. Facies identified are based on classical work on turbidite facies (e.g. Bouma 1962; Walker 1967, 1978; Mutti & Ricci Lucchi 1972; Walker & Mutti 1973; Lowe 1979, 1982). For reservoir modelling purposes only two reservoir facies are applied; these are the 'C-sand' (clean sand) and the 'G-sand' (low reservoir quality).

High-density turbidity current deposits

This facies consists of massive and vaguely laminated green sandstone (up to 4 m thick), occasionally with abundant green mudstone clasts (up to 2 cm in diameter). Most of the sands of this facies are defined as 'C-sand' (Fig. 9). Deformation structures caused by water escape (e.g. dish structures and contorted bedding) are common in most of the sandstone beds. Normal grading may occur in the uppermost part of individual beds and inverse grading is relatively common in the lowermost parts. In most cases, the grain size distribution is remarkably uniform with medium, and occasionally medium- to coarse-grained sandstone dominating. Minor vertical alternations between these grain sizes are evident within individual beds. Another typical vertical arrangement of this facies is a massive or graded lower part, with a non-scoured base, passing upwards into a massive interval with green mudstone clasts. The mudstones are usually well rounded, consisting mainly of chlorite and glauconite. Thin intervals with ripple lamination and convolute lamination may occur on top of some of these massive sandstone beds. The ripple-laminated intervals are very thin (few cm) and commonly poorly developed. Coal fragments are common, especially in the upper part of the beds. Pyrite aggregates, mica and siderite are common as additional minerals. Trace fossils include *Planolites* and *Palaeophycus*. Carbonate cements occurs within a few beds of Facies A and may be restricted to thin intervals, both in the upper and lower part. The thick beds of this facies may reflect sustained, near-steady flow conditions with a gradually rising depositional flow boundary (cf. Kneller & Branney 1995). Beds of this facies are commonly amalgamated or vertically stacked, with some thickening–thinning upward trend, probably reflecting lateral migration of the main feeder system or simply increased/reduced of energy of the turbidity currents. It is also suggested that the presence of shallow channels may be responsible for this arrangement.

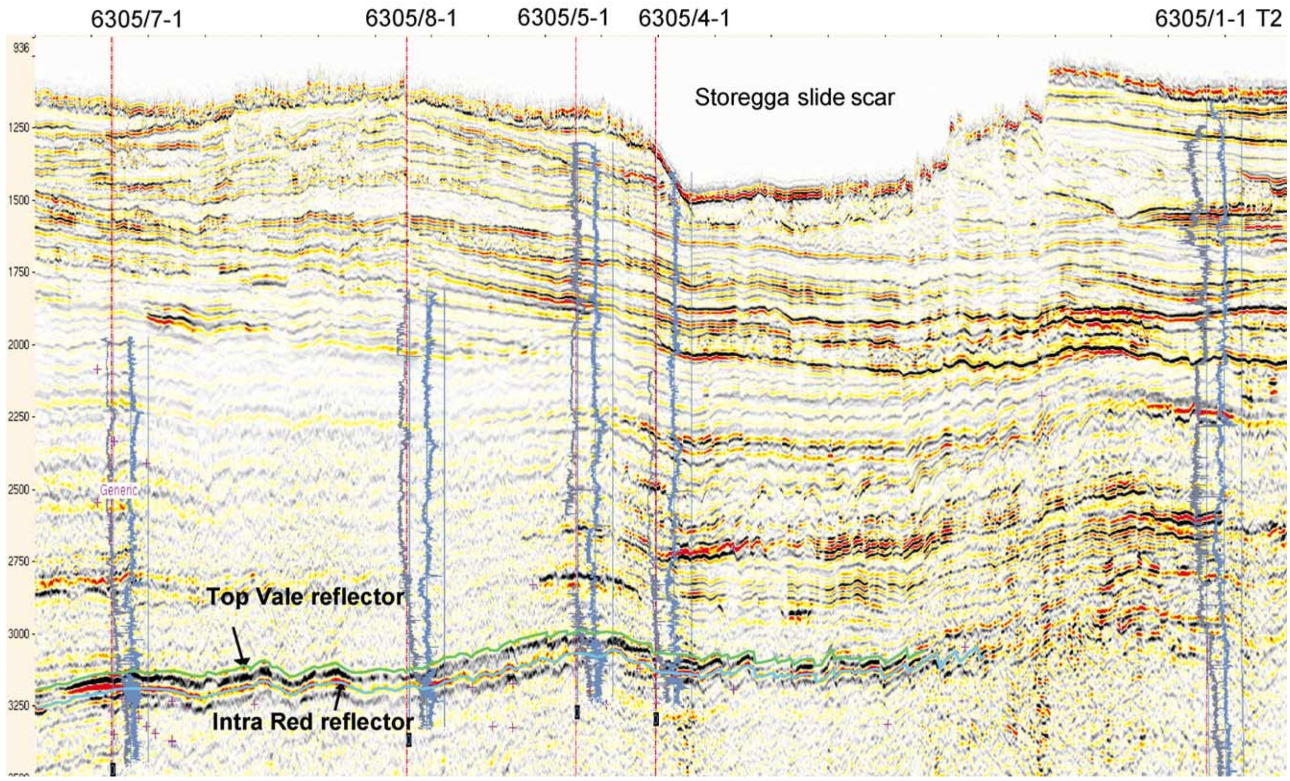

Fig. 6. Composite seismic line through the wells of the Ormen Lange Field, close to the crest of the structure. The positions of the two important seismic reflectors, 'Top Våle' and 'Intra Red' (close to top Jorsalfare Fm.), used in reservoir mapping, are shown. Light blue log curves are gamma logs and dark blue log curves represent sonic logs.

Low-density turbidity current deposits

Most of the fine- to medium-grained sandstone beds with well-developed normal grading and with a partly developed Bouma sequence (e.g. grading from a massive lower part, to parallel lamination and occasional ripple lamination at the top) are interpreted as turbidites of low density currents (Fig. 9). Sandstone in this facies is defined both as 'C-Sand' and 'G-sand' in the modelling framework. Some sandstone beds with contorted lamination have been assigned to this facies, which is relatively rare and comprises only a few beds in the cored interval of the Ormen Lange reservoir.

Highly bioturbated greenish sandy mudstone

This facies is common in the upper sand-rich part of the Egga Member, where it separates individual sandy turbidites. In a few cases sediments of this facies occur isolated within claystones without any association to underlying sand. This is particularly the case in wells 6305/4-1 and 6305/1-1 T2 where a few such isolated beds have been recorded (Fig. 9). Bioturbation mainly comprises *Planolites*, *Palaeophycus* and occasional *Rhizocorallium* trace fossils. The depositional processes of the facies are not understood and it is possible that it may have been deposited as thin sand and silt turbidites in alternation with hemipelagic shale that later was mixed by bioturbation. This facies resembles the greenish-grey mudstone facies (see below), however, it differs slightly due to its thinner and sandier development. It represents strongly deformed and bioturbated turbidite tops, often with a gradual upwards transition to hemipelagic sediments.

Green, greenish-grey and grey mudstones

This facies is very common in the Maastrichtian succession, where it is present as relatively thick beds (up to 3 m) of alternating green-grey and greenish-grey strongly bioturbated mudstones, with occasional centimetre-thick lenses of glauconite-rich sand. The mudstones consist mainly of a mixture of calcareous clay and siltstone, with minor elements of sand. Glauconite grains, pyrite aggregates and mica occur at most levels. Small shell fragments are common throughout, but in relatively small amounts. In two thin intervals there is a significant increase in carbonate, probably primarily related to increased deposition of chalk. The intensity and diversity of bioturbation is very high and, in most cases, the sediments are almost completely homogenized, with no primary sedimentary structures or lamination preserved. The most common trace fossils are *Zoophycos*, *Planolites*, *Palaeophycos* and *Helminthopsis horizontalis*. These mudstones are interpreted to be a combination of turbidite mud and hemipelagic and pelagic mudstones. The greyish and dark green mudstones are probably hemipelagic in origin, whereas the greener and coarser-grained mudstones represents mainly turbiditic mud. This interpretation accords well with the fact that the coarsest mudstones co-occur with the thickest sandstones, thus these are most likely related since the most voluminous turbidity currents probably also transported a considerable mud load. The origin of the green coloration is from glauconite and chlorite. Since glauconite forms in marine-shelf environments, incipient glauconite grains may have been transported via turbidity currents to the deep-water Ormen Lange system and possibly continued their growth there.

