MORPHOLOGY, STRATIGRAPHY AND GENESIS OF BURIED MID-PLEISTOCENE TUNNEL-VALLEYS IN THE SOUTHERN NORTH SEA BASIN

by

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DECLARATION

I have composed this thesis entirely by myself.
The work and results are my own, except where otherwise acknowledged.

There is no exercise of the intellect which is not, in the final analysis, useless.
A philosophical doctrine begins as a plausible description of the universe;
with the passage of the years it becomes a mere chapter - if not a
paragraph or a name - in the history of philosophy.

Borges

ABSTRACT

Conflicting views of the hydrology and dynamics of ice sheets apply to tunnel-valleys in sediment beds. Field data have been attributed to opposed models of superimposed glaciofluvial erosion, either beneath receding ice margins or by 'catastrophic' subglacial floods, whereas theoretical models infer a coupled response to the drainage of water through a deformable substrate. The southern reaches of the Elsterian glaciation are associated with large buried tunnel-valleys, locally over 400 m in relief. Their basal morphology and fill stratigraphy are examined over a 100x150 km area of the southern North Sea Basin, using seismic reflection grids (including a 3D-volume over a 39x22 km area) supplemented by downhole data, mainly collected for hydrocarbon exploration. The results support a diachronous formation by drainage beneath a receding ice margin coupled to its bed.

Basal contours define elongate basins up to 500 m in relief (575 m bsl), shallow relative to their widths (0.5-6 km) and steep-sided (5-40°), containing nested axial sub-basins 10-170 m in relief and 1-45 km long. The basins converge to the south in a rectilinear arborescent plan form. Local divergent or anastomosing patterns, and angular offsets of basal segments up to 20 km long, result from erosional overlap by younger basal elements to the north. The basal morphology is superimposed on trends in the incised Plio-Pleistocene substrata, which have only locally influenced cross-sectional form (stratal benches) and orientation (across faults). However, the basins increase eastwards in size and spacing in parallel with the thickness of sandy Pleistocene sediments.

Contoured reflecting surfaces within the fill reveal three seismic sequences, from bottom to top: **I** - axially downlapping clinoform surfaces, 0.5-3.5° in gradient and 3-20 km in extent, which dominate the fill with thicknesses up to 400 m beneath a depositional upper boundary up to 120 m in axial relief; the northwards prograding sequence includes discordant surfaces of erosion which coincide with plan-form complexity, separating younger cut-and-fill elements to the north; **II** - subhorizontal surfaces which onlap basins at the surface of I and the upper walls with thicknesses up to 150 m; **III** - complex surfaces which fill incised basins up to 70 m in axial relief at the surface of II with thicknesses up to 100 m and extend between some tunnel-valleys.

Downhole data show sand-dominated assemblages beneath a mud-dominated cap. The sediments contain microfossils and secondary components (lignite, shell fragments, glauconite) reworked from series as young as early Pleistocene. A marine foraminiferal zone occurs within the upper muds. The succession records glaciofluvial to lacustrine deposition prior to marine conditions of the Holsteinian interglacial. Correlation to the seismic stratigraphy shows that progradation of glaciofluvial sands (I) was followed by subhorizontal deposition of sands and then muds in lake basins (II) and finally complex deposition in and between marine basins (III).

Evidence for basal erosion to the south versus glaciofluvial fill progradation to the north are reconciled in a model of the contemporaneous excavation and backfilling of basins beneath the northwards receding ice margin. The rectilinear basal morphology, including erosional overlap by younger elements to the north, is attributed to shifting ice margin orientations and local readvances during recession. Clinoform surfaces (I) represent glaciofluvial backsets, previously observed or inferred within eskerine ridges, here attributed to a distributed system of submarginal streams rising to feed over- and backlapping subaqueous outwash fans along a migrating grounding line. Overlying surfaces (II) record pro-glacial outwash of sands and increasingly distal accumulation of muds in lake basins, which elongated with ice recession and persisted until the marine transgression (III).

The tunnel-valleys formed by reworking of sedimentary materials, by systems of streams within larger basins, beneath the outer tens of kilometres of the ice sheet. They are genetically equivalent to eskers, which record the headward evolution of englacial surface-to-bed drainage within melting ice margins during deglaciation. Their inverted morphology, and the observed relation of size and spacing to substrate (aquifer) thickness, are consistent with the response of a deformable sediment bed to seasonally large volumes of surface recharge. The tunnel-valleys record the diachronous coupling of substrate and glacier hydrology beneath the receding, melting ice sheet margin.

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I am especially grateful to Pieter, both for entertaining and challenging my ideas and for being there when it mattered. The thesis and I owe him more than we seem to know how to say.

In the last months, chapters of the thesis were critically read by Geoffrey, Roger Scrutton, Nigel, Pieter, Dave Long, Martyn Stoker (and 5 pages by Bertrand Maillot, including this one). The other Dan (Barlow) read the abstract, and liked it every time. Mais Agnès Baltzer est la seule à avoir tout lu et en plus elle me soutenait tant que dire que je lui suis très reconnaissant n'est guère, voilà.

I began this thesis with the love and support of Corinne. I finished it on my own, but not alone. Thank you Rocio, Roos and Pieter, Chus super-guapa, Ciara and Damian, Annichou et famille, Bertrand, Aloys, Sergei, Gilles, encore Agnès B., Agnès G. et Julian, Manuelito, my family and every day Jill and Bob. I made it through the last year with your friendship and support.

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CHAPTER 1
INTRODUCTION

1.1 Nature of the Research

This thesis concerns certain valley-like features found in glaciated sedimentary basins, which record channelised drainage beneath continental ice sheets. Tunnel-valleys are known both as landforms of the last glaciation, and as buried or stratigraphical forms of the last and earlier glaciations. I present an analysis of a series of large mid-Pleistocene tunnel-valleys within the southern North Sea Basin, using seismic reflection grids and downhole data collected mainly in the course of hydrocarbon exploration. The results include new information on tunnel-valley basal morphology and fill stratigraphy. These are used to support a model of their diachronous formation by drainage beneath the ice margin during its deglacial recession.

I begin by establishing the research context and the thesis rationale. I then review previous work on tunnel-valleys (section 1.2) and the setting of the southern North Sea study area (section 1.3).

1.1.1 Glacial Geological Context

Glacial geology addresses the interaction of glaciers with their beds through processes of erosion, deposition and deformation. Formerly glaciated continental areas provide a record of such processes beneath ice sheets and across a range of bed types. A distinction is made between 'hard' or indurated rock beds and 'soft' or deformable sediment beds (Boulton and Jones, 1979). Glacier hydrology - the drainage of water through the ice and along or through the bed to the margin - is integral to subglacial processes on both types of bed (Paterson, 1994). Geological records of glacial drainage are thus of interest both in relation to the dynamical behaviour of ice sheets, and in their own right as expressions of particular erosional and depositional processes. Channelised subglacial drainage is recorded in two main forms, eskers and tunnel-valleys. Eskers are common to hard bed areas (Clark and Walder, 1994) whereas tunnel-valleys are particular to sedimentary substrata (Boulton and Hindmarsh, 1987).

1.1.1 a) tunnel-valleys

Tunnel-valleys represent one of the larger lineal glacial geological features, with dimensions up to several hundreds of metres in relief, kilometres in width and tens of kilometres in length. They have been discussed for over a century in the literature of northern continental Europe, as meltwater landforms of the last deglaciation of the southern Baltic lowlands (Jentzsch, 1884; Ussing, 1903-07; Werth, 1907-12; Woldstedt, 1913-1929; Nilsson, 1983). Buried equivalents, filled with meltwater deposits, have been identified from borehole evidence and related to glaciations prior to the last (Woodland, 1970; Ehlers et al., 1984). They have also been recognised on the continental shelf, both as bathymetric deeps (Valentin, 1957) and from seismic reflection profiling as at least three levels of channeling in the North and Irish Sea Basins correlated to successive continental glaciations (Cameron et al., 1987; Ehlers and Wingfield, 1991). They thus appear to be characteristic elements of glaciated sedimentary basins and of high preservation potential (Einsele, 1992). They may also be common within the pre-Pleistocene glacial record as 'paleovalleys' (e.g. Beuf et al., 1970; McClure, 1978; Levell et al., 1988; Powell et al., 1994).

Tunnel-valleys were long considered a regional phenomenon and received passing reference in textbooks written in English (e.g. Flint, 1947; Sugden and John, 1976). However, since their recognition by Wright (1973) in Minnesota they have been reported as landforms from wide areas of the last North American ice sheets (e.g. Shaw, 1983; Attig et al., 1989; Brennand and Sharpe, 1993; Booth, 1994). The subsurface extent of a few last-glaciation features has been demonstrated (Christiansen, 1987), but buried tunnel-valleys of previous glaciations have not been identified, although they appear to be present beneath the western Canadian plains (e.g. Klassen, 1989; Christiansen, written communication, 1992). Tunnel-valleys have also been recognised as surficial and buried features of the eastern Canadian continental shelf, attributed to at least one glaciation (Barnes and Piper, 1978; Boyd et al., 1988; Loncarevic et al., 1992). They are generally accepted as "grand manifestations of the complex glacial hydrologic system" (Mooers, 1989a).

1.1.1 b) ice sheet hydrology and dynamics

Tunnel-valleys provide records of channelised drainage, the interpretation of which is implicit in controversies surrounding the hydrodynamics of ice sheets. Two types of hypothesis seek to account for lowland landform assemblages which include tunnel-valleys, differing over the nature and rates of ice sheet drainage and the source of the water. They can be described as diachronous versus catastrophist models. A third type of model seeks to explain tunnel-valleys in (non-catastrophic) terms of a response to subglacial deformation of unlithified substrata.

