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The alluvial architecture of the Coevorden Field (Upper Carboniferous), the Netherlands

H. Kombrink[1,*], J.S. Bridge[2] & E. Stouthamer[1]

1 Department of Physical Geography, Faculty of Geosciences, Utrecht University, P.O. Box 80.115, 3508 TC Utrecht, the Netherlands.

2 Department of Geological Sciences, Binghamton University, Binghamton, NY13902-6000, USA.

* corresponding author. Present address: Department of Earth Sciences, Faculty of Geosciences, Utrecht University, P.O. Box 80115, 3508 TC Utrecht, the Netherlands. Email: h.kombrink@geo.uu.nl

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Abstract

A detailed reconstruction of the alluvial architecture of the Coevorden gas Field (Tubbergen Formation, Upper Carboniferous), which is located in the northeastern part of the Netherlands, is presented. This reconstruction is based on well logs, cross-sections and paleogeographic maps. Sedimentological analysis of a 93 m long core allowed to refine the interpretation of the depositional environment. Accurate width determinations are necessary to correctly correlate fluvial sandbodies and reconstruct alluvial architecture. Without using sedimentological information, sandbody width is likely to be overestimated. A method developed by Bridge and Tye (2000) was used to calculate the width of one sandstone body from cross-set thicknesses. On the basis of this calculation and the paleogeographic reconstructions, it may be stated that on average the width of the channel belts we studied in the Coevorden field does not exceed 4 km. Moreover, our paleogeographic reconstructions, which point to a northwestern direction of paleoflow, are in accordance with earlier observations from the study area. The Tubbergen Formation and time-equivalent sediments in Germany are reviewed briefly to put the Coevorden Field in a regional context. The thickness of the Tubbergen Formation is ~450 m in our study area. In the adjacent German area, time-equivalent sedimentary sequences reach higher thicknesses. This may be attributed to tectonic movements along the Gronau Fault zone and the coming into existence of the Ems Low, of which the Coevorden Field is the westernmost part.

Keywords: Coevorden Field, Tubbergen Formation, Upper Carboniferous, alluvial architecture.

Introduction

The Coevorden gas Field (Westphalian C/D) is located in the northeastern part of the Netherlands and covers an area of 125 km^2 (Fig. 1). The Westphalian C and D sedimentary sequence in the northeastern part of the Netherlands and adjacent areas is characterized by fluvial-lacustrine deposits (Pagnier & Van Tongeren, 1996). Sediments of this age have been extensively described in the Ibbenbüren and Osnabrück areas (David, 1987, 1990; Selter, 1990; Jankowski et al., 1993; Jones & Glover, 2005) and partly in the Ruhr Basin (Drozdzewski, 2005). In the eastern part of the Netherlands these sediments belong to the Tubbergen Formation (Van Adrichem Boogaert & Kouwe, 1993). Although the Tubbergen Formation has been extensively studied in the Netherlands as a consequence of the presence of hydrocarbons, most of these studies are unpublished. Besides a regional study (Pagnier & van Tongeren, 1996), no detailed studies on fieldscale have been published. The Coevorden Field provides a unique location to study this formation. We studied about 400 m of the Tubbergen Formation, of which sandstone percentage is approximately 50%. It is underlain by the coaly, fine-grained Maurits Formation and disconformably overlain by the Permian (Zechstein Group) (Van Adrichem Boogaert & Kouwe, 1993). Because this field is closely located to extensively studied outcrops and wells in Germany, the results can well be compared with the findings from these areas.

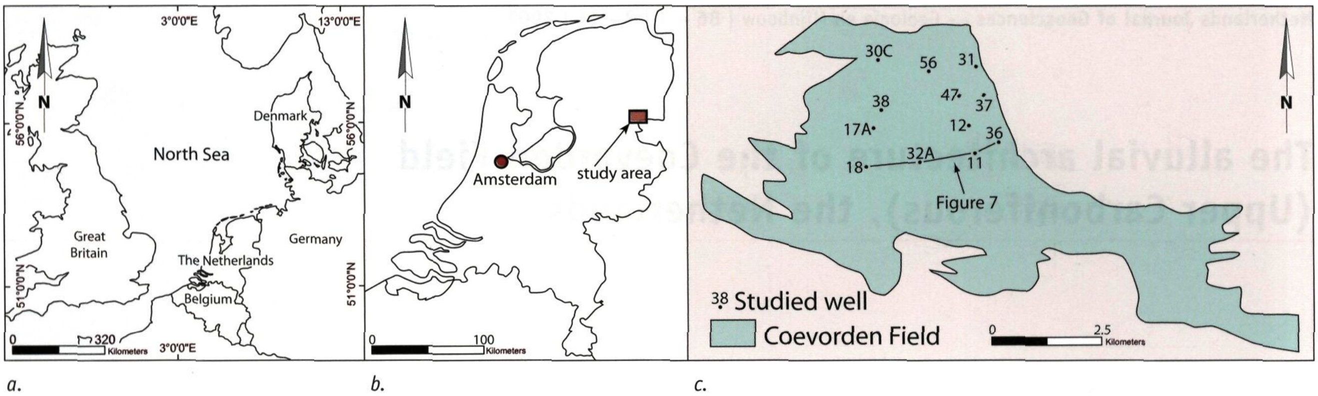

Fig. 1. Location of the Southern North Sea region (a), the Netherlands (b), the Coevorden Field, the studied wells, and correlation panel (c).

The objective of this paper is to describe the alluvial architecture of the Coevorden Field, using the following methodology:

1. interpreting the origin of depositional environments from cores and well logs;
2. correlating channel sandstone bodies between wells;
3. reconstructing the orientation and spatial distribution of individual channel sandstone bodies; and
4. putting the Coevorden Field in a regional context.

Geological setting

The Tubbergen Formation in the northeastern part of the Netherlands

Sediments of the Tubbergen Formation (Early Westphalian C – Late Westphalian D, Fig. 2) occur in the (north)eastern part of the Netherlands, the Ems Low and the Lauwerszee Trough (Fig. 3). Because of the diachronous character of the base and top of the Tubbergen Formation, it is thought to be of Late Westphalian C and Early Westphalian D respectively in the Coevorden area (Van Adrichem Boogaert & Kouwe, 1993).

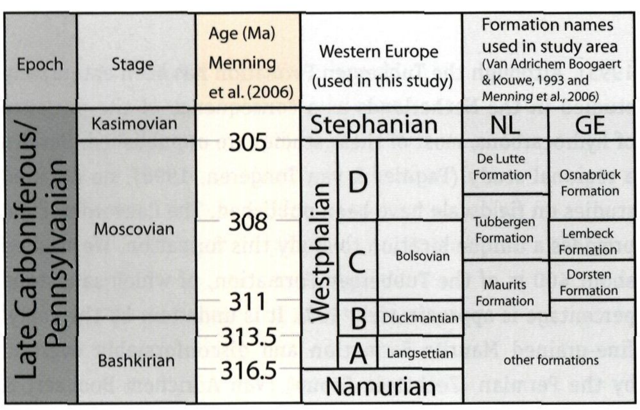

Fig. 2. Diagram showing the global and local chronostratigraphy used for the studied succession, as well as the formation names used in the Netherlands (NL) and Germany (GE). Note that time is not to scale.