Pin-striped mudstones

This fine-grained lithology characterizes the stratigraphy at and immediately above the K–T boundary and basically constitutes the 'Våle Tight' unit (Fig. 9). In contrast to other mudstones within the Ormen Lange area, these mudstones are darker and contain stripes of siltstone and very fine sandstone which do not exceed 1 cm in thickness (thus, 'pin-striped mudstones'). The thin siltstone/sandstone beds show evidence of low-amplitude cross-lamination. Within the mudstones, the bioturbation intensity is reduced in relation to the other mudstones in the succession

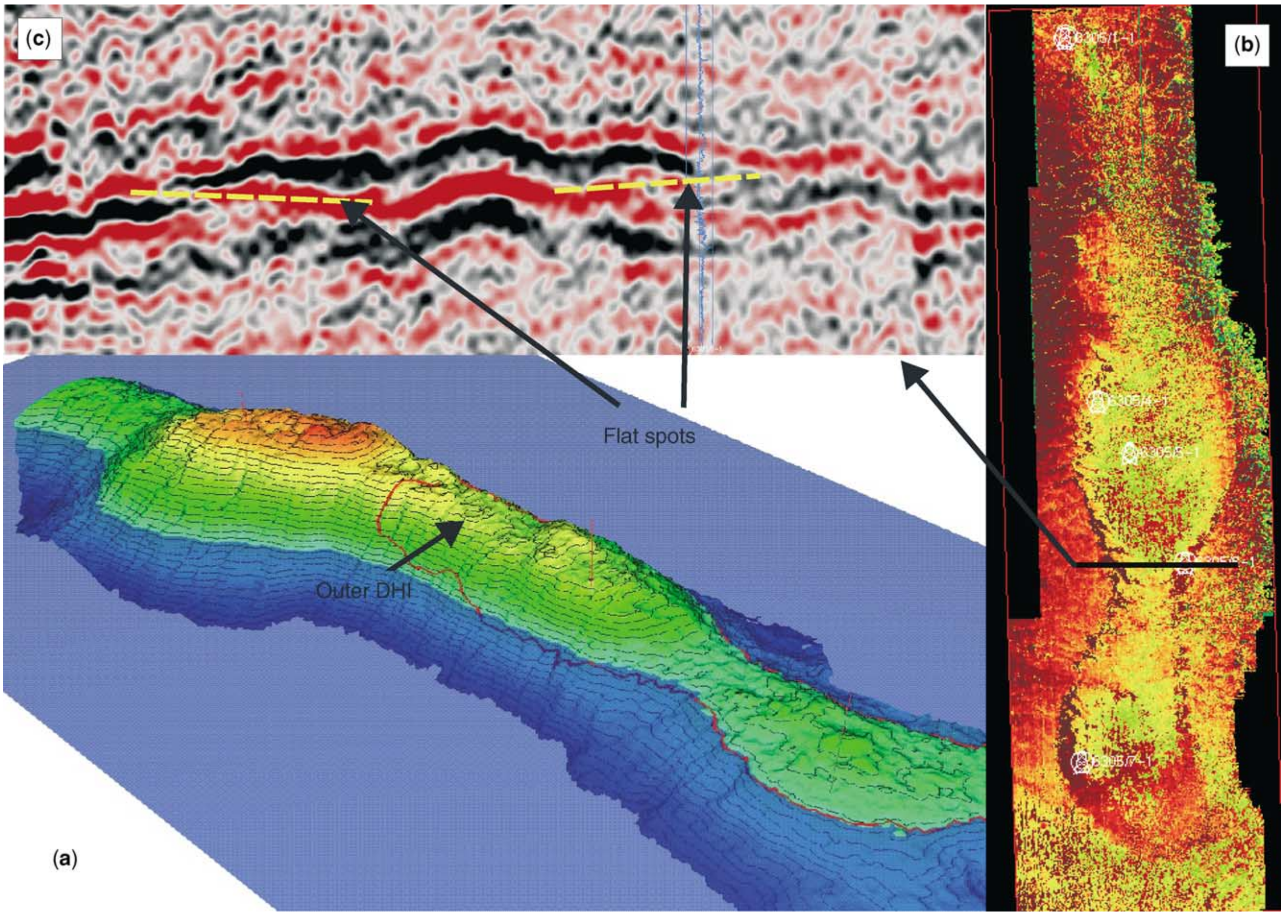

Fig. 7. The Ormen Lange Field. (**a**) Outline of the Ormen Lange structure based on Top Våle reflector level. The blue plane represents the 'regional' gas–water contact level. Red line (arrowed) represents the outer position of the DHI outline on the structure. (**b**) Seismic attribute map (volume-based frequency map from the top Våle to Intra red reflectors) showing the outline of the DHI. (**c**) East–west-orientated seismic section through the central part of the field, with strong DHI.

and a more restricted trace fossil suite occurs (Gjelberg *et al.* 2001).

In addition to the pin-striped facies, the interval just above the K–T boundary contains several relatively thick sandstone beds ('Våle Heterolithic'). The sandstone beds present just below the K–T boundary show evidence of sand injection and water escape. Only in one other part of the succession does prominent deformation occur, namely towards the top of the 6305/8-1 well. Based on biostratigraphic data, the pin-striped mudstones record increased sedimentation rates relative to underlying and overlying sediments. The pin-striped mudstones are suggested to record deposition from suspension interrupted by repeated phases of either reworking by bypassing flows or supply of low-density turbidites. The thick-bedded sandstones are interpreted as turbidites and are identical to the other sandstone beds above and below the pin-striped mudstone interval. The considerable thickness of this facies in the northwestern part of the Ormen Lange Field indicates that it was deposited over a long period of time without interruption by severe turbidity flows. In the southeastern part, on the other hand, it is strongly interbedded by thick-bedded turbidites, splitting it up into thinner units.

Chalk

This facies occurs only in one bed in the Jorsalfare/Springar Formation (Maastrichtian). The lower boundary is sharp, whereas the upper is transitional. It is strongly bioturbated (degree 5) with *Planolites*, *Zoophycos* and possibly *Taenidium satanassi* trace fossils. As mentioned above, there are a few other beds within the Maastrichtian succession that also have increased carbonate content due to chalk; however, these beds are strongly mixed with clastic mud.

Deformed sandstone/mudstone with sand injection

Some of the sandstone beds are significantly deformed and differ from the contorted bedding related to water escape structures within turbidites. One such deformed interval is located at the K–T boundary, where the top of a relatively thick sandstone bed is strongly contorted and broken up into sandstone fragments, associated with sand injections.

Carbonate-cemented intervals

Carbonate-cemented intervals occur frequently both in the Maastrichtian (Jorsalfare/Springar Formation) and, more rarely, in the Paleocene (Egga Member). They are best developed within sandstones of both high and low density turbidites, but may also occur within finer-grained lithology. The top and base of the cemented zones are usually sharp, often curved surfaces that may indicate that they are concretions or irregular beds, probably of coalescing concretions (Fig. 9).

Change of background sedimentation at the Cretaceous–Tertiary boundary

A significant change in sedimentation took place across the K–T boundary. This change is mainly seen in the fine-grained

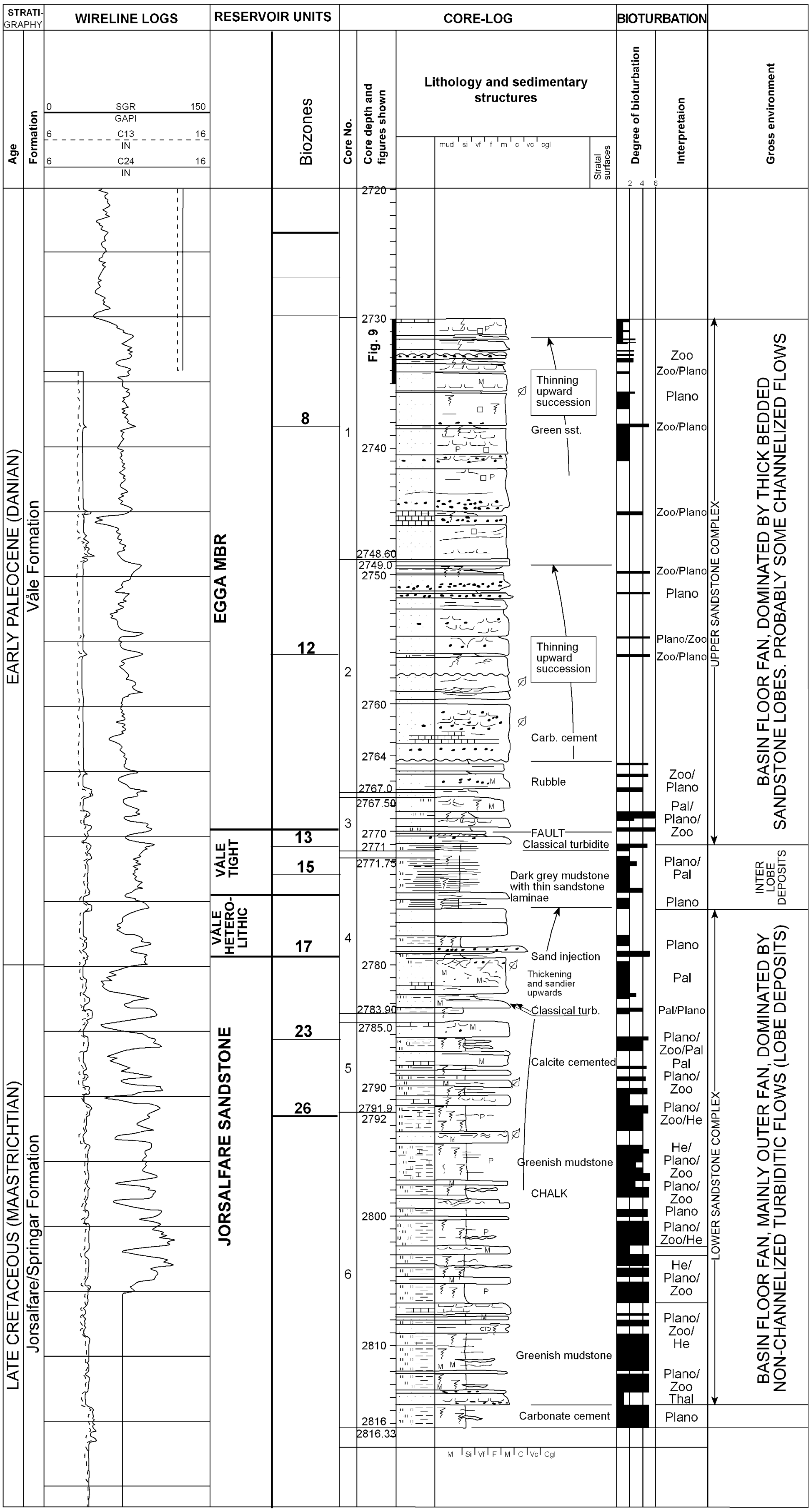

Fig. 8. Core–log of the discovery well 6305/5-1, with reservoir units and some palynozone boundaries.