Diachronous models attribute tunnel-valleys to drainage beneath the ice sheet margin during its deglacial recession. This corresponds to the original concept of Ussing (1903-07) and Werth 1907-12), based on observations of the last deglaciation of the Baltic lowlands, in which the association of tunnel-valleys with outwash and moraines was used to infer drainage of water under pressure, transverse to successive ice margin lines (see Kozarski, 1966/67; Galon, 1983). The model was later extended to buried tunnel-valleys in explanation of their glaciofluvial fill (Woodland, 1970; Ehlers, 1981). The implied uphill and transverse movement of subglacial water is consistent with theoretical expectations (Shreve, 1972). A comparable model was proposed independently by Mooers (1989a), who showed that an apparently anastomosing network of last-glacial tunnel-valleys and eskers in Minnesota was in fact a composite of short segments (10-20 km), leading to outwash fans along successive margins. He emphasised the dominant role of surface water joining the basal system via an englacial network that 'developed headward during ice retreat'. This was significant as an alternative to the water sources invoked in catastrophist models.

Catastrophist models envisage the episodic release of water from large subglacial reservoirs. Two types of model can be distinguished, sub-marginal and subglacial, originating with Wright (1973) and Shaw (1983), respectively. Wright examined the same Minnesotan tunnel-valleys later studied by Mooers (1989a), but treated them as a single system draining 150 km to the margin. He found that the water volumes necessary to maintain flow exceeded estimates of subglacial melting. He proposed flooding from subglacial lakes only because he saw no means by which surface water could penetrate to the bed through what he estimated to be a frozen margin. This was subsequently resolved by Mooers, but the 'frozen margin' flood model was extended to other tunnel-valleys in North America (e.g. Attig et al., 1989) and in northern Europe (Wingfield, 1990; Ehlers and Linke, 1989; Piotrowski, 1994).

Shaw (1983) proposed a model of subglacial sheet-floods to explain the morphology of drumlin fields, which include tunnel-valleys and eskers. Erosion and deposition took place subglacially as discharge waned and the tunnel-valleys and eskers were attributed to late-stage channelised flow. The model has been applied to wide areas of the last deglaciation of Canada, but the location of the ice margin has been of little interest (e.g. Shaw and Gorrell, 1991; Brennand and Sharpe, 1994). Anastomosing tunnel-valley networks have been suggested to be explicable 'only' by synchronous erosion by subglacial floods (Brennand and Shaw, 1994). Surface melting of the margins during deglacial recession has been suggested to contribute to reservoirs beneath the ice sheet centre (Shoemaker, 1992). The sheet-flood model has been suggested to apply to buried tunnel-valleys of the Scotian Shelf (Boyd et al., 1988).

The preceding models are based primarily on the explanation of landform assemblages in terms of hypothetical glacial processes. A third type of model derives from theoretical research into subglacial deformation as a response to effective pore pressures beneath glaciers 'coupled' to groundwater flow in permeable substrata (Boulton and Jones, 1979; Boulton, 1986). The controlled release of subglacial meltwater is considered essential to ice sheet stability, either to prevent deformation (Shoemaker, 1986) or to allow controlled deformation (Boulton and Hindmarsh, 1987). The latter authors suggested that large ice-filled valleys may be excavated by much smaller channels, initiated by piping and enlarged both downward and headward to form a arborescent network beneath the ice margin, with sizes and spacings related to substrate transmissibilities and subglacial melt rates. The model is described as 'uniformitarian' as opposed to catastrophist and is compatible with the diachronous sub-marginal models described above.

Boulton and Hindmarsh (1987) did not address the source of the water in their model, but larger discharges implied larger tunnel-valleys. Shoemaker (1986) considered that discharge variations, such as those due to seasonal surface recharge, could modify such a drainage system. Arnold and Sharp (1992) presented a simple numerical analysis which showed that meltwater discharges increase by up to ten times during deglaciation due to ablation and that the access of this surface water to the bed beneath the margins could influence the deglacial dynamics of an ice sheet as a whole.

The starting point of this thesis is that tunnel-valleys provide large scale records of channelised subglacial drainage in sedimentary basins. The nature of this process is a matter of dispute between differing views of ice sheet hydrodynamics. Questions apply as to the rates and extent of drainage activity (catastrophic and widespread, versus diachronous and sub-marginal), the source of the water (basal melting versus surface recharge), and the influence of substrate characteristics. I address these questions by examination of large buried tunnel-valleys in the southern North Sea Basin.

1.1.2 Study Area and Rationale

1.1.2 a) regional setting

Lowland areas of northwestern Europe are underlain by a series of intracratonic sedimentary basins flanked by structural highs (Fig. 1.1). The Northwest European Basin extends from the North Sea to Poland and has accumulated up to 9 km of sediments since the Palaeozoic. The succession includes widespread Zechstein salts of Permian age, diapirism of which has deformed and faulted younger series. Chalks of upper Cretaceous to Palaeocene (Danian) age are overlain by up to 3.5 km of Cenozoic clastics, which define the North Sea Basin (Fig. 1.1).

Quaternary thicknesses may locally exceed 1000 m along the axis of the North Sea Basin (Caston, 1977). The record is dominated by lower to mid-Pleistocene fluvio-deltaic and marine sediments, accumulation of which continued until the onset of continental glaciation in the mid-Pleistocene (Cameron et al., 1987; Zagwijn, 1989). Evidence for early Pleistocene glaciation of the basin margins is recognised in fluvial deposits of both The Netherlands (Zagwijn, 1985, 1989) and southern England (Bowen et al., 1986). The first continental (lowland) glaciation is recorded by glacial deposits of the Elsterian stage, which corresponds to the Anglian stage of Britain (Bowen and Sykes, 1988; see Jones and Keen, 1993). The Anglian has been correlated with oxygen isotope stage 12 of the deep-sea record, based on amino acid racemization ratios from sediments overlying tills (Bowen et al., 1989). Isotope stage 12 is radiometrically dated to 4.8-4.3 x 10^4 years BP (Johnson, 1982). The Elsterian has also been correlated to stage 12, based on criteria including palaeomagnetic data and radiometric dating of overlying sediments (Sibrava, 1986; Sarnthein et al., 1986), although Zagwijn (1985, 1989) advocated correlation of the Elsterian to isotope stage 10 based on pollen zones constrained by the palaeomagnetic time scale.

The Elsterian is the first of at least three continental glaciations recognised from the onshore record, each of which extended entirely or partly across the North Sea Basin (Fig. 1.2). Correlation of the Elsterian to isotopic stage 12 implies five rather than three glacial stages in the last 4.5 x 10^4 years and the 'additional' glaciations have been proposed from local stratigraphical evidence (e.g. Jones and Keen, 1993). The traditional glaciations of the Elsterian, Saalian and Weichselian stages have been correlated with at least three stacked unconformities recognised within the later Pleistocene seismic stratigraphy of the North Sea Basin, each of which are associated with tunnel-valleys (Fig. 1.3).

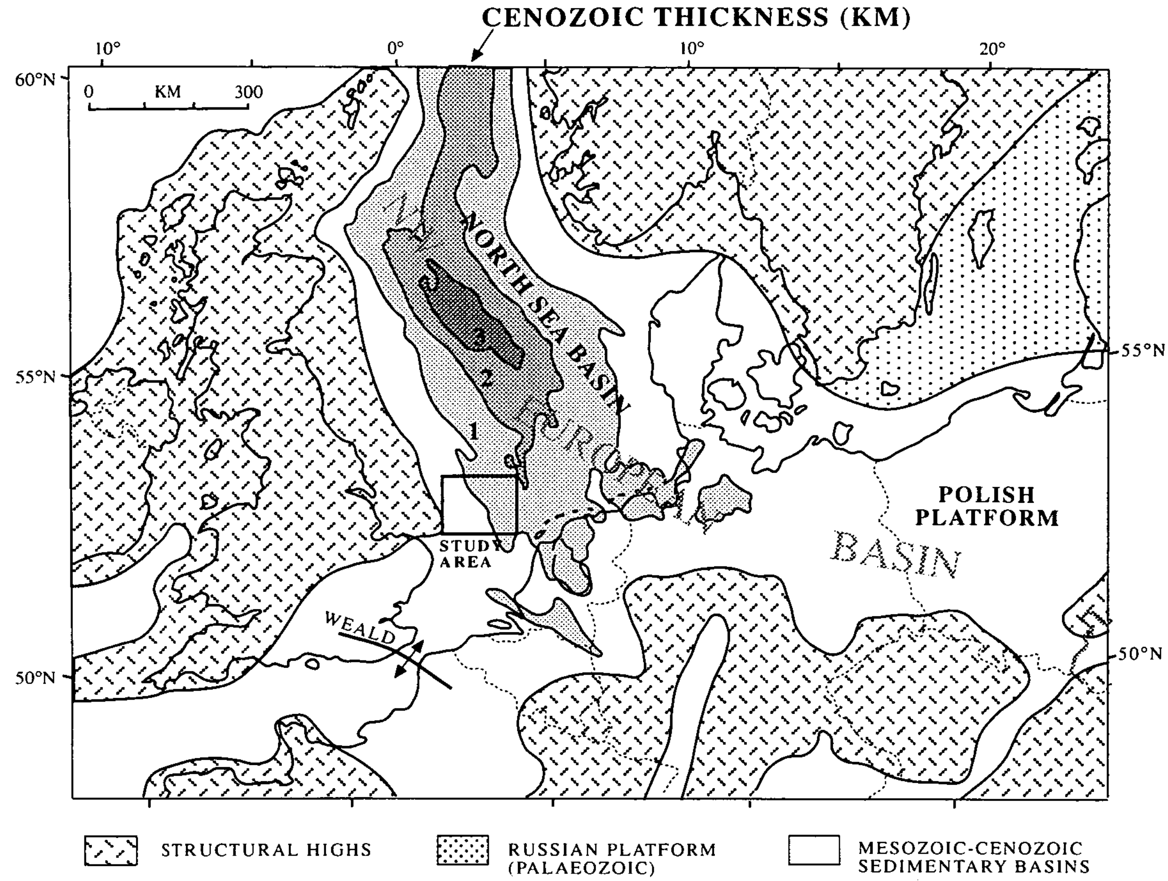


Figure 1.1 - Study area setting within the sedimentary lowlands (basins) of northern Europe. The North Sea Basin and Polish Platform form the upper (Cenozoic) portion of the Northwest European Basin. Structural framework modified from Ziegler and Louwerens (1979), Cenozoic thickness (contoured in kilometres) from Ziegler (1982).