The Tubbergen Formation is characterized by a fluvial-lacustrine depositional environment (Van Adrichem Boogaert & Kouwe, 1993; Pagnier & van Tongeren, 1996; Quirk & Aitken, 1997). Thick and massive sandstones occur (several tens of metres in thickness). They are intercalated in predominantly grey coloured mudstones, which reflect poorly drained conditions. A limited number of coal seams occurs in the Tubbergen Formation. The maximum thickness of the Tubbergen Formation is approximately 500 m. Sandstone percentages range from 30 to 70% (Van Adrichem Boogaert & Kouwe, 1993). The Tubbergen Formation overlies the Maurits Formation, which is a succession of grey mudstones with numerous coal seams and only a minor amount of sandstones. The boundary between the Tubbergen and Maurits Formations is thus characterized by the appearance of thick, massive sandstones and a decrease in the number of coal seams. On top of the Tubbergen Formation the De Lutte Formation occurs, which is characterized by a succession of predominantly reddish-brown sandy mudstones, reflecting well oxidized depositional conditions. Time equivalent sediments have been shown to be present in the Campine Basin in the southern part of the Netherlands and the Cleaverbank High (Van Adrichem Boogaert & Kouwe, 1993, Fig. 3). These sediments suggest a similar depositional environment.

Sediments of Westphalian C and D age in northwestern Germany

The Westphalian C (WC) and Westphalian D (WD) are well studied in the Ruhr Basin (WC only) and the Ibbenbüren and Osnabrück areas (David, 1987; David, 1990; Selter, 1990; Jankowski et al., 1993; Drozdzewski, 2005; Jones & Glover, 2005). Like in the Netherlands, sediments deposited during this time mainly reflect a fluvial dominated environment. David (1990) recognized various depositional environments: braided rivers, overbank siltstones, swamps (coals) and crevasse splays. The transition from coal bearing and sandstone-poor to sandstone dominated as observed in the Coevorden area (Maurits Formation to Tubbergen Formation) during Late WC occurred during the late Early Westphalian B in the Ruhr and

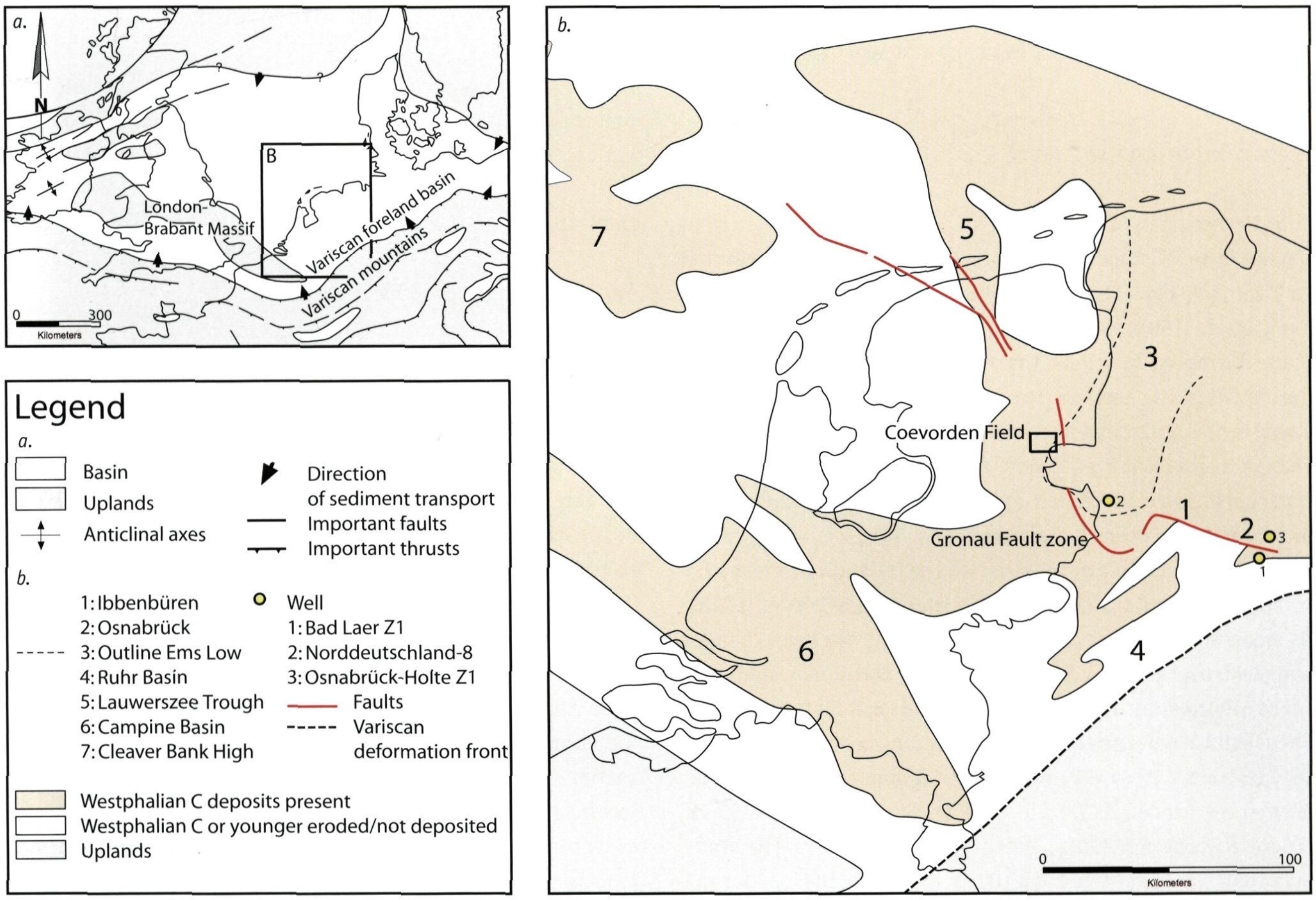

Fig. 3. Location of the Netherlands within the northwest European Carboniferous Basin (A) and the presence of Westphalian C deposits in the Netherlands and adjacent areas in Germany (B). Fig. 3a is based on Ziegler (1990).

Ibbenbüren areas (Drozdzewski, 2005). The entire WC succession is thought to be approximately 850 m thick in the Ibbenbüren and Osnabrück areas (David, 1990 *in:* Jones & Glover, 2005), while the WD is approximately 750 m in thickness. The thickness of the Upper WC (Lembeck Formation) is 465 - 525 m in the Ibbenbüren area (Drozdzewski, 2005). Percentages of sand in the WC vary between 60% in the Ibbenbüren area and 40 - 50% in the Ruhr area (Drozdzewski, 2005). The percentage of sand in the well Bad Laer Z1 is 70% for the WC, while it is 56% for the well Osnabruck-Holte Z1 (Fig., 3B) (Drozdzewski, 2005). For the WD, sand-percentages are approximately 80% (David, 1987). In general, sandstone proportions tend to decrease westwards and northwestwards towards the Ems area (Jones and Glover, 2005). The sandstones encountered in the Osnabrück area (Early WD) show a northward flowing channel system (David, 1990). Jones and Glover (2005) found paleocurrent directions towards west-northwest in the same area. This leads to the assumption of a northwestward direction of paleoflow. David (1990) recognized brackish as well as fresh water environments while Jones and Glover (2005) could not find evidence for brackish water conditions.

Tectonic setting

The tectonic setting of the WC and WD in the Netherlands and adjacent areas must be seen in the light of the establishment of a compressional regime which caused reactivation of pre-existing faults, folding and basin inversion as a consequence of the continuing northward progradation of Gondwana with respect to Laurussia (Leeder & McMahon, 1988; Fraser & Gawthorpe, 1990; Coward, 1993; Süss et al., 2001; Schroot & De Haan, 2003). The transition from the sandstone poor Maurits Formation to the sandstone rich Tubbergen Formation is attributed to this increase in tectonic activity during the WC (Pagnier & Van Tongeren, 1996). Unfortunately, seismic evidence for syndepositional deformation is scarce in the Netherlands (Geluk et al., 2002; Süss et al., 2001; Schroot & De Haan, 2003), although Pagnier & Van Tongeren (1996) were able to identify tectonic activity of the Osning-Gronau fault zone during the WC and WD. In the Ibbenbüren area, Drozdzewski (1985) identified Late Variscan tectonic activity. The area north of the Osning-Gronau fault zone has subsequently been interpreted as a Late Variscan pull-apart basin by Drozdzewski and Wrede (1994). In summary, although there is little direct evidence of syn-sedimentary faulting during the WC and WD in the Netherlands, when combined with the observations from Germany it seems justified not to exclude fault activity.