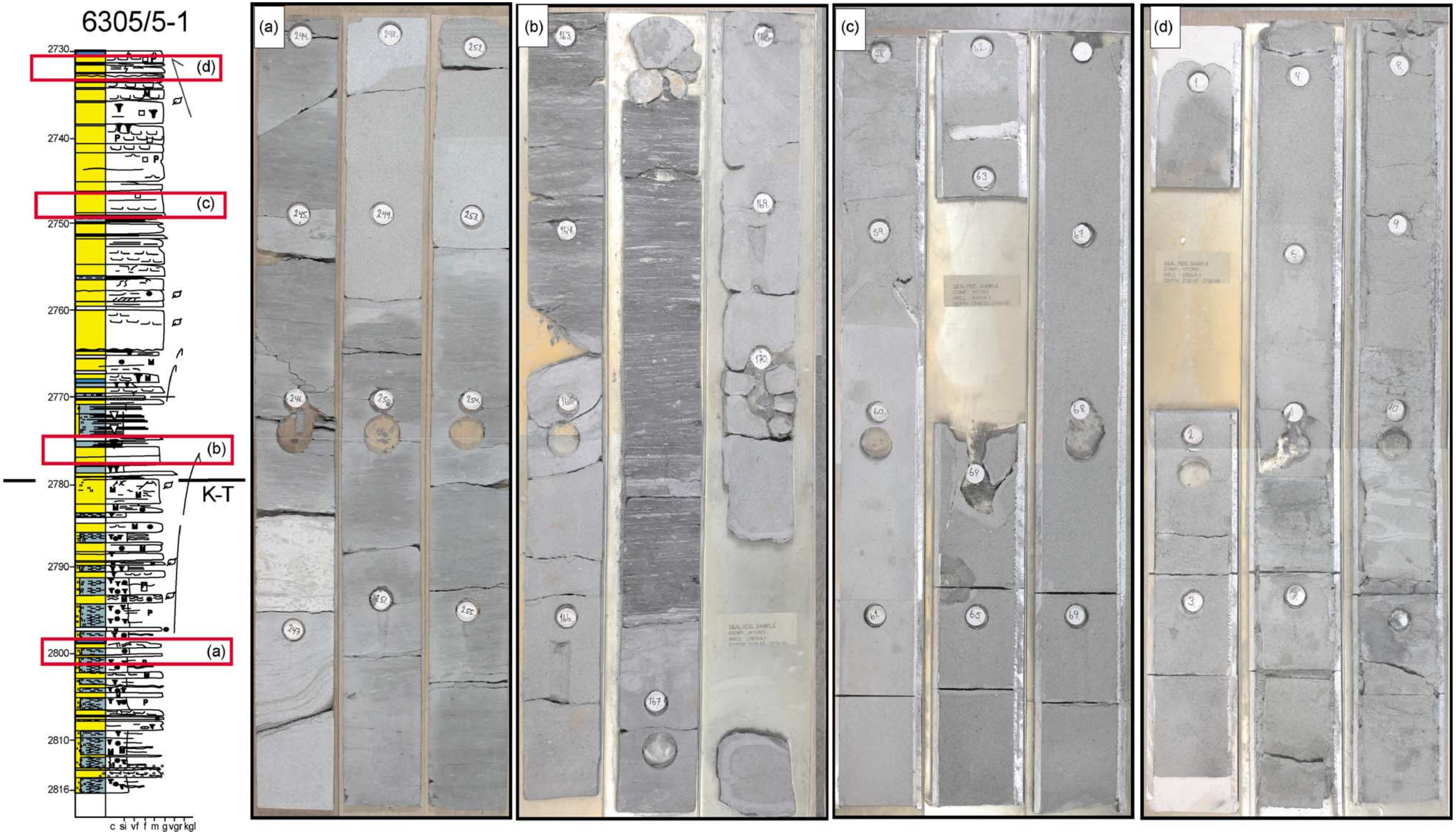

Fig. 9. Some core examples of the Ormen Lange reservoir in well 6305/5-1. (**a**) Turbidites and greenish mudstones within the Jorsalfare Fm. The white bed is the only well-developed chalk bed in the succession. (**b**) Turbidites and pin-striped facies close to the K–T boundary. (**c**) Amalgamated turbidites at the middle part of the Egga Member. (**d**) Turbidites with interbedded mudstones in the upper part of the Egga Member. Core–log to the left shows the intervals of the core photographs.

background sedimentation, which changes from greenish-grey, strongly bioturbated claystones and mudstones of Maastrichtian age into dark grey, less strongly bioturbated mudstones of early Danian age. The Maastrichtian mudstones exhibit a high diversity of trace fossils mainly of the *Zoophycus* ichnofacies together with *Helminthoida and Anconichnus*, whereas the Danian mudstones immediately above the boundary show a very low diversity assemblage (isolated *Planolites*). The sandstones of the Egga Member are all classified as subarkoses according to the classification scheme of Pettijohn *et al*. (1972), with close to 10% feldspar. Mineralogical and elemental composition of the fine-grained background sediments in the well indicate that there is a significant change across the K–T boundary. The concentrations of kaolinite, mica/illite and calcite decrease significantly upwards across the boundary, whereas smectite increases. The greenish shale is rich in chloritic material compared to the black shales, while the clay minerals glauconite and smectite predominate in the black shales. Pyrite is more common in the black shales than in the greenish shales, while rutile predominates in the latter (Gjelberg *et al*. 2001).

The change in the background sedimentation occurs slightly below the boundary, portrayed by a change from greenish, bioturbated mudstones to dark grey mudstones in the Maastrichtian succession. However, the most significant change, both with respect to lithology and ichnofabric occurs at the K–T boundary itself. Gjelberg *et al*. (2001) tentatively suggested an extraterrestrial cause, such as a bolide impact to explain the observed geochemical and sedimentological changes at the K–T boundary. Increased instability of slope sediments and giant-scale slumping have been inferred by several authors. Interestingly, Norris *et al*. (2000) documented that large-scale slumping and sliding of the eastern North Atlantic seaboard during the Cretaceous–Palaeogene transition took place, more or less simultaneously with the Chicxulub impact in the Gulf of Mexico, even though these regions were many hundreds of kilometres away. The first changes in background sedimentation at the K–T boundary (from greenish to dark grey and grey mud) seem to occur slightly below the surface itself and it is, therefore, possible that the changes are related to other external causes, such as gradual climatic change. Such changes could have caused stress on the environment making the fauna and flora more sensitive to other external influences, such as bolide impacts.

Architectural elements and grain size distribution for the Ormen Lange reservoir

The genetic lithofacies of the Ormen Lange reservoir is mainly high-density turbidites, with some intercalations of 'classical' turbidites. The architectural elements of the deposits are all related to depositional settings within a submarine fan system, where depositional lobes are the dominant elements. Based on seismic morphology (sheet geometry), core description and general sedimentological knowledge, it has been suggested that three main depositional elements are present. These are: (1) channel-dominated zone with overbank deposits; (2) channelized lobe; and (3) frontal splay/fan fringe lobe (Fig. 10). Despite it being possible to identify channels and what are believed to be lobes and overbank splays, it must be stressed that in the wells the elements probably had relatively low relief and high aspect ratios (high width to thickness ratios). Thus, the differentiation between a channel and a lobe with a scoured base may, in many instances, be semantic. Identification of axes of deposition defined by the biostratigraphy may be as far one can get with only five wells and poorly resolvable seismic data.

- **Channel zone deposits** (Fig. 10) are mainly located in the more proximal part of the system and are characterized by thick beds of amalgamated high-density turbidites arranged in a fining- and thinning-upwards pattern, associated with intervals of thinly bedded classical and fine-grained turbidites.