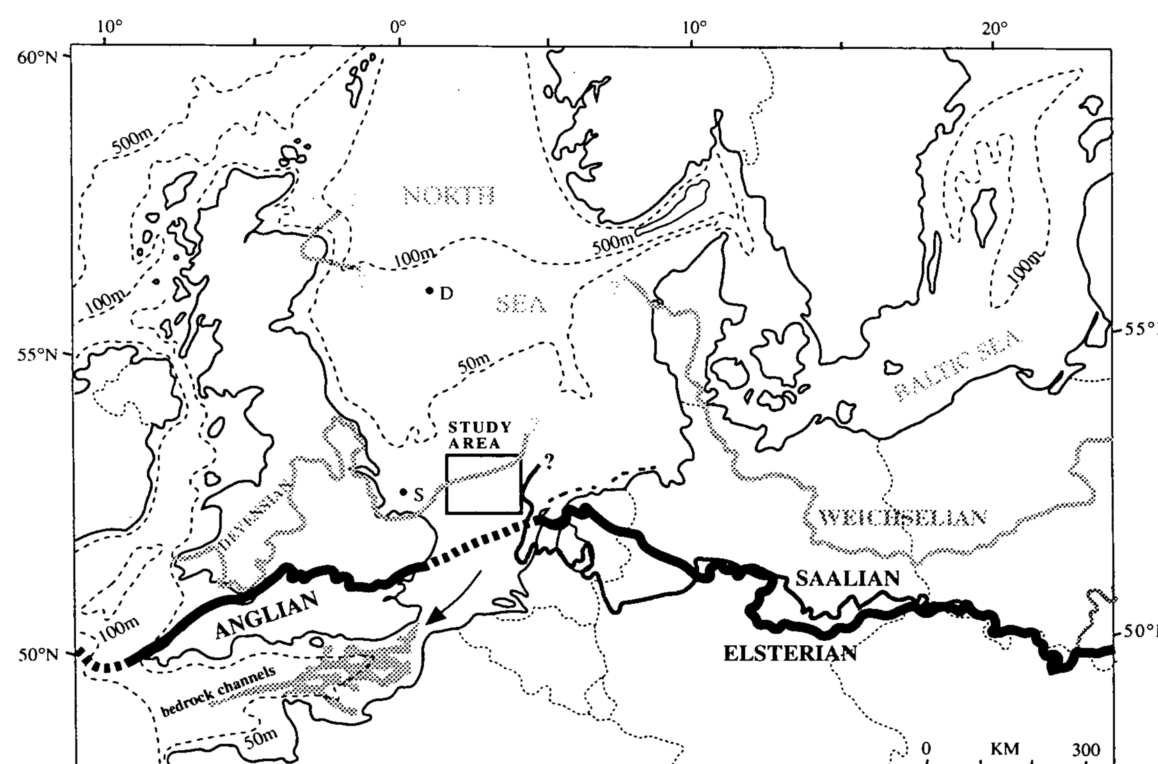


Figure 1.2 - Study area relative to Elsterian (Anglian) and younger glaciations of northwestern Europe. The extent of the Saalian and Weichselian glaciations across the North Sea is uncertain. Glacial limits from Graham and Shaw (1992), Long et al. (1988) and Ehlers et al. (1984). Channel system at the bedrock surface between England and France from Smith (1985). S=Silver Pit, D=Devil's Holes

1.1.2 b) method and contributions

The southern margins of the Elsterian glaciation are associated with a series of large (locally over 400 m relief) buried tunnel-valleys, known from borehole data in lowland areas from East Anglia (Woodland, 1970) to northern Germany and Poland (Ehlers et al., 1984). Those in the southern North Sea (Fig. 1.3) have been examined using a broad grid (5-15 km) of seismic reflection profiles collected by the British and Dutch Geological Surveys (Cameron et al., 1987, 1989b; Balson and Jeffery, 1991). Closer grids of 2D- and 3D-seismic data, supplemented by well data, are available as a consequence of hydrocarbon exploration. I obtained such data across a 100x150 km area (Fig. 1.2) and use it to present a three-dimensional analysis of these elongate glacial basins.

The results include a number of original observations regarding the physical characteristics of buried tunnel-valleys. I show that the southern North Sea examples converge southwards, but include younger arborescent elements to the north which result in local anastomosing patterns. This erosional morphology is comparable to that of tunnel-valley landforms of the last glaciation, interpreted as recording drainage to receding ice margins. I show that the fill is dominated by inclined reflecting surfaces indicating progradation to the north, prior to deposition of subhorizontal surfaces. Such an architecture is in contrast to previous seismic indications of a structureless lower fill. I establish that the fill consists of glaciofluvial sands overlain by lacustrine muds and show that the sands and muds do not correspond to separate seismic intervals as has previously been assumed. I show that sands recording glaciofluvial backfill are overlain by sands recording proglacial outwash in lake basins, a depositional geometry that may be applicable to other buried tunnel-valleys of northern Europe.

I use these results to present a model for the diachronous formation of the tunnel-valleys by the contemporaneous erosion and backfilling of sub-marginal basins in response to drainage through a receding ice margin. Prograding glaciofluvial fill surfaces are identified as backsets, interpreted as subaqueous outwash from a distributed system of subglacial channels. Form and fill characteristics indicate reworking beneath the outer 50 km of the ice sheet margin. Finally, I argue that surface water played an important role in recharging the subglacial drainage system during deglaciation, while substrata characteristics influenced the size and spacing of the tunnel-valley network.

I have presented these results and the model at conferences (Praeg, 1993; 1994; 1996) and in a brief contribution to a forthcoming seismic atlas of glaciated continental margins (Praeg, in press - a).

1.1.3 Thesis Outline

The thesis is divided into five chapters. The remainder of Chapter 1 is given to a review of studies of tunnel-valleys, followed by a description of the southern North Sea study area. In Chapter 2, seismic reflection evidence of tunnel-valley morphology and fill stratigraphy are presented and shown to indicate opposed processes of drainage erosion to the south versus prograding deposition to the north. In Chapter 3, borehole data on the physical nature of the fill are presented and used to show that sand-dominated glaciofluvial sediments are overlain by mud-dominated lacustrine to marine sediments. In Chapter 4, a model is presented for the sub-marginal formation of the tunnel-valleys and related to en- and subglacial drainage during deglaciation. A summary of the conclusions are presented in Chapter 5.

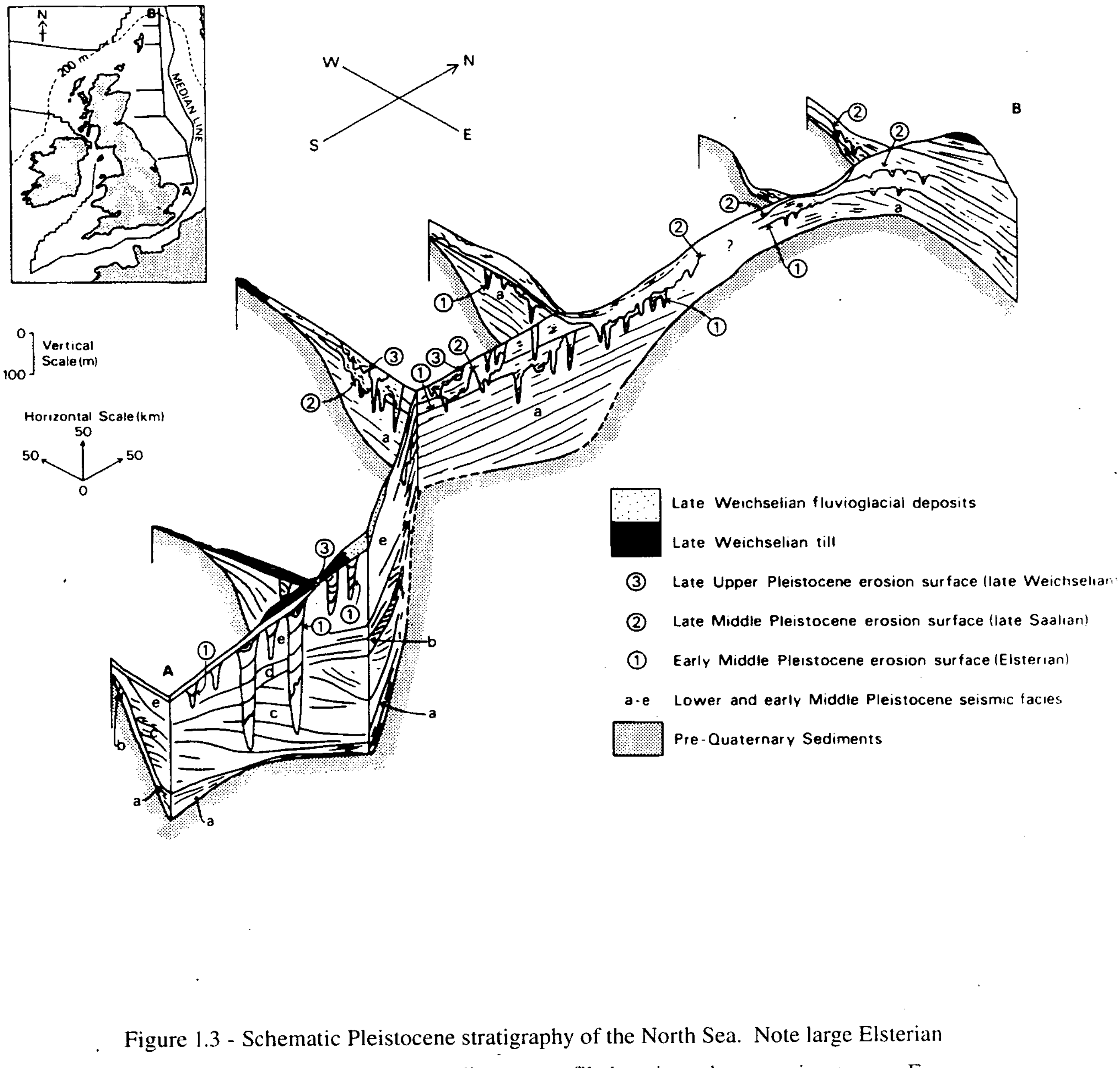


Figure 1.3 - Schematic Pleistocene stratigraphy of the North Sea. Note large Elsterian tunnel-valleys in the south. Fence diagram profile locations shown on inset map. From Cameron et al. (1987).