Paleoclimate and sea level

The Netherlands experienced an equatorial climate from Namurian to WC times, but the climate became more seasonal tropical (wet-dry extremes) by WD (Van der Zwan et al., 1993; Besly et al., 1993). Selter (1990) found that this change took place during early WD in the western part of the basin but during WC in the eastern part of the Carboniferous basin. This climate change is reflected in a decrease in coal seams and a change in paleosol type from gley soils to redder soils with Fe rich concretions and Ca rich concretions (e.g., Barren Red Beds of late WD age; Besly & Fielding, 1989).

The Late Carboniferous is characterized by icehouse conditions (Caputo & Crowell, 1985; Maynard & Leeder, 1992). As a consequence, it is thought that eustatic sea-level changes triggered by the waxing and waning of the Gondwanan ice-sheet occurred during this interval (Wanless & Shepard, 1936). The magnitude of the sea-level fluctuations is however difficult to constrain (45 - 190 m, Crowley & Baum, 1991; >42 m, Maynard & Leeder, 1992; 25 - 155 m, Süss, 1996; 10 - 95 m, Wright & Vanstone, 2001). Moreover, the distance of the study area from the sea is equivocal (Quirk & Aitken, 1997; Hampson et al., 1999a,b) because direct evidence in the form of marine bands or deltaic deposits is lacking. However, hundreds-of-meter thick cycles of change in sandstone proportion (net-to-gross) in the Upper Carboniferous have been interpreted as due to cyclic change in climate and sea level (Geluk et al., 2001; Süss et al., 2001). The last fully marine transgression occurred at the base of the WC (Aegir marine band, Drozdzewski, 2005). A marine band represents a thin (usually <2 m) layer of clay deposited under marine or brackish water conditions during transgression. In the Ruhr Basin the Aegir marine band is well developed, but it is poor in marine fauna in the Ibbenbüren area (Drozdzewski, 2005).

Data and methods

Twelve wells with a spacing of 500 to 1200 m were studied in the northwestern part of the field (Fig. 1). Available data comprised gamma-ray, sonic, density and neutron logs and 93 m of core (COV-17A, Fig. 1). Interpretation of the core led to a better grip on the depositional environment. Next, facies reconstruction based on well logs has been carried out. Then correlation of sandbodies in the wells, based on estimates of their width, resulted in several cross-sections. Widths were estimated using width-thickness cross-plots and according to the method of Bridge and Tye (2000). Using these cross-sections, paleogeographic maps were created.

Core description and interpretation

In the following section a description and interpretation is presented of the core from COV-17A. A detailed sedimentary log can be found in Figure 4.

Description of the core

Sandstone body G (3066 - 3071 m)

The base of sandbody G is developed as a clear erosional surface overlain by a breccia. Up to 3068 m planar strata in fine to medium grained sandstone occur with small scale cross-stratification. Layers of organic debris are common. From 3068 m to the top structureless fine to medium grained sandstone predominates. The top of the sandstone shows an abrupt transition to parallel laminated silts intercalated between mudstone.

Sandstone body H (3043 - 3057 m)

From the base up to 3051 m, this well sorted fine grained sandstone is characterized by planar strata (Fig. 5). The section from 3051 m to 3043.8 m shows cross strata with intraformational breccias but intervals with structureless sandstone can be seen as well. Brown siderite concretions associated with organic debris occur near the fine-grained tops of stratasets. Locally, small scale cross strata associated with symmetrical ripples, mud drapes and drifted organic debris occur. From 3043.8 m a fining upward sequence occurs, which changes into a paleosol at the top of the sandstone body. Red iron staining increases towards the top of the sandstone body.

Sandstone body J1 (3036 - 3041 m)

This sandstone body has a clear basal erosion surface, overlain by planar bedded fine grained sandstone. The remaining part of this sandstone body consists of coarse grained cross bedded sandstones that contain organic debris. A breccia containing pebbles is intercalated in this sandstone. The top consists of a pebble layer which is sharply overlain by laminated mudstones.

Sandstone body J2 (3005 - 3031 m)

Sandstone body J2 occurs as two storeys (J2a and J2b) that have a combined thickness of 26 m. Each storey (bases of these storeys are at 3017 and 3031 m) has a basal erosion surface. Sandstone is yellowish grey, mainly medium to fine grained (rarely coarse), and contains organic debris. J2a has little vertical variation in grain size, while in J2b a slight fining upward can be seen. The sedimentary structures are mainly medium-scale cross sets ranging in thickness from 5 cm to 50 cm, although intervals of structureless sandstone

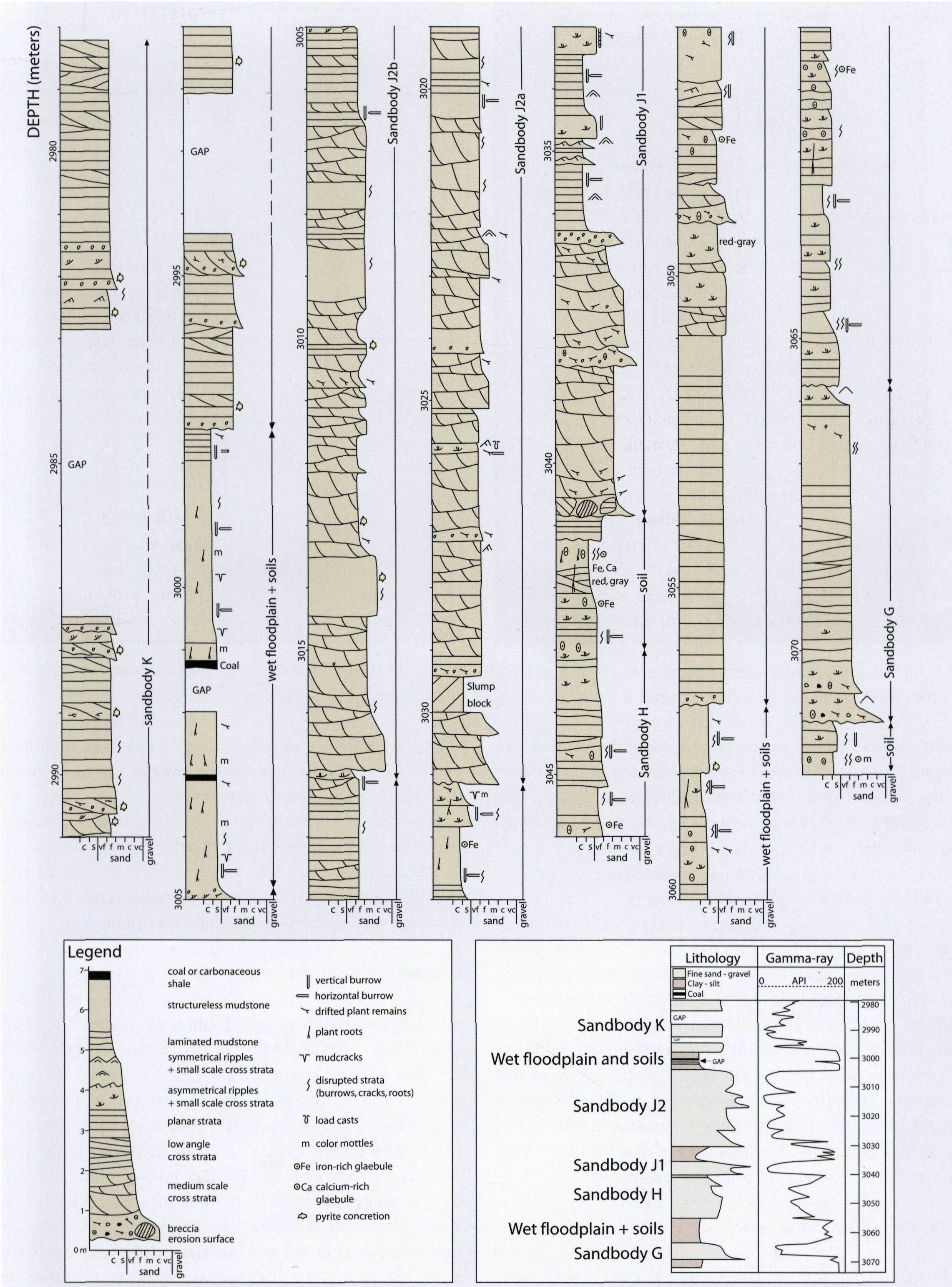

Fig. 4. Sedimentological log of core 17A with interpretation and gamma ray log. For location of the wells, see Fig. 1.