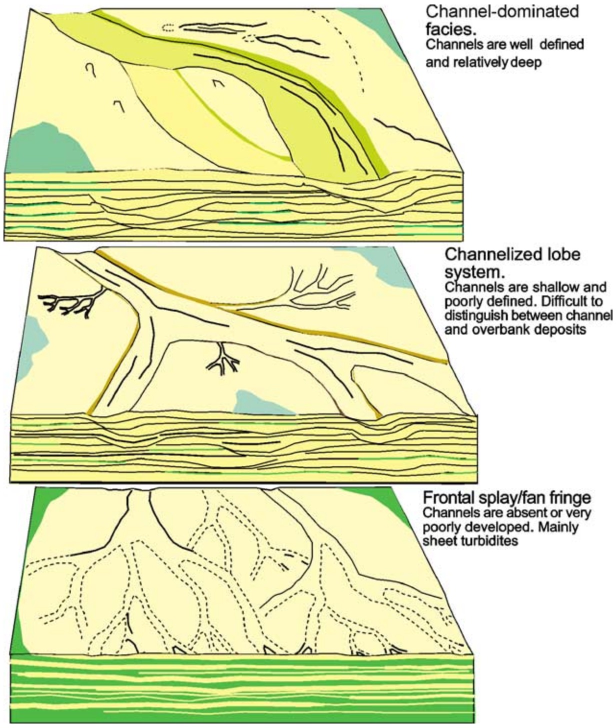

Fig. 10. Sedimentological models for different depositional elements of the Ormen Lange reservoir.

This depositional element is not common within the Ormen Lange Field, but is well developed within the sedimentary fairway to the east. It is also possible that the lower part of the Egga Member in well 6305/8-1 and some of the upper part of the member in well 6305/7-1 represents this depositional element. The element consists mainly of 'C-sand' but may also contain some 'G-sand'.

- **Channelized lobe deposits** (Fig. 10) are thought to be a relative common depositional element within the field; they consist of amalgamated high-density turbidites in alternation with thin classical' turbidites occasionally draped with a thin mudstone package. This depositional element differs very little from the previous one, the main differences being related to how well the channels are developed. The channel belt element is dominated by a relatively deep and well-defined channel or channels in a braided pattern (Fig. 10), whereas the channelized lobe element is mainly dominated by shallow branching channels and sandy overbank deposits. In such a setting it may be very difficult to distinguish between in-channel and overbank deposits, as both the shallow channels and the proximal overbank deposits are dominated by relatively thick (high density) turbidites. This facies element is common in all wells drilled so far. 'C-sand' dominates strongly within the depositional element, with only minor incursion of 'G-sand'.

  In all wells the grain size analyses reveal well-defined fining-upward trends. These trends correspond well with subjective observations on cores, where fining- and thinning-upwards units have been described from all wells (Fig. 11) and it is tentatively suggested that these may reflect infill of shallow channels.

- **Frontal splay or fan fringe deposits** (Fig. 10) are usually laterally connected to the previous facies element and may represent a lateral equivalent. The element consists of 'classical' low density and high density turbidites in alternation with thin, fine-grained turbidites and hemipelagic mudstones. 'G-sand' constitutes a major part of the facies element and it is well developed both within the Jorsalfare Formation, the 'Våle Heterolithic'/'Våle Tight' units and at the top of the Egga Member ('Egga Tight').

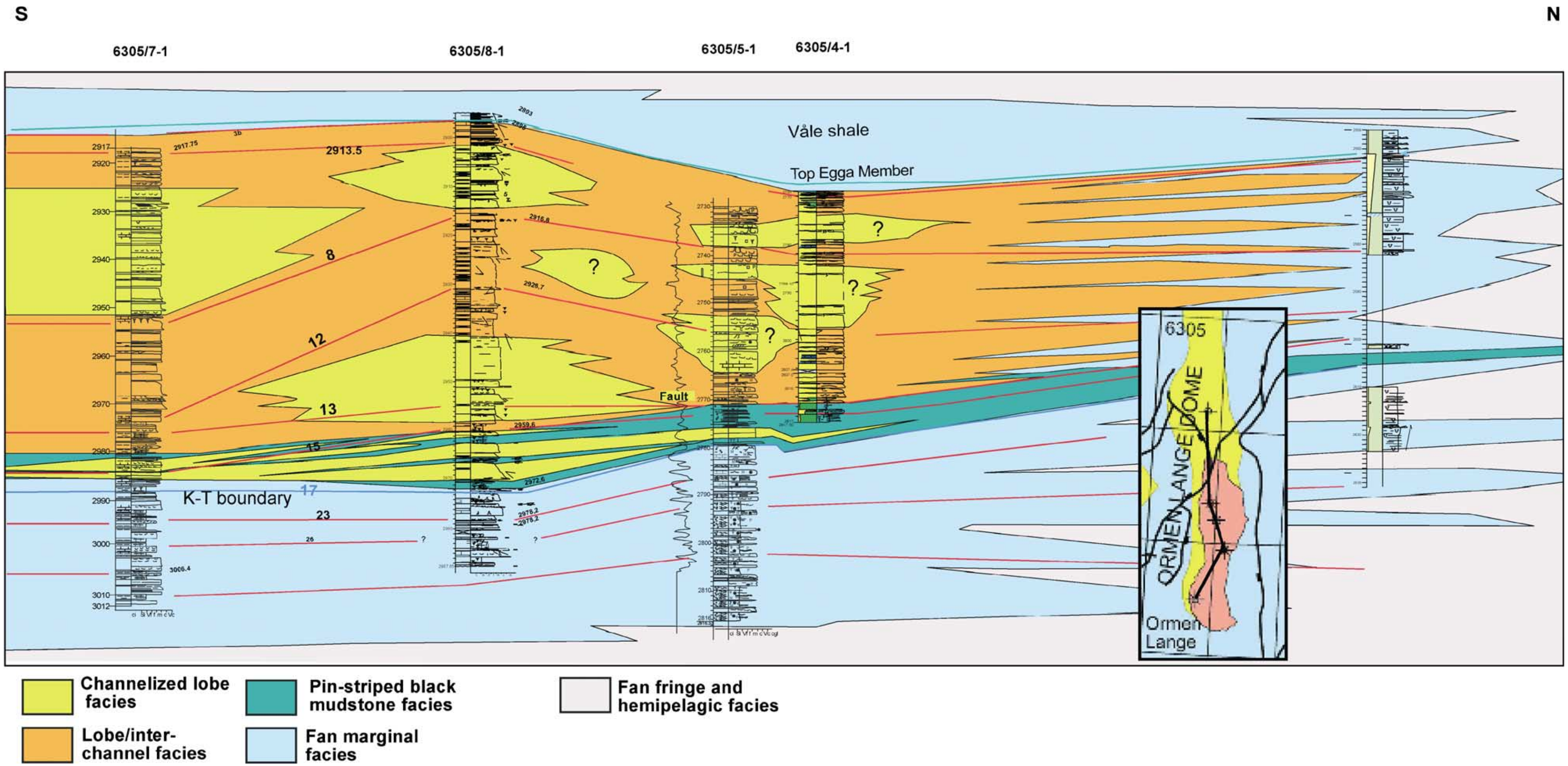

Fig. 11. Core correlation and general facies distribution between wells in the Ormen Lange Field.

A general grain size reduction from the channel-dominated facies to the frontal splay/fan fringe facies is expected. However, grain size distribution seems to be relatively uniform, as most of the samples show an average grain size in the fine to medium range. Grain size analyses (sieve and coulter) suggest that there is a faint grain size reduction between well 6305/7-1 and 6305/5-1. This is interpreted as a consequence of proximal to distal deposition on the submarine fan system, where well 6305/5-1 represents a more distal position compared to the wells 6305/7-1 and 8-1. Within the Egga Member, in well 6305/5-1, the lower part of the interval (up to 2764 m) tends to have a slightly finer grain size distribution than the upper part.

Core porosity and permeability data from wells 6305/7-1, 8-1 and 5-1 show a considerable variation both vertically and between wells. Well 6305/5-1 has porosity values that dominate between 30% and 35% for the Egga Member, but slightly less for the sands in the Jorsalfare Member. The permeability is more than 1 D throughout most of the reservoir and up to 2 D within the some intervals in the upper part. In well 6505/7-1 the porosity is usually less than 30% in the upper part of the Egga Member and slightly more than 30% in the lower part, whereas the permeability is less than 1 D in the upper part and close to 2 D in the lower part. Well 6305/8-1 has porosities between 30% and 35% in the upper part of the Egga Member, less than 30% in the middle part and more than 35% for some amalgamated beds at the base. There is a general trend revealing that the coarser-grained sediments have both higher porosity and permeability than the finer-grained turbidites.

Sedimentation in the 'Ormen Lange sub-basin'

The regional time isochore map of the Paleocene interval within the greater Ormen Lange area shows an elongate, sinuous thickness anomaly emanating from the area of the 6306/10 block running west-northwestwards to the 6305/8 block, then making a sharp northwards turn and fading out in the 6305/1-1 area (Figs 3 and 7). This isochore anomaly is believed to be a geomorphological expression of an extended deep-water turbidite system with its principal source/feeder area at the intersection between the northeastern end of the Gossa High, the Jan Mayen Lineament, the southern end of the Klakk fault complex and the southwestern margin of the Frøya High (Fig. 1). Therefore, the Ormen Lange licence area sits within the feeder system to the Ormen Lange depositional system and is located centrally within the isopach thickness anomaly.