1.2 Previous Research

I review the literature on tunnel-valleys in order to establish what is known about their physical character and what has been thought in consequence about their formation. I give emphasis to the European setting, where the tunnel-valley concept originated and has been extended over wide areas, from lowlands to the continental shelf and to successive glaciations. The associated development of ideas on (and against) subglacial drainage is considered throughout, but a review of the hydrology of ice sheets is deferred to Chapter 4.

The original tunnel-valley concept represented a linking of erosional and depositional forms with inferred glacial processes: trenches (*Ger: rinnen*) marked by chains of lakes were associated with outwash and argued to record subglacial drainage under pressure, transverse to successive ice margins (Ussing, 1903-07; Werth, 1907-12). This model was extended across the last deglaciation of the southern Baltic lowlands (Woldstedt, 1913-29). The subsequent history of research in Europe has included disputes over the significance of water versus ice in the formation of tunnel-valleys and even doubts as to their glacial origin, which persist in some quarters. Extension to the offshore has been accompanied by proposals of a fluvial and marine origin during glacio-eustatic cycles. These discussions have generally appealed to combinations of glacial and non-glacial processes and have not resulted in an alternative model. The original association of tunnel-valleys with sub-marginal meltwater drainage remains valid and has been extended to at least two earlier glaciations of the continental shelf.

In North America, tunnel-valleys have been discussed largely independently of the European literature and interpreted only as records of subglacial meltwater drainage. They are known mainly as landforms, intially attributed to catastrophic subglacial outbursts through frozen ice margins, and later to areal (sheet) floods without reference to ice margins. Recent observations have suggested formation beneath receding ice margins during deglaciation. Surficial and buried features have been recognised on the eastern continental shelf.

I consider tunnel-valleys as landforms (section 1.2.1), as buried forms (1.2.2) and as features of the continental shelves (1.2.3). The southern North Sea setting is presented in section 1.3.

1.2.1 Lowland Meltwater Landforms

The literature of the turn of the last century attests to a robust investigation of processes marginal to ice sheets, by means of observations of modern glaciers and their application to fossil glacial landscapes of North America and Europe (see Price, 1973). Expeditions to Malaspina Glacier in Alaska were particularly influential in suggesting that water was an important element of glacier systems and that it flowed into the glacier and along its base through tunnels to emerge under considerable pressure (e.g. Russell, 1893; Tarr, 1909). Reconstructions of the last deglaciation assumed the dominant role of surface water and its ready access to the glacier bed beneath the margin, notably in interpretation of esker systems (e.g. de Geer, 1897; Hershey, 1897; Stone, 1899). This is the conceptual background against which the introduction of tunnel-valleys took place in the southern Baltic lowlands.

I describe a) the formulation of the tunnel-valley concept, b) controversies over their origin, c) axial relief and association with eskers and d) the extension to North America.

1.2.1 a) formulation of the tunnel-valley concept

Kozarski (1966/67) reviewed the history of ideas on the 'subglacial channels' of the southern Baltic lowlands. I draw on his paper in review of a discussion conducted mainly in German, as well as Danish and Polish, and on other English-language summaries in Charlesworth (1957, p. 241-243), Hansen (1971), Pasierbski (1979), Galon (1983) and Nilsson (1983). However, translations from German included below are mine, unless otherwise indicated.

Interest in networks of *Rinnen* (channels, trenches or gutters), marked by chains of elongate lakes (*Rinnenseen*), pre-dated general acceptance of the glacial theory in northern Europe. The landforms were initially regarded as surface expressions of underlying tectonic activity, an association which has persisted in this century (Kozarski, 1966/67 for a review). Following O. Torell's 1875 address to the *Deutsche Geologische Gesellschaft*, *Rinnen* and *Rinnenseen* were among features considered in subsequent addresses which embraced the theory of continental glaciation. Thus Berendt (1879) rejected his earlier advocacy for a tectonic origin and suggested that the trenches were eroded by glacial meltwater, flowing in open crevasses in the ice sheet. Geinitz (1886) suggested the action of meltwater flowing from the ice margin and eroding basins by evorsion.

Jentzsch (1884) was the first to invoke subglacial streams, filled with water moving under pressure (Kozarski, 1966/67). He contrasted the Alpine lake-basins, recently attributed to ice erosion by Penck, with the sinuous and convergent morphology of the *Rinnenseen*. and argued that 'the valley form must be a starting point for any proposed theory on the origin of these lakes':

> "die Thalgestalt muss einen Ausgangspunkt jeder über die Bildung dieser Seen aufzustellenden Theorie bilden" (p. 699).

He considered 'what force could interrupt the regularity of the valley-axes', which included reverse slopes of up to 14°, and suggested water moving uphill under pressure. He also appears to have equated the valleys with retreating moulins during ice margin recession (Charlesworth, 1957).

From 1903 N.V. Ussing, the director of the Danish Geological Survey (DGU), presented three publications containing influential observations on the relations of valleys containing 'long-lakes' (*Dan: langsøer*) to the outwash plains of Jutland (Fig. 1.4). Ussing (1903) observed that the outwash plains to the west of the Main Stationary Line (last-glacial maximum) were large conical flats and that their apices coincided with the mouths of valleys which breached the end-moraine belt at 'gates'. He concluded that the outwash plains had been formed by the outflow of subglacial streams. Ussing (1904, 1907) went on to consider the subglacial stream-beds, calling them fjord-valleys (*Dan: fjorddale*) due to their extension up to 75 km to an expression in the eastern coastline (Fig. 1.4). He noted their tendency to converge towards the former ice margin, their irregular axial profiles and the

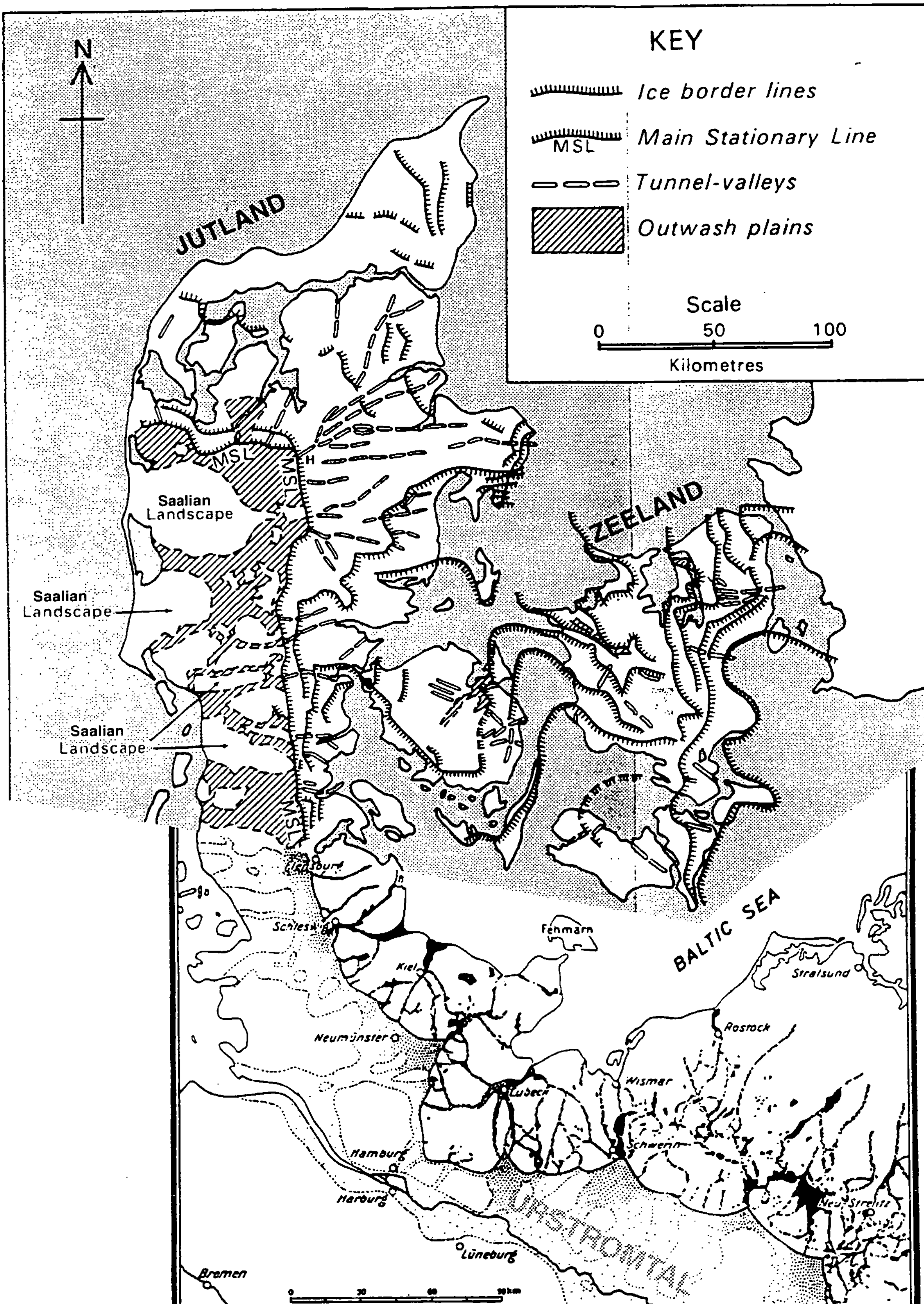


Figure 1.4 - Tunnel-valleys of the last glaciation of Denmark and northern Germany. The Danish area is taken from Woodland (1970, after Milthers and Nielsen), the German area from Nilsson (1983, after Woldstedt). The Main Stationary Line in Jutland records the maximum extent of the last ice sheet. Note tunnel-valley convergence to this line, and to younger ice margin lines including in Zeeland. In Germany, tunnel-valleys (including coastal Föhrden) are shown in black, outwash plains in stipple. Younger ice margin lines are not shown. H=Hald Sø (northern Jutland, original study area of Ussing, 1903-07).

elevation of the outwash cone apices over 50 m above the adjacent valley floors. He suggested that the water flowing in the subglacial valleys had moved uphill under pressure.