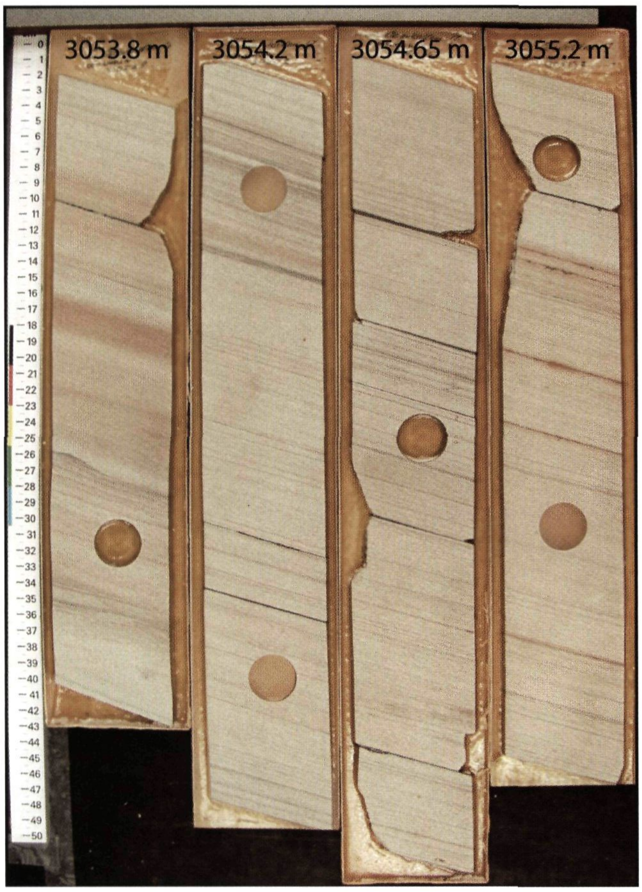

Fig. 5. Photograph showing planar strata in sandstone body H. Depth corresponds to depth in Fig. 4.

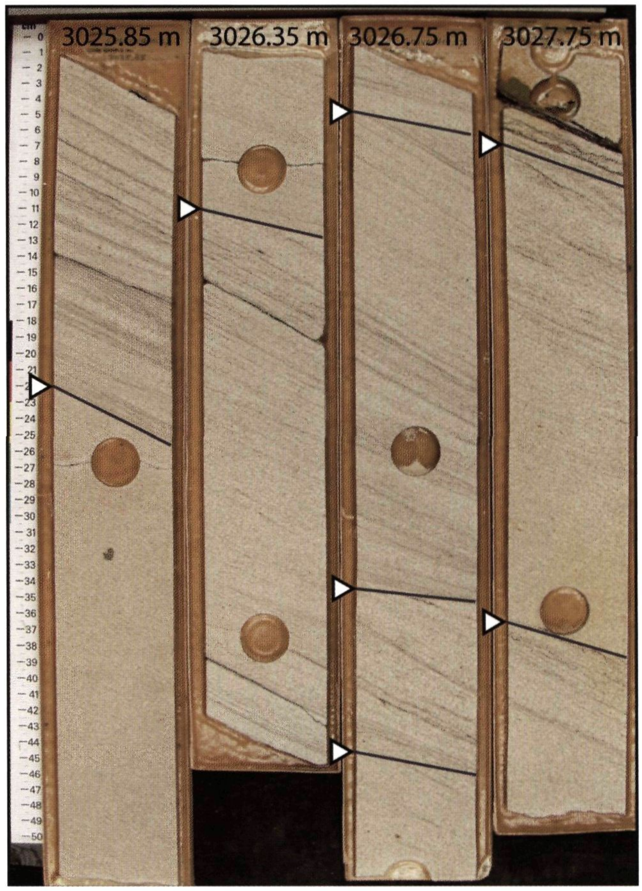

Fig. 6. Photograph showing medium scale cross strata in storey J2A. Depth corresponds to depth in Fig. 4.

occur as well. Figure 6 shows a core photo in which several cross-sets observed in sandstone body J2a can be seen. Planar strata are most abundant in the finer-grained stratasets at the top of the storey. The sandstone body has a sharp top overlain by mudstone.

Sandstone body K (2978 - 2997 m)

The interval from the base (2997 m) to approximately 2982 m is characterized by sharp, erosional surfaces overlain in places by intraformational breccia containing mudstone granules or pebbles, organic debris and siderite concretions. Planar strata are mostly observed on top of the breccias, although structureless sandstone occurs as well. The top part of this sandstone body (from 2982 m onwards) is characterized by planar strata. The thickness of these sets varies between 5 cm to several decimeters. The grain size of this sandstone body ranges from very fine grained to fine grained.

Fine grained sediments

Overall, the fine grained sediments are laminated with intercalations of small-scale cross-stratified siltstone to very fine-grained sandstone (Fig. 4, 3057 - 3065 m). Symmetrical and asymmetrical ripples are associated with the small-scale cross strata. The strata are dark grey to brownish grey in colour. Burrows (horizontal and vertical) and organic debris are common. Strata commonly occur in fining upward stratasets (very fine-grained sandstone to mudstone) that are generally decimeters thick but rarely over 1 m. The top 20 cm of a strataset tends to be dark grey and organic rich and very disrupted with burrows, cracks, and brown stains.

Interpretation of the core

All sandstone bodies described above are interpreted to be deposited in a fluvial dominated environment because of (1) the clear basal erosion surfaces, (2) one dominant transport direction and (3) the dominance of cross-bedded and planar strata. The sediments allow a slightly more detailed subdivision. The main difference between sandstone bodies J1/J2 and G, H and K is the dominance of medium scale cross strata in J1/J2 and planar and low-angle cross strata in G, H and K (Fig. 4). Planar strata are commonly associated with mouth bar or crevasse splay deposits (e.g., Tye & Coleman, 1989; Pèrez-Arlucea & Smith, 1999; Bristow et al., 1999). For instance, the succession from planar strata to (low angle) cross strata observed in sandbody H can be interpreted as a crevasse splay overlain by a fining upward channel bar deposit. Based on the distinct basal erosion surface, sandbody G is thought to represent a channel bar deposit although the remainder of the sandstone body is characterized by planar and low angle cross strata. Because of the absence of a basal erosion surface and the presence of planar strata all through sandbody K, it is interpreted as a mouth bar deposit. The combination of a fining upward sequence of very fine sandstone to mudstone and the relatively thin sandstone body J1 leads to the interpretation of a basal channel bar sandstone overlain by a channel fill. Sandstone body J2 clearly shows two storeys dominated by medium scale cross-strata. These storeys are interpreted as channel bar deposits.

The fine grained sediments are thought to represent deposition in waterlogged floodplains or lakes. These types of deposits are common in modern marshes, swamps, lakes, and distal levees (e.g., Coleman & Gagliano, 1964; Coleman & Prior, 1982; Saucier, 1994). We did not find strong evidence for marine or brackish depositional environments. No marine fossils were found and apart from some muddrapes in sandstone body H, no sedimentary structures were found that could indicate marine or tidal influence.