Detailed seismic morphology of the upper Cretaceous–lower Paleocene section in the Ormen Lange licence area is difficult to interpret because of the poor seismic data quality and the strong overprint of the gas–water contact. Even though thinning and thickening trends can be seen on isopach maps (Fig. 7), internal reservoir architecture is not seen either on seismic lines nor on volume- and surface-based attribute maps (Figs 5 and 7). The attribute maps tend to highlight the distribution of gas but, in addition, they may also, to some extent, show the distribution of sand. These maps suggest that a reservoir with some gas saturation may be present as far north as well 6305/1-1, but distributed just to the east of the well.

A series of palaeoenvironmental reconstructions has been drawn to illustrate the deposition of the latest Maastrichtian and earliest Paleocene in the 'Ormen Lange sub-basin' (Fig. 13). The reconstructions are based on detailed biostratigraphical analysis (Figs 8, 11 and 12), combined with regional geological studies and knowledge. According to the classification system of Reading & Richards (1994) for turbidite systems in deep-water basin margins (classified by grain size and feeder systems), the Ormen Lange reservoir should be regarded as a sand-rich submarine fan system deposited from a point source.

Early and early late Maastrichtian (Palynozones 31–26)

This period was strongly dominated by greenish mudstones deposited as background sediments with interbedded high and low density turbidites. The turbidites occur sporadically in the vertical section with a tentatively identified thickening-upwards trend. The thickest sandy development is in the 6305/7-1 well in the southern region, where more than 20 m of sand-dominated facies are recorded. Northwards, the sand intercalations decrease and, in the 6305/1-1 T2 well, sand is absent even though the total thickness of the zone increases significantly. Bypass occurred on the basin margin (evidenced by the biostratigraphy) but it is possible that some turbidite deposition occurred in the 'Gossa sub-basin' at this time. The thickening toward the northwest may be explained by interfingering of the greater Ormen Lange system with other (fine-grained) sedimentary systems (e.g. from the north or west).

Late Maastrichtian (Palynozones 26–18)

This period was dominated by deposition of greenish mudstone separated by deposition of thin- to thick-bedded turbidite sandstones, which tend to thicken upwards. This palynozone interval shows, in general, little variation between wells 6305/7-1, 6305/8-1 and 6305/5-1, both with respect to thickness and facies development. However, it is slightly thicker (only a few metres) and sandier in wells 6305/7-1 and 6305/4-1. No turbidites are present in well 6305/1-1 T2. The bedding style suggests deposition in a basin plain fed by turbidity currents entering into the basin along the Ormen Lange fairway to the southeast (Fig. 13).

Latest Maastrichtian–earliest Paleocene (Palynozones 18–16)

This period appears to have been dominated by thicker-bedded turbidite sedimentation than the previous period. Some thickness changes occur, where the lowermost Paleocene section in well 6305/8-1 is a few metres thicker than in well 6305/7-1 (Fig. 12), but thicker-bedded turbidite sedimentation seems to have started earlier in 6305/7-1.

These zones seem to record a progression of sedimentation in the Ormen Lange fan system, probably related to progradation of the system (Fig. 13). The changes in thickness observed may be random, but emphasis is put on the thick, coarse-grained beds in 6305/8-1 well, which can be put into context with later patterns at this location (see below). Preliminary biozonation suggests that there is a considerable thickness increase from well 6305/5-1 to 6305/4-1 and that a more than 10 m thick package of amalgamated turbidites occurs in well 6305/4-1. It is tentatively suggested that this thickness increase may be local, related to syn-sedimentary faults creating local depressions on the seafloor, or due to an increased accumulation rate at the western dip-slope margin of the 'Ormen Lange sub-basin'. It has been demonstrated from field studies in Annot (SE France) that turbidites tend to accumulate with increased thickness along such slopes when the turbidity currents hit the slope obliquely (Puigdefabregas *et al.* 2004).

'Våle Tight' reservoir interval: early Paleocene (Palynozone 15)

Palynozone 15 (including the immediately preceding time) is different from any of the other time zones in that the pin-striped mudstone facies dominates. Palynozone 15 is less than 2 m thick in well 6305/7-1, but close to 12 m in well 6305/8-1. In 6305/5-1 it is about 7 m thick and close to 10 m in well 6305/4-1, whereas it is not identified in well 6305/1-1 T2. The relatively thin development of this zone in well 6305/5-1 compared to the 4-1 well may be because a small fault crosses the well path in the upper part of the zone, implying that several metres have been faulted out.

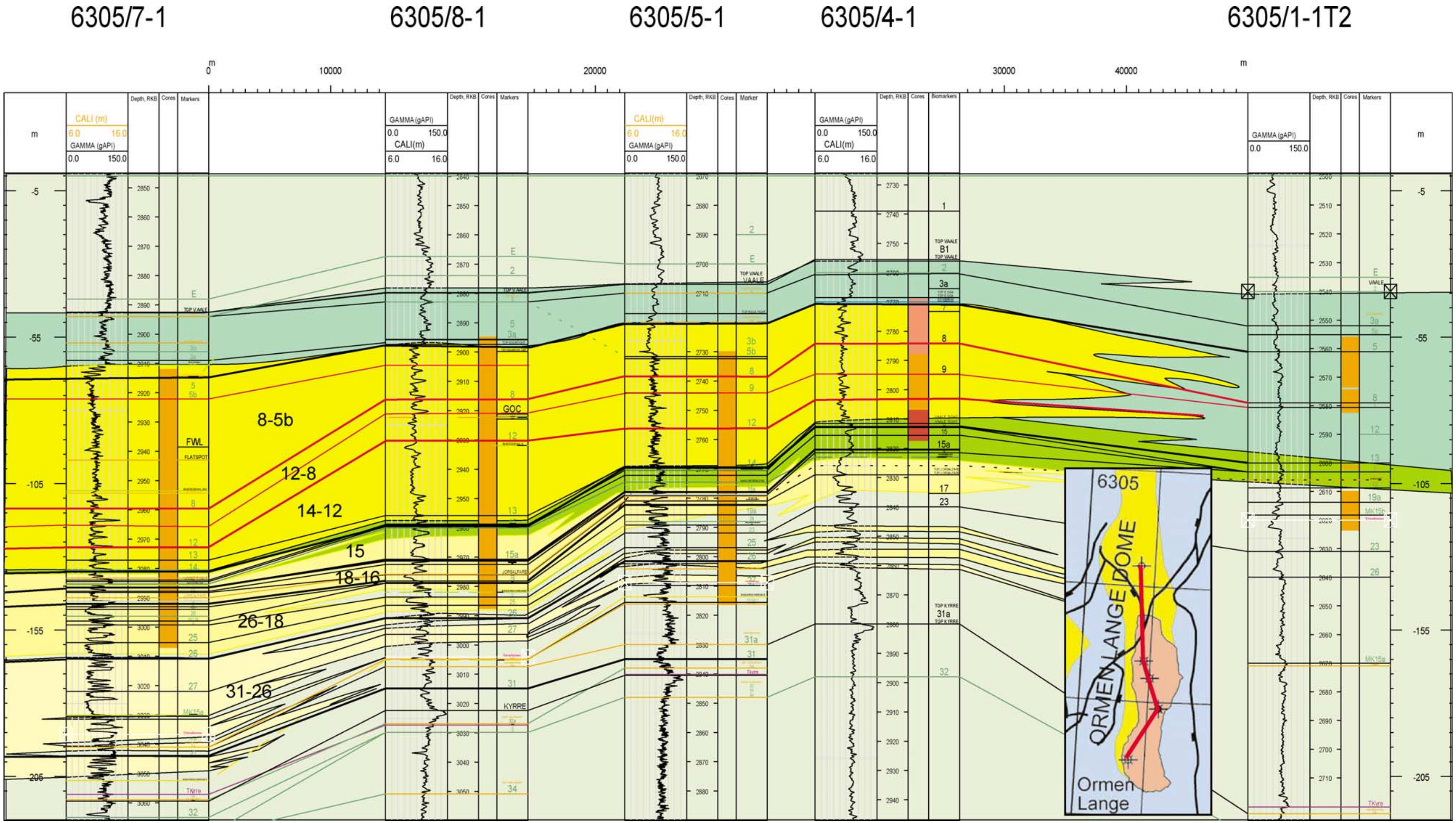

Fig. 12. Well correlation on the Ormen Lange Field with gamma-ray logs and biozonation. The zones mentioned in the text are highlighted with thick lines. Cored intervals are indicated with red bars.