The role of pressurised subglacial streams was developed by Werth (1907, 1909, 1912) in a comprehensive treatment of glacial valleys of Norway, the northern Baltic (Sweden/Finland) and the southern Baltic (Denmark/Germany), which he classified respectively as *Fjorde*, *Fjärde* and *Föhrden*. The coastal terms from different languages (Norwegian, Swedish, German) were intended to apply to features both at and away from the coast and to reflect differences in form and process. Thus *Fjorde* occurred in mountainous areas of topographically constrained glacial erosion, *Fjärde* in low-lying areas of indurated rocks and *Föhrden* in sedimentary lowlands. The latter corresponded to the Danish *fjorddale* and German *Rinnen* and were attributed to meltwater erosion. Werth inferred erosion by water moving from high to low pressure (thicker to thinner ice) and so both uphill and transverse to former ice margin lines. He argued that pressurised water had determined not only the steep upward terminations of *Föhrden* at end moraines, but 'during separate phases of backwasting the irregular axial profile of all channel-forms of the formerly glaciated lowlands':

> "sowie mit Beziehung zu einzelnen Phasen des zurückschmelzenden Eises die Schwellenbildung bzw. das unausgeglischene Bodengefälle in allen den rinnenförmigen Senken der vergletschert gewesenen Tiefländer"
> (Werth, 1912, p. 164-165).

Werth's terminology was confusing, particularly the extension of coastal terms to inland features, and his classification was never fully accepted. However, his ideas on subglacial water were influential and his distinction of different glacier bed types relative to subglacial processes remains of interest. Ironically, his emphasis on the morphological similarities of fjords and tunnel-valleys was to be a persistent element of later arguments for erosion by ice.

Madsen (1921), who replaced Ussing at the DGU after his death in 1911, introduced the term tunnel-valley (*Dan: tunneldale*), as an alternative to *fjorddale* and *Föhrden*. The term was accepted within the Danish literature and, following Woodland's (1970) review, in English language publications. The DGU mapped tunnel-valleys across the last deglaciation of Denmark (Madsen, 1928) and during the war dedicated itself to a summary of the geology of Denmark, presented as a beautifully illustrated atlas (Schou, 1949). The latter includes block diagrams of tunnel-valleys emerging from the receding ice margin and maps of tunnel-valleys leading to successive marginal lines, which form the basis of the upper part of Figure 1.4.

Woldstedt (1913, 1923, 1926) extended Ussing's and Werth's concept of subglacial meltwater valleys across the *Jungmoränenlandschaft* or last-glacial landscape of the German and Polish lowlands. He noted the arrangement of trench-valleys (*Rinnentäler*) transverse (and often convergent) to former ice margin lines, including individual lobes, and to outwash cones and other outwash accumulations within the large proglacial meltwater valleys or *Urströmtäler* parallel to ice margins (Fig. 1.4). He attributed the *Rinnentäler* to erosion by pressurised streams beneath recessional stages of the ice margin. However, he suggested (1926) that wider features, such as the coastal Föhrden of Schleswig-Holstein, may have been remodelled at a late stage by glacial ice.

The sub-marginal meltwater model was widely applied, notably by Polish geologists (e.g. Pawlowski, 1927; Nechay, 1932; see Kozarski, 1966/67; Pasierbski, 1979; and Galon, 1983). The last-glacial landscape south of the Baltic is up to 350 km wide in Poland (see Fig. 1.2) and contains a number of prominent moraines marking recessional stages or readvances of the ice margin. Galon (1965) presented a map of tunnel-valleys in Poland (reproduced in Pasierbski, 1979), showing their change in orientation in correspondence with successive marginal lines, and noted examples of overlap of channels due to changes in ice margin orientation. The age of the outermost moraines is between 17-20 ka BP, while moraines along the southern Baltic coast are about 13-14 ka BP (e.g. Andersen, 1981), so that tunnel-valley formation across the area spans at least 3000 years of deglaciation.

Majdanowski (1949) used the distribution of tunnel-valleys, contoured as 'lake-channel density', to define the limits of the last glaciation across the European plain. His map implies that tunnel-valleys extend far eastwards from Poland, at least 1000 km to Leningrad. Lake-channel densities were negligible beyond the last-glaciation limit (see also Andersen and Borns, 1994, Figure 2-15B). Tunnel-valleys are not recognised as landforms within the *Altmoränenlandschaft* or old-moraine landscape, although Liedtke (1981) observed that the distributions of some unusually parallel streams are suggestive of former *Rinnen* and cited work on the Hümmlig of northwestern Germany.

1.2.1 b) controversies over water versus ice origin

In the first edition of his influential textbook Das Eiszeitalter (1929), Woldstedt expressed reservations concerning the origin of the *Rinnentäler*, suggesting that the valleys were largely eroded by ice, while meltwater was responsible for late-stage remodelling. In 1952, he published a paper in which he stated that a long period of investigation had persuaded him that the role of meltwater was very limited and that the valleys were the result of erosion by marginal ice lobes or tongues. This view was based primarily on the similarity of their long- and cross-sectional slopes to those of basins eroded by Alpine glaciers, and their narrowing near ice margin lines where the erosive power of water beneath a glacier should have been greatest. However, he suggested proglacial meltwater may have been significant in cutting depressions later shaped by ice tongues during the advance, in order to account for their valley-like plan form. Similar but less definite views were expressed in a later edition of Das Eiszeitalter (1954).

Woldstedt's reversal of opinion was significant, as his advocacy of glacial erosion found many supporters (see Kozarski, 1966/67). The extreme proponent was Gripp (1964, 1975), who proposed the term *Glaziellen* to refer to channels scooped out by narrow ice tongues during deglaciation, as a series of connected terminal basins, the irregular relief of which could also be due to melting of buried ice. Echoing Werth, he described them as the *Fjordtäler* of the glacial lowlands (Galon, 1983, p. 87-88). Liedtke (1975) more closely followed Woldstedt in promoting a combined glacial and glaciofluvial origin. The latter was invoked to account for the association of tunnel-valleys with outwash plains, and their sinuous plan form (e.g. Fig. 1.4), for which he suggested modification of glacially eroded forms by water flowing under pressure beneath the ice margin. Glacial erosion was invoked due to the imperfect association of outwash with tunnel-valleys, their 'bayonet-like' narrowing to ice margins and their similarity to Alpine basins, as per Woldstedt.

In Denmark, a school of thought formed around the consensus that Ussing's original concept, accepted until the 1960s, was inconsistent with both observations and theory. This was initiated by Hansen (1971), who argued that subglacial water should not run uphill. Hansen also suggested that the association of tunnel-valleys with outwash cones was inexact and that the lack of recognition of tunnel-valleys in reference works was 'peculiar' and indicative of the unreliability of the concept as a 'special Danish phenomenon'. His alternative was erosion by ice tongues as advocated by Woldstedt and Gripp in some cases and other explanations in other cases. The result of this 'case-by-case' approach in Denmark was summarised twelve years later by Krüger (1983). Berthelsen (1972) suggested the influence of selective proglacial meltwater erosion of permafrost, subsequently overridden and exploited by ice, as well as the formation and later decay of buried ice (cryolaccoliths) to account for the uneven long profiles. Kronborg et al. (1977) argued that straight, subparallel segments of adjacent tunnel-valleys, separated at angles, record neotectonic fault offsets and selective glacial erosion. Lykke-Andersen (1972) and Sjørring (1979) suggested the glacial modification of pre- or proglacial fluvial valleys. Marcussen (1977) suggested the valleys were elongate kettle-holes, possible located along the lines of older erosional valleys or tectonic lines.

The Danish setting is in fact characterised by the superimposition of last-glacial landscapes on relief of the penultimate glaciation (Fig. 1.4). Larsen et al. (1977) summarised evidence from eastern Jutland for the influence of remnant landscape elements on erosional and depositional patterns of the last glaciation: tills (and so relief) of Saalian age were found, but no earlier glacial or pre-glacial sediments. Lykke-Andersen (1986) presented borehole and geoelectrical evidence from the Nørreå valley in Jutland, one of the original tunnel-valleys studied by Ussing (1903), and showed it to be filled by up to 100 m of mainly glaciofluvial sands. He nonetheless argued for the glacial overdeepening of pre-glacial river valleys, of this valley and of all of the buried valleys of the northern European lowlands. The basis of this view was that "the classical conception of melt water flowing in tunnels at the base of the ice sheet towards the outwash plains... is unnecessarily complicated" (p. 211). Meltwater is known to flow in subglacial tunnels and glacial modification of pre-existing valleys is a classic example of an untestable hypothesis. Such views nonetheless persists in Denmark today in the extension of research into the offshore (section 1.2.3).

The arguments of Woldstedt and successors against a meltwater origin for tunnel-valleys were considered in some detail and quite effectively refuted by Kozarski (1966/67) and Galon (1983). They noted the rarity of marginal glaciotectonic disturbances and suggested that any local evidence for mechanical activity by glacial tongues could be attributed to the modification of features already formed by meltwater activity. They noted the small size of most tunnel-valleys relative to putative ice tongues and their sinuous or winding courses. Essentially, they reiterated the observational evidence of Jentszch, Ussing, Werth and Woldstedt himself regarding the morphology of the valleys, notably their arborescent convergence to successive ice margin lines and their association with outwash (Fig. 1.4), including eskers.