Paleosols consist of mottled grey-brown siltstones with burrows, crack fills, root traces, and concretions of calcium carbonate and siderite (e.g., Fig. 4, 3071.5 m and 3041.5 m). Some grey paleosols with relatively few concretions contain cm-thick, dark-grey, plant-rich layers and cm-thick coals (Fig. 4, 2998-3005 m). Most paleosols could be classified as weakly developed (inceptisols), with some moderately developed (histosols and vertisols) (Bridge, 2003, Tables 6.2 and 6.3). Such paleosols are typical of wet floodplains in tropical seasonal climates (Cojan, 1999; Kraus & Aslan, 1999). The paleosol with calcareous concretions (Fig. 4, 3041.5 m) is indicative of relatively drier (better drained) depositional conditions (Wright, 1999).

Interpretation of lithology in well logs

In the following section an interpretation is given of the lithology as encountered in the well logs. We present one cross-section (Fig. 7) in which four wells have been correlated. For location of the wells, see Fig. 1.

Sandstone bodies in well logs

Channel-bar deposits can be recognized based on consistent fining upward trends within sandstones, but also sandstone bodies that do not change grain size vertically and even a few sequences that coarsen upward occur. These sandstone bodies (or storeys as is the case of J2) range in thickness from about 8 to 14 m (mainly 10 to 12 m), indicating a maximum bankfull depth of about 14 m and a mean bankfull depth of around 8 m (Bridge & Tye, 2000). Usually, systematic lateral changes in thickness and vertical sequence of grain size in these

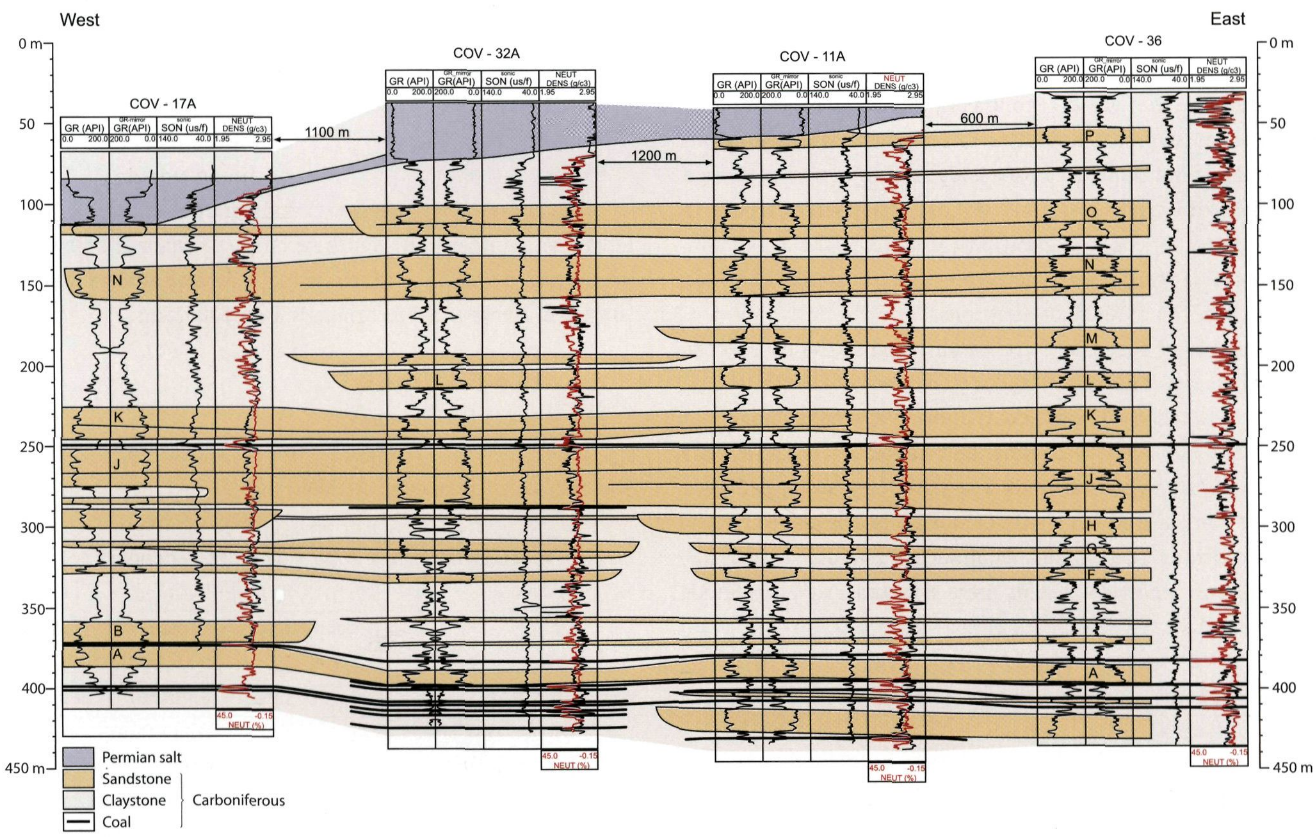

Fig. 7. Well log panel showing interpretation (for location of this panel see Fig. 1).

channel bar deposits occur (Bridge, 1993; Lunt et al., 2004). For example, the bar thickness may change by a factor of two over a lateral distance of the order of a kilometre. Fining upward sequences may change into those with little vertical variation in grain size over a similar distance.

Most of the channel sandstone bodies in Figure 7 are interpreted to be composed of untruncated channel bar deposits within isolated channel belts. It is possible that truncated channel bars are preserved near the bases of some of these sandstone bodies, especially the relatively thick ones, as is the case in sandstone body J2 at 3017 m (Fig. 4). Criteria to determine this include different grain size trends within the different channel bars, and the lower truncated bars being much thinner than the uppermost untruncated one (e.g., Fig. 7, sandstone bodies O and N in COV-11A).

Crevasse deposits in well logs can be recognized by the relatively thin sandstone bodies with a thickness of 2 - 5 m (between sandbody A and F, Fig. 7). Often, these thin sandstone bodies cannot be correlated in all wells, which indicates their restricted width. In some cases the crevasse deposits can be correlated to a channel belt (H, Fig. 7).

Fine grained deposits and coals in well logs

Floodbasin mudstones are meters in thickness and laterally continuous. Meter-scale variations in grain size (clay proportion) of floodplain deposits are clearly discernible. These are fining upward and coarsening upward, and normally include sandstone bodies interpreted as crevasse splays/levees, lacustrine deltas and mouth bars. Four to 8 of these sequences commonly occur within the floodplain sequences between channel belt deposits. According to Bridge (1984), such meter-scale floodplain sequences may be related to avulsion of main channels. However, avulsions do not necessarily result in discernible changes in deposition throughout the floodplain, and crevasse splays and lacustrine deltas can develop and prograde in the absence of avulsions.

A number of coal seams can be observed at the base of the studied succession (Fig. 7) and a prominent coal seam occurs on top of sandbody J. Some coals are laterally continuous, but others are truncated by channel-belt deposits. For instance, a coal seam just below sandbody J in COV-32A (Fig. 1) is absent in the other wells which suggests erosion or non-deposition at these locations. It is possible that such isolated coal seams at the base of sandstone bodies may represent a(n) (isolated) channel fill.

Correlation

In this section the correlation scheme (Fig. 7) and the paleogeographic maps (Fig. 8) will be discussed. As correlation data we used the top of the Maurits Formation (Fig. 2) where a characteristic group of coal seams occurs and a prominent coal seam immediately above sandstone body J (Fig. 7). Another common feature of the studied wells is the cyclic nature of the sandstone proportion (well developed in COV-17A). This also served as a guideline for correlation.