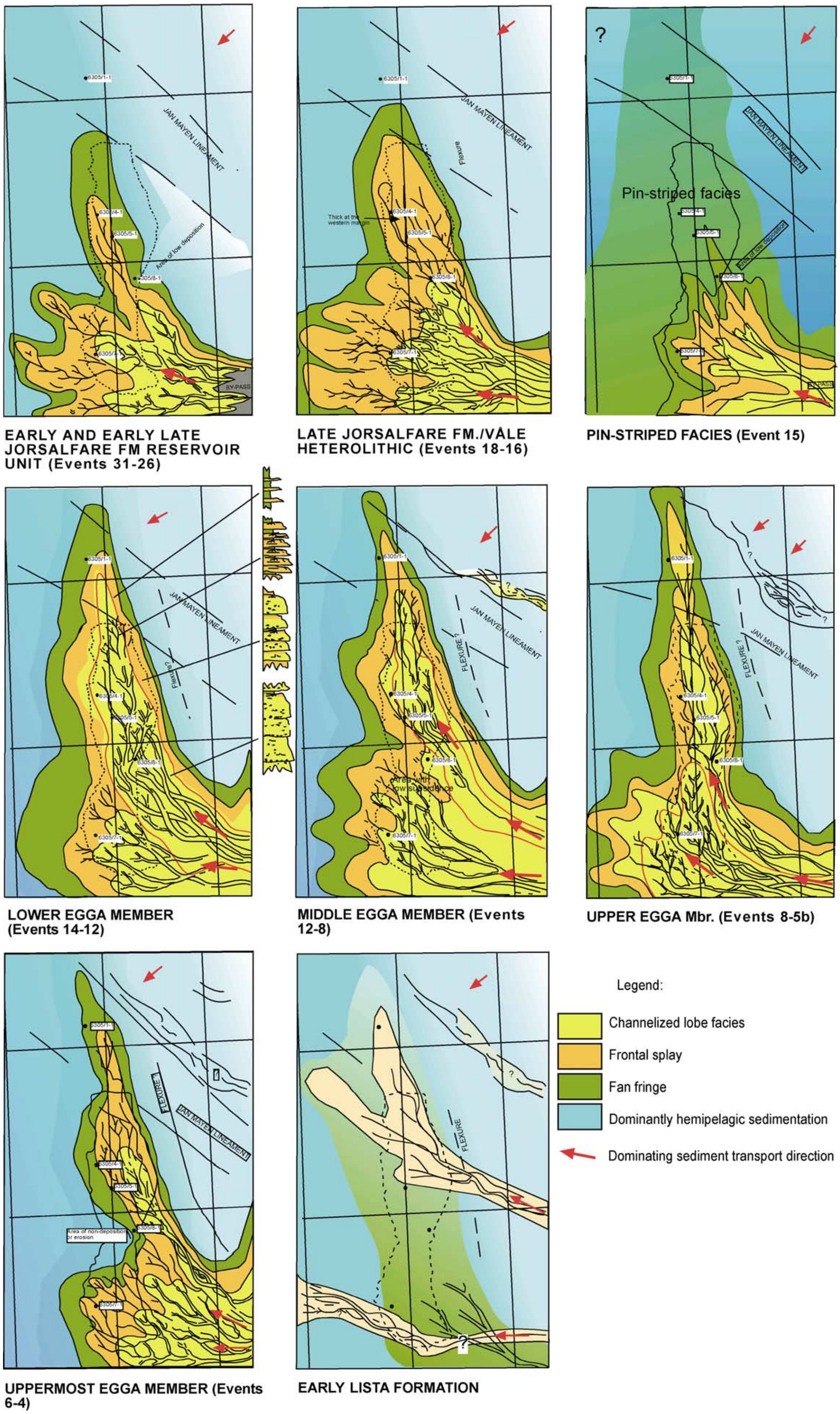

Fig. 13. Palaeoenvironmental setting during deposition of the Ormen Lange reservoir through Late Maastrichtian–Early Paleocene. See Figure 12 for event zonation.

The lower part of the palynozone includes turbidites that vary in thickness and facies. The thickest turbidites are located in the 8-1 well, where one of the beds is more than 3 m thick and where turbidites constitute up to 60% of the zone. In well 5-1 two turbidites are located in the lower part of the zone. Calculations of depositional rate of this facies based on the zonation of microfossils shows evidence of a much higher sedimentation rate during deposition of this mudstone interval than the units below.

The palynozone records a field-wide and sub-basin-wide temporary change of deposition (Gjelberg *et al.* 2001). The higher sedimentation rates and facies change suggest altered sediment supply. As alluded to above, this change may be related to slope instability during or right after the Chicxulub impact in the Gulf of Mexico. In fact, then, it is the added thickness of mudstone and the pin-stripes that are the anomalous parts of this interval, not the apparent absence of sandstones. The pin-stripes suggest a quasi-continuous supply of fine sand, perhaps as a consequence of continuous release of material on the contemporaneous slope from sliding and slumping or hyperpycnal flow. Such changes in depositional conditions could also be a consequence of climatic or relative sea-level changes.

Lower Egga Reservoir Unit, early Paleocene (Palynozones 14–12)

This time period is marked by prominent thickness changes in the wells (Fig. 12). The thickest and coarsest reservoir section occurs in well 6305/8-1, where it is approximately 36 m thick, and where the bedding pattern of amalgamated high density, relatively coarse-grained turbidites suggests the presence of channels. The interval is only 17 m thick in well 6305/7-1 and 23 and 16 m thick in wells 6305/5-1 and 4-1, respectively. The biozones in the lowermost part of this unit (Events 14–13) are very thin in well 6305/5-1, where a syn-sedimentary fault is taken as the main reason for the thinning (see above). It is suggested that the dominant coarse clastic sediment influx during this period was in the 8–1 area, entering the sub-basin through the well-established sediment fairways in the southeast (Figs 13 and 14).

Middle Egga Reservoir Unit (Palynozones 12–8)

Zones 12–8 have a more even thickness across the area, but the coarsest, thickest and most amalgamated beds occur in well 6305/5-1, where the unit is approximately 26 m thick and may contain channelized lobe/channel mouth facies (Fig. 12). The interpreted channels are mainly localized along the eastern margin of the ‘Ormen Lange sub-basin’ (Fig. 13), suggesting control by subtle differential subsidence along this margin; the underlying Jurassic fault trend was possibly the main reason for this. It is inferred that this reservoir zone extends along the eastern margin further north than the 8-1 well.

Upper Egga Reservoir Unit, early Paleocene (Palynozones 8–5b)

These zones record a major shift in sedimentation pattern where the thickest and coarsest beds now occur in well 6305/7-1, where the unit is close to 40 m thick (Fig. 11). The bedding pattern there tentatively suggests the presence of channels with relative coarse-grained, amalgamated beds. The same interval thins (individual

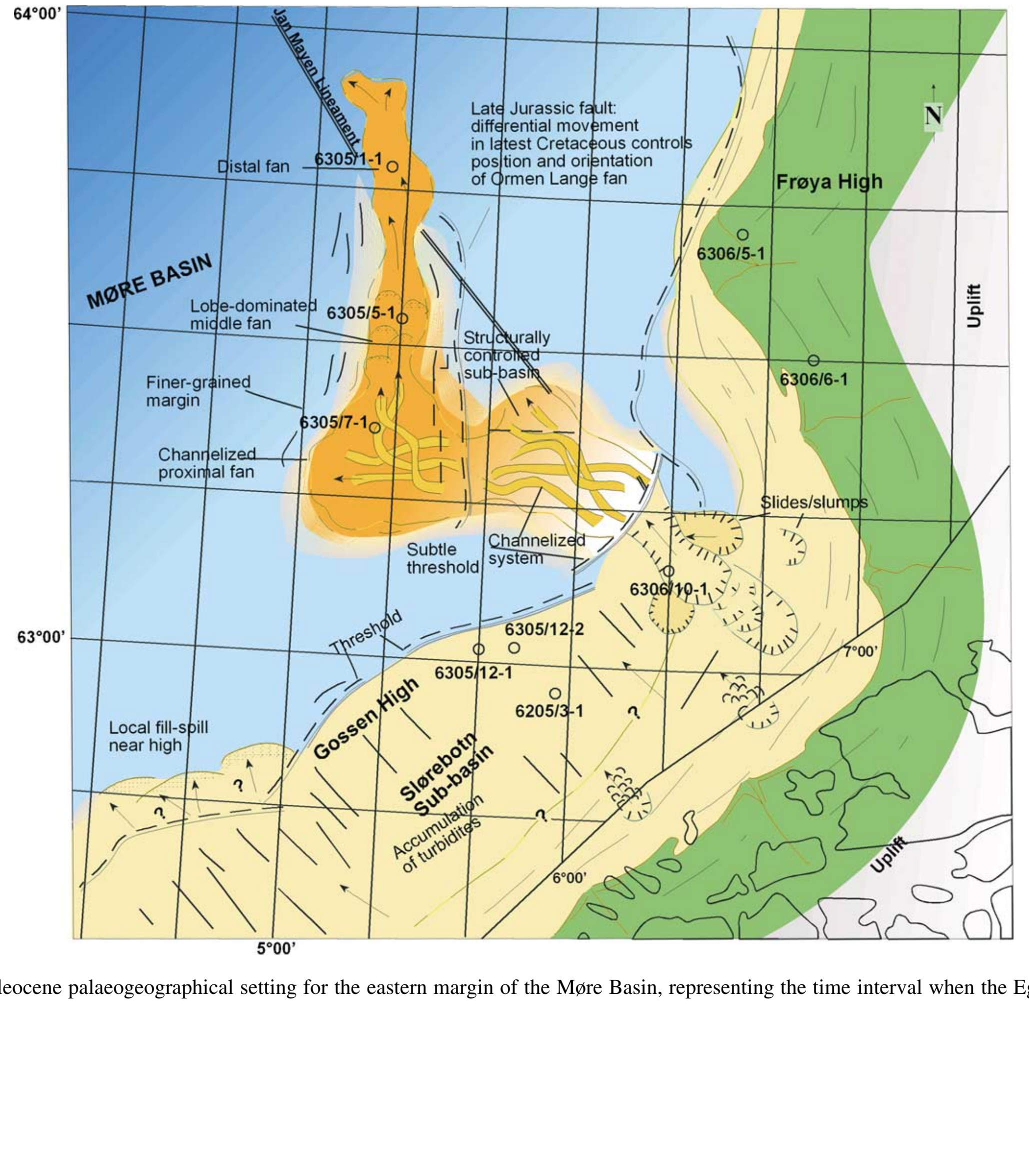

Fig. 14. Early Paleocene palaeogeographical setting for the eastern margin of the Møre Basin, representing the time interval when the Egga Member was deposited.