1.2.1 c) axial relief and association with eskers

The 'up-and-down' axial profiles of tunnel-valleys, expressed in lakes, have long constituted evidence of their subglacial origin, be it by the action of ice or water. Werth (1909, 1912) proposed an origin of the axial relief in outwash during ice margin recession, as noted above. He related the reverse slopes to deposition beneath successive margins, by extension from the outer margins where outwash cones apices are observed to be elevated above the floors of the valleys. Ussing (1903) had observed an elevation of over 50 m in northern Denmark, relative to the adjacent lake of Hald Sø (Fig. 1.4), and this elevation is 100 m when the depth of water in the lake is considered (Hansen, 1971). The elevation is over 150 m if the thickness of sediment in the lake is considered, which consists mainly of glaciofluvial sands (Lykke-Andersen, 1986). The presence of a glaciofluvial fill within the tunnel-valleys appears to be generally recognised in Germany (Grube, 1983). In Poland, Galon (1965) and Kozarski (1966/67) both observed that the transversal thresholds separating the lake-basins are composed of glaciofluvial deposits, i.e. the valleys are partly filled, indicating a possibly less irregular but even greater overall axial relief.

However, both Galon (1983) and Kozarski (1966/67) regarded the present relief as being due to the melting of buried ice. The idea of giant kettle-holes may have been first proposed by Ussing (1903), who suggested a lake 1 km wide and 6 km long (which he regarded as distinct from tunnel-valleys) to represent the 'fusion' of much smaller (<200m) dead-ice depressions observed in surrounding deposits. The idea was later applied to tunnel-valleys by Woldstedt (1926, 1929, 1952) and Gripp (1964, 1975), among others. Galon (1983) noted several examples of such an interpretation, including tunnel-valley lakes and adjacent kettle-holes in the Brda sandur in Poland, also illustrated in review by Ehlers and Wingfield (1991, Figure 2). In all examples, the lakes are at least an order of magnitude larger than dead-ice depressions known from Pleistocene or modern glacial settings. In addition, field evidence for 'disturbed deposits' is often open to dispute (e.g. discussion of Miedwie lake - Pasierbski, 1979). It is thus unclear to what extent melting of buried ice might have been of regional significance in the formation of the Rinnenseen. The alternative of depositional relief is seldom discussed.

Relief of depositional origin is also recognised in the form of eskers within tunnel-valleys. Galon (1965, 1983) cites and illustrates several examples from Poland. Mapping in Denmark has shown that eskers, found mainly in the east, are often associated with small tunnel-valleys (Krüger, 1983). Eskers are rare in northern Germany (Ehlers and Grube, 1983), but Grube (1983) notes an example of a small esker in a tunnel-valley northeast of Hamburg and Piotrowski (1994) notes examples farther north toward Kiel. The eskers are generally regarded as subsequent, superimposed landforms, localised by the tunnel-valley depressions. A more intimate relationship between eskers and the inland *Fjärde* of the southern Swedish peninsula was proposed by Werth (1914), who argued that erosion and deposition were contemporaneous, by application of de Geer's (1897) model of sub-marginal esker formation during deglaciation. The same approach was applied by Andersen (1931) in an interesting paper on tunnel-valleys and eskers in southeastern Zeeland (Fig. 1.4).

Andersen was addressing the issue of regional deglaciation by downwasting versus backwasting, a distinction introduced by Flint (1929), who proposed that subglacial meltwater activity would only be important in the case of stagnant (downwasting) ice, in order for channels to remain open. The concept of regional stagnation was strongly criticised, and effectively withdrawn by Flint and Demarest (1942) in favour of local marginal stagnation (see Price, 1973). Andersen (1931) considered the tunnel-valleys of Zeeland to be a priori evidence of stagnant ice, but within the terms of his discussion he presented evidence for their diachronous erosion during esker deposition.

Andersen examined an arborescent system of up to four tunnel-valleys, convergent to two outlets along a common transverse (marginal) line. He also referred to them as esker-valleys due to their discontinuous glaciofluvial fill, which he assumed was deposited by the same waters that eroded the valleys. He further argued that each of the four esker-valleys were "composed of similar successions of different links", the links comprising both arcuate valley segments (bows) and enclosed esker beads, and that these could be correlated between adjacent valleys across distances up to 50 km. Applying de Geer's model of beaded esker formation, Andersen interpreted the bows as annual and drew successive connecting 'winter-lines', constrained in places by marginal lineations. He regarded the winter-lines as recording the up-glacier migration of a boundary between active and stagnant ice.

Reverting to an ice marginal perspective, Andersen presented a case for the erosion of tunnel-valleys and their contemporaneous filling by glaciofluvial outwash, during ice margin recession. His explanation also accounts for the angular subparallelism of adjacent tunnel-valley segments, a factor in the tectonic theories of the previous century and regarded as evidence of faulting in Jutland by Kronborg et al. (1977).

In summary, tunnel-valleys were originally proposed to record the drainage of pressurised meltwater beneath the ice sheet margin during its recession. This model was based on their erosional morphology and an association with depositional landforms, including outwash fans, eskers and ice margin lines. This landform assemblage was recognised across the last deglacial landscape of the Baltic lowlands. Disputes over the meltwater origin of tunnel-valleys did not result in an alternative model. However, they may have contributed to the failure of the tunnel-valley concept to be disseminated outwith the European literature.

1.2.1 d) extension to North America

Tunnel-valleys were not recognised in North America in the first half of the century, nor was subglacial drainage considered as a significant erosional or depositional process. Ironically, the lack of interest was in part due to turn-of-the-century observations of the margins of Malaspina Glacier, which are covered with debris. Combined with Flint's (1929) proposals of regional stagnation, this contributed to a guiding perception that meltwater activity comprised areal wash of the ice margin surface (e.g. Newcombe and Lindberg, 1935). This idea has persisted in North American glacial geology, for example in the 'dirt machine' hypothesis of southeastern Laurentide deglaciation, discussed and refuted by Gustavson and Boothroyd (1987) in favour of glaciofluvial activity beneath the ice margin, once again on the basis of comparison with Malaspina Glacier.

Flint (1947, p. 167) first compared 'stream-trench systems' of the western Canadian plains to the more extensive marginal *Urstromtäler* and transverse *Rinnentäler* of northern Germany. Williams (1929) had commented on the complex trench systems of southern Alberta and Saskatchewan, emphasising their potential for reconstructing the retreat of the last ice margin across the area. This potential was exploited by Bretz (1943), who mapped former lakes and drainage routes, the latter including elements parallel and transverse to former ice margins. Bretz used the parallel elements to establish temporal constraints on intervening marginal moraines, but attributed the transverse elements to remnant pre-glacial drainage systems. Gravenor and Bayrock (1956) and Gravenor and Kupsch (1959) later followed Flint in comparing the stream-trench systems of Alberta with those of northern Europe, but interpreted the Canadian systems as 'ice-walled channels' formed in cracks beneath stagnant ice, similar to the views of Andersen (1931).

Wright (1973) identified elongate trenches in Minnesota as tunnel-valleys, following a visit to Denmark by E. Cushing. They were described as elongate trenches, marked by lakes, locally over 1 km wide and 70 m deep, cut through drumlins, containing eskers and oriented transverse to former ice margins. Wright regarded the subparallel, locally anastomosing, fan-shaped pattern as 'cut simultaneously along its length' by drainage across a 100 by 150 km area to a stationary ice margin, which marked the maximum extent of the Superior Lobe. He considered that surface water would not be able to penetrate to the bed to contribute to erosion and so attributed them to episodic drainage through a frozen margin. However, he allowed that the associated eskers may have been formed diachronously and by surface water as the ice thinned. The tunnel-valleys did not actually extend to the indicated ice margin, nor were terminal outwash accumulations observed.

Tunnel-valleys were subsequently identified along the last-glacial maximum immediately to the east in Wisconsin and locally to the southeast in northern Illinois and Indiana (Mickelson et al., 1983; Clayton et al., 1985). Attig et al. (1989) summarised mapping in Wisconsin, which showed that tunnel-valleys up to 500 m wide extend up to 40 km from the ice margin, in places with an arborescent pattern. The trenches cut through a marginal zone of hummocky moraine up to 20 km wide, led to outwash fans and contained small eskers, the latter more common farther from the margin in association with drumlins. They applied Wright's model of episodic drainage through a frozen margin, over a long period to allow for the proglacial accumulation of outwash, while the eskers were again allowed to form diachronously as the ice thinned.

Mooers (1989a, 1990) re-examined the Superior Lobe system and showed the tunnel-valleys and eskers to comprise segments 10-20 km long, associated with outwash fans along recessional ice margin lines. The eskers were interpreted to indicate the influence of surface water and the system was attributed to the headward development of the englacial drainage system within the outer 20 km of the ice sheet during deglaciation. Episodic release from subglacial reservoirs was thus shown to be both unnecessary and unfeasible in explanation of the system. Contemporaneous tunnel-valley erosion and fan deposition was implied, but the eskers were regarded as recording a 'later stage' of decreased discharge or increased sediment load.

Mooers did not attempt to estimate the duration of the deglacial recession across the area, but ice is thought to have stood at a maximum position between 15-18 ka BP and to have withdrawn to the north by 12 ka BP (Wright, 1973; Attig et al., 1989), suggesting tunnel-valley formation over several thousand years. Evidence of tunnel-valley formation over time is also available from the adjoining Des Moines lobe, which withdrew from northern Iowa across southern Minnesota between about 14-12 ka BP. Tunnel-valley segments 2-3 km wide and up to 15 km long are reported along at least two of the outer marginal lines of the lobe, corresponding to up to 1000 years of deglaciation (Patterson, 1993). Tunnel-valleys are also reported from the northern Des Moines lobe, possibly of similar age, and are argued by Hoffman and Cotter (1989) and Cotter (1993) to be inconsistent with the singular catastrophist model of Wright (1973).

A recessional model had previously been put forward by Christiansen (1987) in interpretation of a partly buried tunnel-valley of the last deglaciation of Saskatchewan. The Verendrye tunnel-valley (over 100 km long, about 1 km wide) terminates in a fan-shaped esker and is filled with over 200 m of sediments which fine upwards from sand and gravel to lacustrine muds. Christiansen proposed diachronous formation by headward erosion of a subglacial stream draining to a stationary ice margin. A fining-up sequence within a similar, partly buried tunnel-valley to the northwest in Alberta was also interpreted to record subglacial erosion and backfilling during deglaciation (McClung and Mollard, 1987).