Width of sandbodies

When correlating fluvial sandstone bodies it is important to take into account the direction of paleoflow. The correlation scheme in Figure 7 is perpendicular to the expected direction of paleoflow to the north or northwest based on regional paleogeography (Jones & Glover, 2005), decrease in sediment size to the north-northwest (Den Hartog Jager et al., 1993), and paleocurrent data from cross strata measured in cores from nearby wells. As in most subsurface studies, the width of most of the fluvial sandstone bodies has been obtained using a width-thickness cross-plot. However, Bridge and Tye (2000) used the mean thickness of cross-sets to calculate the potential channel belt width. For sandbody J the medium-scale cross sets could be measured in the COV-17A core (Figs 4 & 5). For the other sandbodies present in the core (G, H and K) there were too little cross-sets available to apply this method as well. Sandstone bodies J2A and J2B have average medium-scale cross-set thicknesses (s) of 27 +/- 17 cm (n = 22) and 26 +/- 18 cm (n = 23) respectively. The mean dune height (hm) can be calculated as follows:

$$h_m = 6 * (s/1.8) \tag{1}$$

The relationship between flow depth (d) and mean dune height is (Bridge & Tye, 2000):

$$d/h_m = 6 - 10 \tag{3}$$

From the obtained flow depth, a range in channel belt width (cbw) can be calculated by using the empirical equations (Bridge & Mackey, 1993; Bridge & Tye, 2000):

$$cbw = 59.9\ d^{1.8} \quad \text{(min)} \tag{4}$$
$$cbw = 192\ d^{1.37} \quad \text{(max)} \tag{5}$$

The results are presented in Table 1.

Table 1. Widths of sandbodies J2A and J2B.

	Sandbody J2A	Sandbody J2B
Flow depth	5.4 - 9 m	5.2 - 8.7 m
Width (min-max)	1246 - 3895 m	1164 - 3719 m

The calculated flow depths are in agreement with the interpretation that sandbody J consists of two superimposed channel belts (J2A and J2B). This is because the calculated flow depths are in agreement with the thickness of the individual

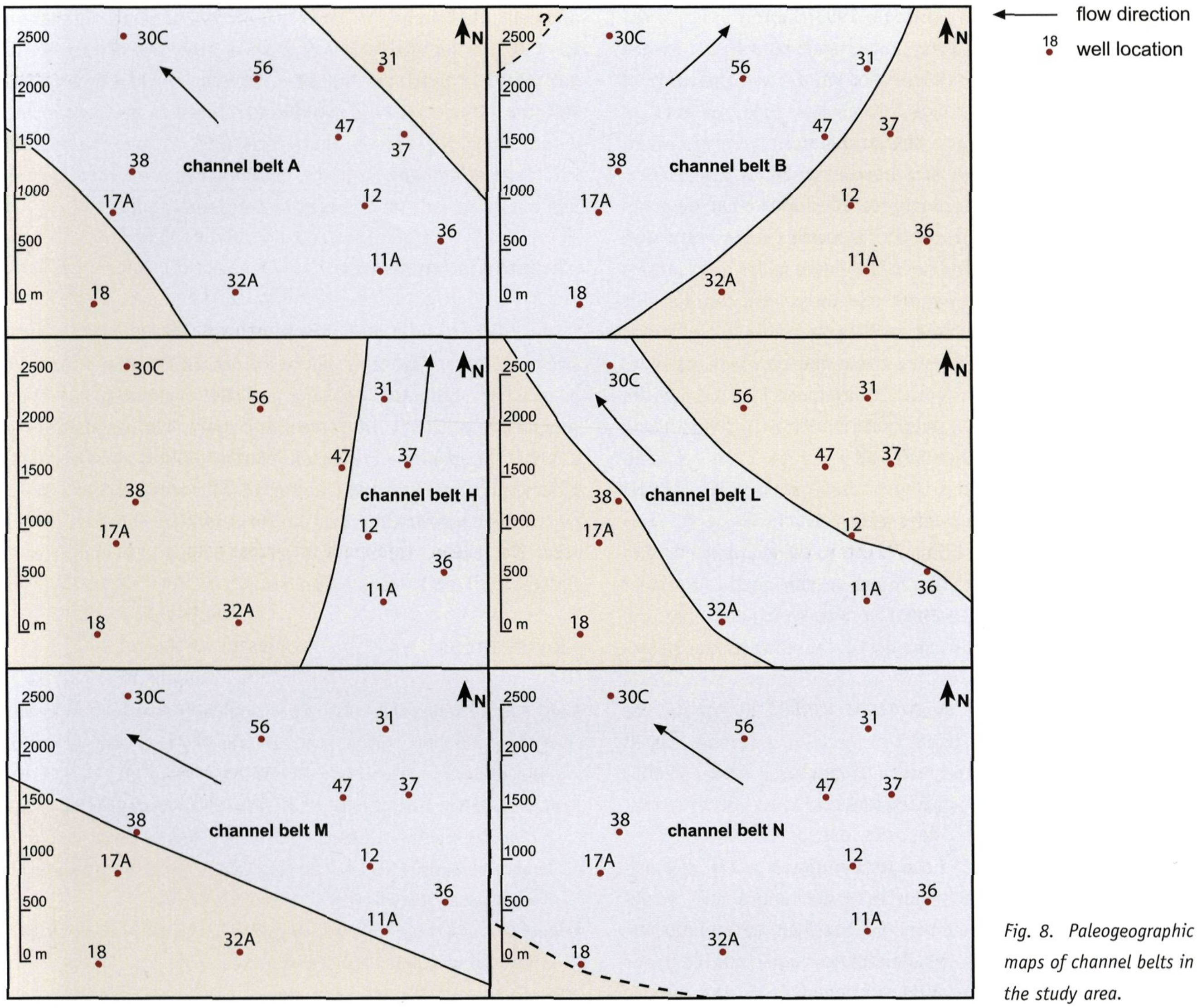

Fig. 8. Paleogeographic maps of channel belts in the study area.

storeys. When sandbody J2 would be interpreted as one single channel belt, the calculated flow depth would be much less than the thickness of the entire sandbody. Superimposed channel belts are likely to result in wider sandstone bodies than those made of single channel belts (Bridge & Mackey, 1993). As the mean width of sandstone bodies J2A and J2B is approximately 2500 m, superposition of these channel belts is likely to result in a width exceeding 2500 m. This is supported by the observation that sandstone body J is present across the entire study area which is 3 km in length and width.

Paleogeographic maps

Some sandstone bodies extend across the whole study area. For sandstone body J, this is expected based on calculations of the width. One margin of some channel-belt sandstone bodies can in certain cases be identified because it is lacking in one or more wells (Fig. 8; sandstone bodies B, H, M and N). In two cases (sandstone bodies A and L), both sides of the channel belt can be identified. This illustrates the benefit of making paleogeographic reconstructions on such a scale. The sandstone bodies are oriented between NW (A, L, M and N?) and NNE (B and H). The northwestward directions agree with the paleoflow directions from literature (David, 1990; Jones and Glover, 2005). The inferred north-northeast direction is considered not to be in direct conflict with the paleoflow directions from literature because (1) a small area has been studied and (2) only one margin of these sandstone bodies could be reconstructed.

Concluding discussion

The alluvial architecture of the Coevorden Field is dominated by untruncated channel bar deposits within isolated channel belts although one sandstone body is probably composed of stacked channel belts (J2). This is in accordance with studies on time-equivalent rocks in the Osnabrück and Ibbenbüren areas (David, 1987, 1990; Jones & Glover, 2005). From our paleogeographic reconstructions (Fig. 8) and the calculation of the width of a channel belt on the basis of cross-set

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thicknesses, it may be stated that on average the width of the channel belts we studied in the Coevorden Field does not exceed 4 km. However, Jones and Glover (2005) interpret the width of the stacked channel belts to be in the order of tens of kilometres. According to our results this is an overestimation.