beds also thin and fine) towards the north, where the successions in the 8-1 and 5-1 wells are less than 20 m thick (Fig. 13).

The thickness and facies changes imply that avulsion took place upslope of the 'Ormen Lange sub-basin' and shifted the axis of channels towards the south (Fig. 13). However, this period is also the interval in which sedimentation seems to have extended furthest to the north, where the succession in the 6305/1-1 T2 well is about 27 m thick. However, it is possible that part of the fine-grained sediment supply in the north came from different sediment sources than the reservoir sand facies dominating in the southern region.

Uppermost Egga Reservoir Unit and Egga Tight (Palynozones 5b–3a)

These zones are present both in the southern and central/northern areas of the Ormen Lange Dome, but missing in the 6305/8-1 well. In well 6305/7-1 the zones consist of turbidites separated by thin mudstone interbeds. In well 6305/5-1 this interval was not cored and the log quality is poor and it is, therefore, difficult to evaluate the lithology. However, it is suggested that it consists of relatively thin turbidites separated by thin mudstones. The absence of the zone in the 8-1 well is not yet understood, but may represent an area of non-deposition and/or erosion due to tectonically induced basin floor topography at the end of deposition of the Egga Member.

Post Egga Member deposition

Following these phases of sedimentation, the Ormen Lange system because abandoned, shown by an overall thinning-upward character in wells such as 6305/8-1. During deposition of the Våle Shale the old coarse clastic systems of the Egga Member were terminated and fine-grained sediments were deposited from suspension over the whole area.

Attribute maps from the top Våle Formation suggest that two minor sediment fairway systems remained active during deposition of the overlying Lista Formation, but now crossing the structure in a west–northwest transport direction (Fig. 13). These systems may represent shallow channels with possible sediment bypass in the Ormen Lange area and deposition further west.

Subregional palaeogeography and stratigraphical evolution

The time isochore map for the Paleocene succession in the Greater Ormen Lange area shows two protrusions that are interpreted to define two palaeosubmarine feeder systems stretching from the Slørebotn Sub-basin and from the Frøya High area towards the Ormen Lange area (Fig. 3). They occur in the area of the intersection between the northeastern end of the Gossa High and the southern end of the Klakk fault complex, approximately at the landward extension of the Jan Mayen Lineament. It is suggested that this was a structural weakness zone, which attracted clastic supply because of a relatively narrow shelf landward of the area (see above). Furthermore, the transition between the Klakk fault complex and the faults controlling the Gossa High can be interpreted as a palaeo-relay ramp between two major fault systems. This interpretation is supported by the fact that the Gossa High disappears in this zone. Relay ramps are prime *loci* of sediment transport both in non-marine, shallow-marine and deep-water settings in extensional basins (Leeder & Gawthorpe 1987; Leeder & Jackson 1993).

If the deposition within the four sub-basins is genetically linked, two modes of chronological development are possible: (1) fill, or ponding and subsequent bypass to successively more basinward sub-basins, in essence similar to the minibasins of the Gulf of Mexico (Prather *et al*. 1998; Badalini *et al*. 2000); (2) initial bypass by highly effective turbidity currents which flowed to the 'Ormen Lange sub-basin', later becoming less effective and depositing their loads within progressively more landward subbasins.

Isochore data suggest that the Gossa High may have been a barrier to turbidity current flow. However, at its northeastern end, the thickness data indicate that the turbidity currents could bypass into the 'Gossa sub-basin'. Therefore, ponding seems to have been effective locally within the Slørebotn Sub-basin, but did not play a role in the development of the thickness in the 'Gossa sub-basin', in which fill is considered to be older (Fig. 3).

A complicating relationship is that only the Ormen Lange wells show a continuous section from the Maastrichtian through the Paleocene. In the wells in the Slørebotn area and on the Frøya High, most of the Maastrichtian and the lowermost Paleocene are missing. (Thin units of Maastrichtian age may be preserved locally at the Møre Basin margin as suggested from seismic data and biostratigraphy from well 6205/3-1.) This observation suggests initial bypass through the feeder system to the 'Gossa sub-basin' or erosion of most of the eventual Maastrichtian and lowermost Paleocene sediments prior to the deposition of the Egga Member. However, even though the Slørebotn Sub-basin is considered to have had negative topographic expression at that time, it is most likely that little sediment was deposited. Therefore, a lateral shift must have occurred or new feeder systems must have been activated to explain the fill of this sub-basin.

If the barriers between the sub-basins were very small, an initial bypass phase to the 'Ormen Lange sub-basin' is possible. Later, accumulation could have also been initiated in the 'Gossa sub-basin'. In this scenario, deposition of a coarse-grained lag should be expected in the 'Gossa sub-basin', which records dropping from bypassing turbidity currents. An alternative possibility is that there was no, or a laterally limited, structural barrier between the 'Ormen Lange and the Gossa sub-basins'. In this hypothesis, only a 'normal' proximal–distal relationship would exist with channels in the 'Gossa sub-basin' and more sheet-like elements in the 'Ormen Lange sub-basin'. An important observation may be that immediately west of where the prominent northwards protrusion of the isochore anomaly in the 'Gossa sub-basin' occurs (Fig. 3), there is marked thinning of the isochore. This thinning could reflect the presence of a subtle high that prevented turbidity currents from flowing directly downslope to the Ormen Lange system. Thus, where turbidity currents were deflected around the subtle high, the system narrowed significantly and flows entered the Ormen Lange system from the southeast rather than directly from the east.

Unfortunately, the seismic resolution is so poor that stratigraphical relationships within the different sub-basins (apart from the Slørebotn Sub-basin) are impossible to map. Unlike in the Gulf of Mexico sub-basins where baselap/onlap relationships are crucial for defining ponded and bypass facies assemblages (Prather *et al*. 1998), the situation in the Ormen Lange area is, perhaps, that structural relief was so small that a combination of different cases is most likely. Furthermore, the lateral variability of structural barriers and sills is important. The Gulf of Mexico model is, therefore, probably not applicable, as this model requires fully silled basins, which are filled later and bypassed.

A palaeogeographical map for the late Danian/early Paleocene (Fig. 14) illustrates the sub-regional palaeogeographical situation at the eastern margin of the Møre Basin during deposition of the reservoir units of the Ormen Lange Field.

Conclusions

Seismic data from the Ormen Lange structure are of insufficient resolution to yield information about sedimentary facies, depositional morphology and distribution of reservoir units. Regional studies and well data are, therefore, crucial to the interpretation of the stratigraphical development and distribution of sedimentary facies in the reservoir. Regional studies show that the reservoir is located at the transition zone between the wide shelf of the Halten Terrace/Trøndelag Platform area and the narrow shelf of the Møre Basin. This zone is defined by the Jan Mayen Fracture Zone and its extension towards the Norwegian mainland. The lineament probably represents a polarity change in the initial asymmetric rift basins between Norway and Greenland, with a change from upper to lower plate across the lineament. The lineament had a very important influence on sediment distribution in the area. Sand was probably focused into the area through NE–SW-orientated valley systems that developed along the important Møre–Trøndelag fault zone.

The reservoir rocks consist entirely of turbidites, but comprise architectural elements that range from channel-dominated facies, channelized lobes to frontal splay/fan fringe facies. The reservoir is best developed in the southern part of the structure, where channel facies are common, but tends to become more fine grained northwards. The reservoir tends to become thinner and probably finer grained to the west and northeast of the structure. However, it is suggested from examination of attribute maps that a sandy facies with relatively good reservoir properties is present all the way north to well 6305/1-1, but is mainly developed just east of this well. The spatial distribution of these facies elements indicates a dynamic system with changing sediment focus through time, with a compensational infill of seafloor palaeotopography,

Polygonal faults intersect the reservoir into compartments, probably with some reduced flow capacity across some of the faults. Clay smear of phyllosilicates seems to be responsible for the fault-sealing mechanism (cf Fisher & Knipe 1998). Some of the faults were active during deposition of the reservoir.