Tunnel-valleys associated with drumlins and eskers have been widely reported from other parts of the Laurentide ice sheet, including Saskatchewan, Ontario and the high arctic, but have not been related to ice margins (Shaw, 1983; Shaw and Gorrell, 1991; Brennand and Sharpe, 1993; Brennand and Shaw, 1994). Instead, they have been interpreted to record the waning stages of subglacial sheet floods, invoked to explain the form of the drumlins. The complex plan-form morphology of tunnel-valleys is regarded as evidence of their synchronous erosion by subglacial flood waters, 'temporally disconnected' from subsequent glaciofluvial deposition. The complex tunnel-valley systems nonetheless contain arborescent esker systems and other glaciofluvial sediments, the sedimentology of which are interpreted to indicate surface-to-bed water connections within the outer part of the ice sheet. The deposits are acknowledged to be consistent in every respect with models of sub-marginal esker formation during deglaciation (e.g. arborescence, beads, fans and sedimentology - Bannerjee and McDonald, 1975), but such time-transgressive models are largely rejected in favour of synchronous erosional and depositional events due to flood events, including but not restricted to the seasonal cycle (Brennand, 1994).

The sheet-flood model is clearly incompatible with models for the sub-marginal formation of tunnel-valleys and eskers, such as those of Minnesota (Mooers, 1989a), which are also associated with drumlins. Mooers (1989b, 1990) presented a related model for the sub-marginal formation of the drumlins, based in part on correspondence of their orientations with those of the receding ice margin. He noted that an erosional relation of the tunnel-valleys to the drumlins was not necessarily easy to establish, as they were generally subparallel, and that zones of contemporaneous tunnel-valley and drumlin formation may have overlapped considerably.

Booth (1994) reported elongate troughs incised through proglacial outwash of the Puget Lowland, along the outer margin of the Cordilleran ice sheet in Washington. The troughs are up to 400 m deep and apparently (his Figure 4) locally over 10 km wide, but Booth compared them with large tunnel-valleys and argued for erosion by 'subglacial fluvial erosion'. He noted that such features challenge the usual assumptions regarding the importance of ice versus water in glacial erosion, a point earlier made by Ehlers (1981) in reference to buried tunnel-valleys.

1.2.2 Subsurface Meltwater Basins (Boreholes)

Borehole data from East Anglia to Poland have demonstrated the presence of buried tunnel-valleys, in the form of elongate depressions filled with glaciofluvial and lacustrine sediments, stratigraphically related to at least one glaciation prior to the last. An origin in meltwater drainage is inferred from their basinal relief, complex plan form and the character of the fill. The largest and best-known features are those of Elsterian age between East Anglia and Germany, and a detailed review of their fill character is presented in Chapter 3 by way of comparison with my results. Here I summarise the evidence of form and meltwater origin and review evidence for similar features to the east, from Poland to Siberia.

a) East Anglia to Germany

The buried valleys of East Anglia were compared to those of the Baltic Lowlands by Boswell (1914), who suggested they were "hollows analogous to the true Föhrden described so graphically and in so much detail by Dr. Werth" (p. 611). Woodland (1970) also compared them to the tunnel-valleys of Denmark, showing them to be elongate depressions locally over 100 m deep and up to 1 km wide. He found that they contained sands overlain by muds, both locally derived, and observed that their distribution was coincident with (and transverse to the edges of) the associated till sheet. He argued that the till and the tunnel-valleys had formed contemporaneously beneath the receding ice margin and that the valley fill sands recorded deposition from subglacial streams fed by water from the melting ice surface. The tunnel-valleys have since been related to the Anglian (Elsterian) glaciation (e.g. Ehlers and Gibbard, 1991) and a number of local borehole and geophysical studies have yielded information consistent with Woodland's observations (see section 3.5.1).

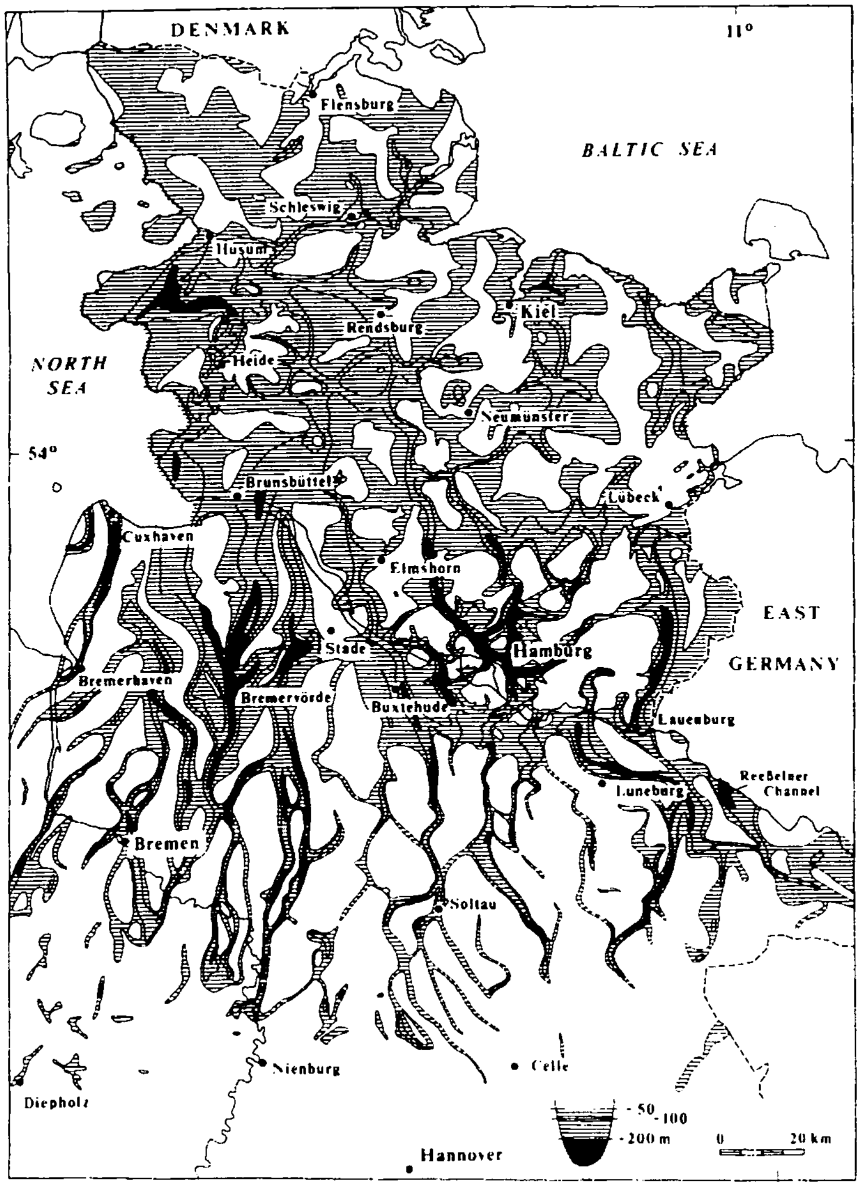
Buried Elsterian valleys in the northern Netherlands were described by ter Wee (1983), as depressions locally over 350 m in depth, filled mainly with sands and an upper layer of lacustrine muds, similar to the adjacent tunnel-valleys of Germany. Ter Wee preferred a southern origin, mainly due to the lack of Scandinavian pebbles in the sands, but was unable to account for their deep erosion by a non-glacial mechanism. The composition of the sands is also consistent with a northern origin (see section 3.5.1) and Ehlers et al. (1984) regarded the features as tunnel-valleys. Smaller buried valleys, up to 100 m in relief, were recognised within Pleistocene sediments in the eastern-central Netherlands and correlated to the Elsterian and Saalian glacial stages (van Rees Vellinga and de Ridder, 1973). The authors attributed them to fluvial erosion, but they are filled with glaciofluvial sediments and could also be tunnel-valleys.

In northern Germany, elongate depressions (Rinnen) incised into Tertiary sediments were recognised early in the century and initially attributed to tectonics or pre-glacial fluvial erosion (see Ehlers and Linke, 1989). All were eventually determined to be of Elsterian age by an upper fill of Lauenburg Clay and (Holsteinian) marine sediments. Their regional distribution beneath northwestern Germany was presented in a series of papers by Kuster and Meyer (1979), Grube (1979), Hinsch (1979) and Eissmann and Müller (1979), summarised by Ehlers (1981), Grube (1983) and Ehlers et al. (1984). They are locally over 400 m deep, 0.5-4 km wide and form an interconnected network mainly oriented transverse to former ice margins (Fig. 1.5a). The bulk of the fill is sands (Fig. 1.5b), with pebbles of northern derivation which demonstrate their glacial (i.e. non-fluvial) origin (Kuster and Meyer, 1979). They are interpreted as tunnel-valleys, formed by glaciofluvial erosion and deposition during the retreat stages of the Elsterian glaciation (Ehlers, 1981). A number of smaller (generally <100 m) features are filled with till and are suggested to be the result of ice rather than water erosion (Grube, 1979, 1983).

Ehlers and Linke (1989) presented borehole profiles along and across tunnel-valleys in Hamburg (e.g. Fig. 1.5b). An axial profile 30 km long indicated a wedge of interbedded sands and muds, thinning and dipping gently (≤1°) south, away from the former ice margin. They explained the interbedding by episodic release of meltwater in the marginal areas of the ice sheet, from en- or subglacial reservoirs, which also eroded the channels. Water due to melting of the ice sheet during deglaciation was related to later depositional stages. They suggested that the large valleys would have been filled with ice and drainage would have taken place along a narrow channel at the base. This was argued to be consistent with evidence of horizontal stresses along some buried valley flanks (Bruns, 1989) and of rafts of glaciotectonically dislocated Tertiary sediments within the fill.