Interpretation of cores is a useful source of information along with well logs because a better description can be given of the sedimentary environment. However, using only well logs, it remains very difficult to distinguish facies associations such as fining-upward sequences associated with channel fills from a sequence of overbank sandstone overlain by muddy floodbasin deposits. Meters-thick coarsening-upward sequences (mud to sand) in well logs could be produced by progradation of a levee, a crevasse splay, a lacustrine delta (in a floodbasin or into an abandoned channel), or a mouth bar of a main channel. This is the reason why in Fig. 7 only channel belts and crevasse-splays were interpreted.

From paleocurrent measurements in the Coevorden area and a decrease in sandstone percentage to the north-northwest (Den Hartog Jager et al., 1993) it was concluded that the direction of paleoflow has been to the northwest. Our paleogeographic reconstructions are in accordance with these observations. Moreover, David (1987, 1990) interprets the sandstone bodies in the Osnabrück and Ibbenbüren areas as northward flowing systems, while Jones and Glover (2005) found a direction to the west-northwest. These observations match our results.

Although David (1990) interprets brackish water deposits in the Osnabrück and Ibbenbüren areas, Jones and Glover (2005) could not find evidence for brackish water deposits. However, Hedemann et al. (1984) interpret some marine bands in the WC and even WD in northwest Germany. In the core we studied, some diagenetic pyrite in the sandstones was found but pyrite has not been recognized in the claystones or coal. Besides a small amount of muddrapes that were recognized in the sandstones, no further indications of marine (or brackish) water influence has been found. Altogether, in the light of the findings from literature, periodic marine influence can not be excluded.

Percentages of sand are thought to decrease in a northwestward direction (Jones & Glover, 2005). The percentage of sand of the studied succession is approximately 50%. This is indeed less than the 60 % inferred from the Ibbenbüren area (Drozdzewski, 2005) and much less than the 80% found for the Early WD in the Osnabrück area (David, 1987). As our studied succession is probably of Late WC to Early WD age, it is likely that a northwestward decrease in sand-percentage occurs. This may be related to a decrease in thickness of the WC and WD successions to the northwest. The Upper WC has a thickness of 500 m in the Ibbenbüren area (Drozdzewski, 2005) and 450 m in the well Norddeutschland 8 (Fig. 3B) (Hedemann et al., 1984). The Lower WD has a thickness of about 300 m in Norddeutschland 8 (Hedemann et al., 1984) and the Ibbenbüren and Osnabrück areas. The thickness of the Tubbergen Formation is ~450 m in our study area and has a likely age of Late WC to Early WD. In that case indeed a decrease in thickness of the WC and WD occurs. This may be attributed to tectonic movements along the Gronau Fault zone (Fig. 3) and the coming into existence of the Ems Low, of which the Coevorden Field is the westernmost - and therefore peripheral - part.

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References

Besly, B.M. & Fielding, C.R., 1989. Palaeosols in Westphalian coal-bearing and red-bed sequences, central and northern England. Palaeogeography, Palaeoclimatology, Palaeoecology 70: 303-330.

Besly, B.M., Burley, S.D. & Turner, P., 1993. The Late Carboniferous 'Barren Red Bed' play of the Silver Pit area, Southern North Sea. *In:* Parker, J.R. (ed.): Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference. The Geological Society (London): 727-740.

Bridge, J.S., 1984. Large-scale facies sequences in alluvial overbank environments. Journal of Sedimentary Petrology 54: 583-588.

Bridge, J.S., 1993. The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers. *In:* Best, J.L. & Bristow, C.S. (ed.): Braided Rivers. Geological Society Special Publication (London) 75: 13-72.

Bridge, J.S., 2003. Rivers and floodplains. Blackwells (Oxford): 491 pp.

Bridge, J.S. & Mackey, S.D., 1993. A theoretical study of fluvial sandstone body dimensions. *In:* Flint, S.S. & Bryant, I.D. (ed.): Geological modeling of hydrocarbon reservoirs. International Association of Sedimentologists Special Publication 15: 213-236.

Bridge, J.S. & Tye, R.S., 2000. Interpreting the dimensions of ancient fluvial channel bars, channels, and channel belts from wireline-logs and cores: American Association of Petroleum Geologists Bulletin 84: 1205-1228.

Bristow, C.S., Skelly, R.L. & Ethridge, F.G., 1999. Crevasse splays from the rapidly aggrading, sand-bed, braided Niobrara River, Nebraska: effect of base-level rise: Sedimentology 46: 1029-1047.

Caputo, M.V. & Crowell, J.C., 1985. Migration of glacial centers across Gondwana during Paleozoic Era. Geological Society of America Bulletin 96: 1020-1036.

Cojan, I., 1999. Carbonate-rich palaeosols in the Late Cretaceous-Early Palaeogene series of the Provence Basin (France). *In:* Thiry, M. & Simon-Coicon, R. (ed.): *Palaeoweathering, Palaeosurfaces and related Continental Deposits.* International Association of Sedimentologists Special Publication 27: 323-335.

Coleman, J.M. & Gagliano, S.M., 1964. Sedimentary structures: Mississippi River deltaic plain. *In:* Middleton, G.V. (ed.): Primary sedimentary structures and their hydrodynamic interpretation. Society for Sedimentary Geology (SEPM) Special Publication 12: 133-148.

Coleman, J.M. & Prior, D.B., 1982. Deltaic environments. *In:* Scholle, P.A. & Spearing, D. (ed.): Sandstone Depositional Environments. American Association of Petroleum Geologists (Tulsa, Oklahoma): 139-178.

Coward, M.P., 1993. The effect of late Caledonian and Variscan continental escape tectonics on basement structure, Paleozoic basin kinematics and subsequent Mesozoic basin development in NW Europe. *In:* Parker, J.R. (ed.): Petroleum Geology of Northwest Europe: Proceedings of the 4th Conference. The Geological Society (London): 1095-1108.

Crowley, T.J. & Baum, S.K., 1991. Estimating Carboniferous sea-level fluctuations from Gondwanan ice extent. Geology 19: 975-977.

David, F., 1987. Sandkörper in fluviatilen Sandstein des Unteren Westfal D (Oberkarbon) am Piesberg bei Osnabrück. Facies 17: 51-58.

David, F., 1990. Sedimentologie und Beckenanalyse im Westfal C und D des nordwestdeutschen Beckens. DGMK-Bericht 384-3.

Den Hartog Jager, D.G., Boekelman, W. & Mijnlieff, H.F., 1993. Regional geology of the Carboniferous in the NE Netherlands: Nederlandse Aardolie Maatschappij (NAM) internal report No. 24929, Assen.

Drozdzewski, G., 1985. Tiefentectonik der Ibbenbürener Karbon-Scholle. *In:* Drozdzewski, G., Engel, H., Wolf, R. & Wrede, V. (ed.): Beiträge zur Tiefentectonik westdeutscher Steinkohlenlagerstätten. Geol. L.-Amt Nordrh.-Westf. (Krefeld): 189-216.

Drozdzewski, G., 2005. Zur sedimentären Entwicklung des Subvariscikums im Namurium und Westfalium Nordwestdeutschlands. Cour. Forsch.-Inst. Senkenberg 254: 271-320.

Drozdzewski, G. & Wrede, V., 1994. Faltung und Bruchtektonik – Analyse der Tektonik im Subvariscikum. Fortschritte Geol. Rheinl. U. Westfalen, Krefeld, 38: 7-187.

Fraser, A.J. & Gawthorpe, R.L., 1990. Tectono-stratigraphic development and hydrocarbon habitat of the Carboniferous in northern England. *In:* Hardman, R.F.P. & Brooks, J. (ed.): Tectonic events responsible for Britain's oil and gas reserves. Geological Society Special Publication (London) 55: 49-86.