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References

Badalini, G., Kneller, B. & Winker, C. D. 2000. Architecture and processes in the Late Pleistocene Brazos-Trinity turbidite system, Gulf of Mexico continental slope. *In*: Weimer, P., Slatt, R. M., Coleman, J., Rosen, N. C., Nelson, H., Bouma, A. H., Styzen, M. J. & Lawrence, D. T. (eds) *Deep Water Reservoirs of the World*, Proceedings of the 20th Annual Research Conference, GCSSEPM Research Foundation, 16–34.

Blystad, P., Brekke, H., Færseth, R. B., Larsen, B. T., Skogseid, J. & Tørudbakken, B. 1995. *Structural elements of the Norwegian continental shelf, Part 2. The Norwegian Sea Region*, Norwegian Petroleum Directorate Bulletin, **8**.

Bouma, A. H. 1962. *Sedimentology of some Flysh Deposits*. Elsevier, Amsterdam.

Brekke, H. 2000. The tectonic evolution of the Norwegian Sea Continental Margin with emphasis on the Vøring and Møre Basins. *In*: Nøttvedt, A. (ed.) *Dynamics of the Norwegian Margin*. Geological Society, London, Special Publications, **167**, 327–378.

Brekke, H., Sjulstad, H. I., Magnus, C. & Williams, R. 2001. Sedimentary environments offshore Norway – an overview. *In*: Martinsen, O. J. & Dreyer, T. (eds) *Sedimentary Environments Offshore Norway – Palaeozoic to Recent*. Elsevier, Amsterdam, Norwegian Petroleum Society, Special Publication, **10**, 7–33.

Dalland, A., Worsley, D. & Ofstad, K. 1988. *A lithostratigraphic scheme for the Mesozoic and Cenozoic succession offshore mid- and northern Norway*, Norwegian Petroleum Society Bulletin, **4**.

Doré, G., Lundin, E. R., Birkeland, Ø., Eliassen, P. E. & Jensen, L. N. 1997. The NE Atlantic Margin: implications of late Mesozoic and Cenozoic events for hydrocarbon prospectivity. *Petroleum Geoscience*, **3**, 117–131.

Fisher, Q. J. & Knipe, R. J. 1998. Fault sealing processes in siliciclastic sediments. *In*: Jones, G., Fisher, Q. J. & Knipe, R. J. (eds) *Faulting, Fault Sealing and Fluid Flow in Hydrocarbon Reservoirs*. Geological Society, London, Special Publications, **147**, 00–01.

Gjelberg, J., Enoksen, T., Kjærnes, P., Mangerud, G., Martinsen, O. J., Roe, E. & Vågnes, E. 2001. The Maastrichtian and Danian depositional setting along the eastern margin of the Møre Basin (Mid Norwegian shelf): implications for the reservoir development of the Ormen Lange Field. *In*: Martinsen, O. J. & Dreyer, T. (eds) *Sedimentary Environments Offshore Norway – Palaeozoic to Recent*. Elsevier, Amsterdam, Norwegian Petroleum Society, Special Publication, **10**, 421–440.

Grunnaleite, I. & Gabrielsen, R. H. 1995. Structure of the Møre Basin, mid-Norway continental margin. *Tectonophysics*, **252**, 221–251.

Isaksen, D. & Tonstad, K. 1989. *A revised Cretaceous and Tertiary lithostratigraphic nomenclature for the Norwegian North Sea*, Norwegian Petroleum Society Bulletin, **5**.

Jongepier, K., Rui, J. C. & Grue, K. 1996. Triassic to Early Cretaceous stratigraphic and structural development of the northeastern Møre Basin, off mid-Norway. *Norsk Geologisk Tidsskrift*, **76**, 199–214.

Kneller, B. C. & Branney, M. 1995. Sustained high-density turbidity currents and the development of thick massive sands. *Sedimentology*, **42**, 607–616.

Leeder, M. R., Gawthorpe, R. L. 1987. Sedimentary models for extensional tilt-block/half-graben basins. *In*: Coward, M. P., Dewey, J. F. & Hancock, P. L. (eds) *Continental Extensional Tectonics*. Geological Society, London, Special Publications, **28**, 139–152.

Leeder, M. R. & Jackson, J. A. 1993. The interaction between normal faulting and drainage in active extensional basins, with examples from the western United States and central Greece. *Basin Research*, **5**, 79–102.

Lowe, D. 1979. Sediment gravity flows: their classification and some problems of application to natural flows and deposits. *In*: Doyle, L. J. & Pilkey, O. H. (eds) *Geology of Continental Slopes*. Society of Economic Paleontologists and Mineralogists, Special Publication, **27**, 75–82.

Lowe, D. 1982. Sediment gravity flows: II Depositional models with special reference to the deposits of high-density turbidity currents. *Journal of Sedimentary Petrology*, **52**, 279–297.

Martinsen, O. J. & Dreyer, T. 2001. (eds) *Sedimentary Environments Offshore Norway – Palaeozoic to Recent*. Elsevier, Amsterdam, Norwegian Petroleum Society, Special Publication, **10**.

Martinsen, O. J. & Gjelberg, J. 2001. Paleocene deep-water depositional systems along the eastern margin of the frontier Møre and Vøring basins, Mid-Norway. *AAPG Annual Convention, Denver, Colorado, Program with Abstracts*. A126.

Martinsen, O. J., Bøen, F., Charnock, M., Mangerud, G. & Nøttvedt, A. 1999. Cenozoic development of the Norwegian margin: sequences and sedimentary response to variable basin physiography and tectonic setting. *In*: Fleet, A. J. & Boldy, S. A. R. (eds) *Petroleum Geology of Northwest Europe: Proceedings of the 5th Conference*. Geological Society, London, 293–304.

Mutti, E. & Ricci Lucchi, R. 1972. Le tobiditi delt Apennino settentrionale: introduzione all'analisi di facies. *Memorie Società Geologica Italiana*, **11**, 161–199.

Norris, R. D., Firth, J., Blusztajn, J. S. & Ravizza, G. 2000. Mass failure of the North Atlantic margin triggered by the Cretaceous–Paleogene bolide impact. *Geology*, **28**, 1119–1122.

Pettijohn, F. J., Potter, P. E. & Siever, R. 1972. *Sand and sandstone*. Springer, Berlin.

Prather, B. E., Booth, J. R., Steffens, G. S. & Craig, P. A. 1998. Classification, lithologic calibration, and stratigraphic succession of seismic facies of intraslope basins, deep-water Gulf of Mexico. *AAPG Bulletin*, **82**, 701–728.

Puigdefabregas, C., Gjelberg, J. & Vaksdal, M. 2004. The Annot Sandstone in the Annot syncline: basin-margin onlap and associated soft-sediment deformation. *In*: Joseph, P. & Lomas, S. A. (eds) *Deep-water Sedimentation in the Alpine Basin of SE France: New Perspectives of the Grès d'Annot and Related Systems*. Geological Society, London, Special Publications, **221**, 367–387.

Reading, H. G. & Richards, M. 1994. Turbidite systems in deep-water basin margins classified by grain size and feeder systems. *AAPG Bulletin*, **78**, 792–822.

Riis, F. 1996. Quantification of Cenozoic vertical movements of Scandinavia by correlation of morphological surfaces with offshore data. *Global and Planetary Changes*, **12**, 331–357.

Riis, F. & Fjeldskaar, W. 1992. On the magnitude of the late Tertiary and Quaternary erosion and its significance for the uplift of Scandinavia and Barents Sea. *In*: Larsen, R. M., Brekke, H., Larsen, B. T. & Talleraas, E. (eds) *Structural and Tectonic Modelling and its Application to Petroleum Geology*, Norwegian Petroleum Society, Special Publication, **1**, 163–185.

Smith, R. & Møller, N. 2004. Sedimentology and Reservoir Modelling of the Giant Ormen Lange Field, Mid Norway. *European Association of Geoscientists and Engineers Annual Conference and Technical Exhibition, Firenze 2002, Abstract volume*.

Vergara, L., Wreglesworth, I., Trayfoot, M. & Richardsen, G. 2001. The distribution of Cretaceous and Paleocene deep-water reservoirs in the Norwegian Sea basin. *Petroleum Geoscience*, **7**, 395–408.

Vågnes, E., Gabrielsen, R. H. & Haremo, P. 1998. Late Cretaceous–Cenozoic intraplate contractional deformation at the Norwegian continental shelf: timing, magnitude and regional implications. *Tectonophysics*, **300**, 29–46.

Walker, R. G. 1976. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *Journal of Sedimentary Petrology*, **37**, 25–43.

Walker, R. G. 1978. Deep-water sandstone facies and ancient submarine fans: models for exploration for stratigraphic trans. *AAPG Bulletin*, **62**, 932–966.

Walker, R. G. & Mutti, E. 1973. Turbidite facie and facies associations. *In*: Middleton, G. V. & Bouma, A. H. (eds) *Turbidites and Deep Water Sedimentation, Pacific Section*, Society of Economic Paleontologists and Mineralogists, Short Course Notes, 119–157.