Geluk, M., De Haan, H., Schroot, B. Wolters, B. & Nio, S.D., 2002. The Permo-Carboniferous gas play, Cleaver bank High area, southern North Sea, the Netherlands. *In:* Hills, L.V., Henderson, C.M. & Bamber, E.W. (ed.): Carboniferous and Permian of the World. Canadian Society of Petroleum Geologists Memoir 19: 877-894.

Hampson, G.J., Davies, S.J. Elliott, T., Flint, S.S. & Stollhofen, H., 1999a: Incised valley fill sandstone bodies in Upper Carboniferous fluvio-deltaic strata: recognition and reservoir characterization of Southern North Sea analogues. *In:* Fleet, A.J. & Boldy, S.A.R. (ed.): Petroleum geology of Northwest Europe: Proceedings of the 5th Conference. The Geological Society (London): 771-788.

Hampson, G.J., Stollhofen, H. & Flint, S., 1999b. A sequence stratigraphic model for the Lower Coal Measures (Upper Carboniferous) of the Ruhr district, north-west Germany. Sedimentology 46: 1199-1231

Hedemann, H.-A., Schuster, A., Stancu-Kristoff, G. & Lösch, J., 1984. Die Verbreitung der Kohlenflöze des Oberkarbons in Nordwest-deutschland und ihre stratigraphische Einstufung. Fortschr. Geol. Rheinld. U. Westf. 32: 39-88.

Jankowski, B., David, F. & Selter, V., 1993. Facies complexes of the Upper Carboniferous in northwest Germany and their structural implications. *In:* Gayer, R.A., Greilling, R.O. & Vogel, A.K. (ed.): Rhenohercynian and Subvariscan fold belts. Vieweg, Wiesbaden: 138-158.

Jones, N.S. & Glover, B.W., 2005. Fluvial sandbody architecture, cyclicity and sequence stratigraphical setting – implications for hydrocarbon reservoirs: the Westphalian C and D of the Osnabrück and Ibbenbüren area, northwest Germany. *In:* Collinson, J.D., Evans, D.J., Holliday, D.W. & Jones, N.S. (ed.): Carboniferous hydrocarbon geology – The southern North Sea and surrounding areas. Yorkshire Geological Society Occasional Publication 7: 57-74.

Kraus, M.J. & Aslan, A., 1999. Palaeosol sequences in floodplain environments: a hierarchical approach. *In:* Thiry, M. & Simon-Coicon, R. (ed.): Palaeoweathering, Palaeosurfaces and related Continental Deposits. International Association of Sedimentologists Special Publication 27: 303-321.

Leeder, M.R. & McMahon, A.H., 1988. Upper Carboniferous (Silesian) basin subsidence in northern Britain. *In:* Besly, B.M. & G. Kelling (ed.): Sedimentation in a synorogenic basin complex. The Upper Carboniferous of Northwest Europe: 43-52.

Lunt, I.A., Bridge, J.S. & Tye, R.S., 2004. A quantitative, three-dimensional model of gravelly braided rivers. Sedimentology 51: 377-414.

Maynard, J.R. & Leeder, M.R., 1992. On the periodicity and magnitude of late Carboniferous glacio-eustatic sea-level changes. Journal of the Geological Society, London: 149, 303-311.

Menning, M., Alekseev, A.S., Chuvashov, B.I., Davydov, V.I., Devuyst, F.-X., Forke, H.C., Grunt, T.A., Hance, L., Heckel, P.H., Izokh, N.G., Jin, Y.-G., Jones, P.J., Kotlyar, G.V., Kozur, H.W., Nemyrovska, T.I., Schneider, J.W., Wang, X.-D., Weddige, K., Weyer, D. & Work, D.M., 2006. Global time scale and regional stratigraphic reference scales of Central and West Europe, East Europe, Tethys, South China, and North America as used in the Devonian-Carboniferous-Permian Correlation Chart 2003 (DCP 2003). Palaeogeography, Palaeoclimatology, Palaeoecology 240: 318-372.

Pagnier, H.J.M. & Van Tongeren, P., 1996. Upper Carboniferous of borehole 'De Lutte-6' and evaluation of the Tubbergen in the eastern and southeastern parts of the Netherlands. Mededelingen Rijks Geologische Dienst 55 (Haarlem): 3-30.

Pèrez-Arlucea, M. & Smith, N.D., 1999. Depositional patterns following the 1870's avulsion of the Saskatchewan River (Cumberland Marshes, Saskatchewan): Journal of Sedimentary Research 69: 62-73.

Quirk, D.G. & Aitken, J.F., 1997. The structure of the Westphalian in the northern part of the southern North Sea. *In:* Ziegler, K., Turner, P. & Daines, S.R. (ed.): Petroleum Geology of the Southern north Sea: Future Potential. Geological Society Special Publication (London) 123: 143-152.

Saucier, R.T., 1994. Geomorphology and Quaternary Geologic History of the Lower Mississippi Valley. Mississippi River Commission (Vicksburg): 364 pp.

Schroot, B.M. & De Haan, H.B., 2003. An improved regional structural model of the Upper Carboniferous of the Cleaver Bank High based on 3D seismic interpretation. *In:* Nieuwland, D.A. (ed.): New insights into structural interpretation and modeling: Geological Society Special Publication (London) 212: 23-37.

Selter, W., 1990. Sedimentologie und Klimaentwicklung im Westfal C/D und Stefan des nordwestdeutschen Oberkarbon-Beckens. DGMK-Bericht 384-4.

Süss, M.P., 1996. Sedimentologie und Tektonik des Ruhr-Beckens: Sequenzstratigraphische Interpretation und Modellierung eines Vorlandbeckens der Varisciden. Dissertation Rheinische Friedrich-Wilhelms University (Bonn): 147 pp.

Süss, M P., Drozdzewski, G. & Schäfer, A., 2001. The Ruhr and Aachen basins – sedimentary environments, sequence stratigraphy model, and synsedimentary tectonics of Variscan foreland basins (Namurian B/C to Westphalian C, W. Germany). *In:* Hills, L.V., Henderson, C.M. & Bamber, E.W. (ed.): Carboniferous and Permian of the World. Canadian Society of Petroleum Geologists Memoir 19: 208-227.

Tye, R.S. & Coleman, J.M., 1989. Depositional processes and stratigraphy of fluvially dominated lacustrine deltas: Mississippi Delta Plain. Journal of Sedimentary Petrology 59: 973-996.

Van Adrichem Boogaert, H.A. & Kouwe, W.F.P. (ed.), 1993. Stratigraphic nomenclature of the Netherlands, revision and update by RGD and NOGEPA. Mededelingen Rijks Geologische Dienst (Delft/Haarlem) 50: 1-40.

Van der Zwan, C.J., Van der Laar, J.G.M., Pagnier, H.J.M. & Van Amerom, H.W.J., 1993. Palynological, ecological and climatological synthesis of the Upper Carboniferous of the well De Lutte-6 (East Netherlands): Comptes Rendus XII ICC-P 1: 167-186.

Wanless, H.R. & Shepard, F.P., 1936. Sea level and climate changes related to late Palaeozoic cycles. Geological Society of America Bulletin 47: 1177-1206.

Wright, V.P., 1999. Assessing flood duration gradients and fine-scale environmental change on ancient floodplains. *In:* Marriott, S.B. & Alexander, J. (ed.): Floodplains: Interdisciplinary Approaches. Geological Society Special Publication (London) 163: 279-287.

Wright, V.P. & Vanstone, S.D., 2001. Onset of late Palaeozoic glacio-eustasy and the evolving climates of low latitude areas: a synthesis of current understanding. Journal of the Geological Society of London 158: 579-582

Ziegler, P.A., 1990. Geological Atlas of Western and Central Europe (2nd edition). Shell Internationale Petroleum Maatschappij B.V.; Geological Society Publishing House (Bath): 232 pp.