Piotrowski (1994) also presented borehole profiles across and along tunnel-valleys in the north of Germany, south of Kiel. He suggested that a single polygenetic feature recorded glacial erosion and deposition during three successive glaciations. His profiles show that an Elsterian valley up to 200 m in relief is present, as shown by Hinsch (1979). However, his attribution of the upper part of the fill to the Saalian glaciation is based only on identifications of till, and sands overlying clays, both features of the fill of Elsterian tunnel-valleys beneath Hamburg (Ehlers and Linke, 1989). The Weichselian glaciation is represented by the upper tens of metres of section, including *Rinnenseen* which coincide with the buried valley location but do not appear to be determined by it.

a)



b)

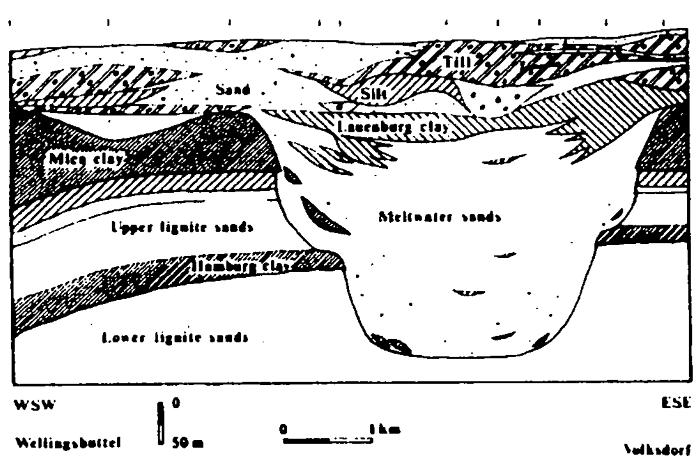


Figure 1.5 - a) distribution of Elsterian tunnel-valleys in northwestern Germany, based on depths to the base of the Quaternary in boreholes (from Ehlers et al., 1984); b) cross-section of a tunnel-valley in Hamburg, incised over 300 m into Miocene sediments (from Ehlers, 1981).

Piotrowski (1994) suggested that the location of the buried tunnel-valley was controlled by flanking salt structures within the incised Mesozoic-Cenozoic substrate. The regional correspondence of some Elsterian tunnel-valleys with the trends of underlying salt ridges was also noted by Hinsch (1979). Ehlers *et al.* (1984) observed that while examples of local control by near-surface salt domes exist, the regional correspondence is approximate and poorly constrained. Ehlers and Linke (1989) further observed that the course of major features appear to be unaffected by halokinetic structural trends. Comparison with features of the last glaciation would suggest that the ice margin was the primary control on orientation, but ice margin lines of the Elsterian deglaciation are unknown.

b) Poland to Siberia

Buried valleys have a substantial literature in Poland (Mojski, 1985). They contain sands locally intercalated with tills, and were traditionally regarded as fluvial in origin, by erosion and deposition during successive glacial-interglacial cycles. Mojski (1981) compared them to the surficial features of the last glaciation and argued that they were tunnel-valleys formed during at least two preceding glaciations, filled with glaciofluvial sediments. There are doubts as to their formation during more than one glaciation, an hypothesis based on tills within individual valleys (Ehlers et al., 1984). Contours on the base of the Quaternary presented by Mojski and Rühle (1965) show valleys locally over 150 m below sea level (bsl) near the Baltic coast, while recent work showed somewhat greater relief (up to 190 m bsl - Marks, 1988). The earlier map was reproduced by Ruszczynska-Szenajch (1993, Figure 2), who noted the occurrence of rafts of Tertiary strata within one such valley, described however as a depression of glaciotectonic origin. Marks (1988) also reflected uncertainty as to the influence of tectonics, fluvial processes and glacial activity in the formation of the valleys.

Buried valleys are recognised east of Poland, at the surface of the Palaeozoic strata in Belarus where they are up to 100 m deep (Matveyev, 1995), and in the Baltic States (e.g. Miidel and Raukas, 1991, Figure 16.1) where they are locally over 300 m deep and filled with glaciofluvial sediments, tills and lacustrine muds (Raukas and Gaigalas, 1993). A single system at the bedrock surface is attributed to successive glaciations, younger to the north, and it is not clear to what extent this might reflect stratigraphical differentiation based on tills alone, as in Poland, or a real overlap and so polygenetic features.

Buried tunnel-valleys may be widespread in the glaciated lowlands of Russia and adjoining regions, as indicated by a selection of papers which describe features at the base of the Quaternary which are attributed to lower to mid-Pleistocene glaciations. In southwestern Timan, buried valleys locally over 100 m deep are incised into Mesozoic strata (Potapenko, 1979). The author distinguished troughs of glacial erosion, up to 5 km wide and filled with morainal material, from troughs of meltwater origin, up to 2 km wide and filled with sands (and locally rafts of incised strata), and referred to work on the latter features by Spiridonov and Mislevets (1977). In the southern Pechora Lowlands, buried valleys up to 100 m deep and filled mainly with glacigenic sands, are incised into Palaeozoic strata (Simonov and Stepanov, 1988). In the Ukraine, Khristophorova (1995) described an overdeepened valley up to 80 m deep, filled with glacigenic sediments, incised into Mesozoic strata, and referred to others. In western Siberia, overdeepened valleys up to 340 m bsl, incised into Mesozoic-Cenozoic strata, are argued to be of glacial origin (Astakhov, 1991).

In summary, buried tunnel-valleys from East Anglia to Germany have been shown to contain a sandy glaciofluvial fill and to be related to meltwater drainage beneath the margins of the Elsterian ice sheet (Woodland, 1970; Ehlers, 1981; Ehlers and Linke, 1989). Equivalent features of at least one glaciation may be widespread across formerly glaciated areas to the northeast.

1.2.3 Continental Shelves

Submarine tunnel-valleys were first identified as bathymetric deeps on the floor of the North Sea, by comparison with landforms of the Baltic lowlands. Following the advent of seismic reflection profiling in the 1960s, multiple levels of buried valleys were recognised beneath the North and Irish Seas and correlated to successive glaciations. Their origin has been controversial, due to attempts to interpret some of them in terms of fluvial and marine processes and to comparisons with similar features of non-glacial origin in the English Channel. They have recently been argued to record sub-marginal drainage during deglaciation, either catastrophically or diachronously. Tunnel-valleys have been reported from other glaciated continental shelves, notably offshore eastern Canada where they are interpreted in similar terms.

I describe a) the UK North Sea and Irish Sea, b) the German/Danish North Sea, c) the English Channel and d) other shelves.

1.2.3 a) UK North Sea and Irish Sea

> "There are, indeed, some long trough-like cavities in the North Seas, named the Silver Pits, but they present us with nothing like a system of valleys resembling those on dry land" (de la Beche, 1834, p. 190).

Elongate holes, deeps or pits in the bottom of the sea have long been known to fisherman and cartographers. De la Beche may have been the first geologist to propose an origin, as cracks due to upward crustal bending.

A century later the deeps were attributed to fluvial processes, influenced to varying degrees by glaciation. Gregory (1931) and Beaugé (1937) reported soundings of the Devil's Holes (see Fig. 1.2) and adjoining steep-sided trenches in the northern North Sea and considered them to be traces of the valley of the Rhine. Lewis (1935) recognised additional elongate deeps and tried to account for them by a river valley which changed position over time. He suggested their isolation to be of tectonic origin, the result of drainage disruption and overflow across former rising anticlines. H. Baulig (in Guilcher, 1958, p. 208) considered a similar process, due to glacio-isostatic uplift of areas beyond the last ice margin. Robinson (1952) suggested that an ice margin in the North Sea would have influenced fluvial drainage, by analogy with 'urstromtal' valleys of Germany, and proposed the deeps to result from partial fill of valleys due to dead ice which later melted.

Furnestin (1939-43) may have been the first to propose a subglacial origin. He reproduced Beaugé's 1937 soundings of the Devil's Holes - steep-sided basins up to 4 km wide, 40 km long and 150 m in relief, with longitudinal basins and sills - and observed that such features could not be interpreted as fluvial ("ceux-ci ne sauraient, en effet, être considérés comme un ancien delta fluvial" - p. 486). He suggested that they presented a glacial aspect comparable to fjords and related them to erosion by a glacier from the Firth of Forth.

Valentin (1957) extended the tunnel-valley concept of the Baltic lowlands to the submerged glacial landscape of the North Sea. He identified a belt of *Rinnen* along the eastern coast of England, north of East Anglia, and traced their line of termination offshore to the southern limit of the belt of elongate deeps which include the Silver Pit (see Fig. 1.2). He attributed the system to sub-marginal drainage transverse to a lobate recessional margin of the last ice sheet. This model was subsequently applied by Flinn (1967) to an arcuate belt of linear deeps in the northern North Sea, including the Devil's Holes. He used them to infer a former ice margin, suggesting that "they formed under the front of an ice sheet as siphons through which melt water escaped under pressure" (p. 1152).

Donovan (1965) used sample and seismic data to examine the Silver Pit (up to 100 m deep, 4 km wide and 35 km long) and showed it to be eroded through last-glacial boulder clay. However, he interpreted it and adjacent deeps as post-glacial tidal scour hollows, as had Guilcher (1951) based on soundings. Robinson (1968) observed that this idea was difficult to apply to other, even larger trenches such as the Devil's Hole. He favoured Valentin's hypothesis of subglacial erosion and suggested that unrounded gravels and sands at the 'mouth' of the Silver Pit represented glaciofluvial outwash. Donovan (1973) reconsidered the tidal scour hypothesis, noting the lack of modern day analogues, but allowed that subglacial tunnel-valleys may have been enlarged by post-glacial tidal scour.

Dingle (1971) presented the first seismic reflection evidence for buried tunnel-valleys, beneath a 60 by 100 km area northwest of the Dogger Bank. Channels 1-5 km wide incised up to 200m into Tertiary strata, filled with concave reflectors, were of undulating axial relief, with a plan form generally too complex to trace across the 4-10 km grid. Dingle compared them to the tunnel-valleys of Woodland (1970) and followed his interpretation of subglacial streams near the edge of the last ice sheet, which both eroded channels and filled them with outwash sands and clays. He raised the question of why others are not filled, for which he saw 'no obvious explanation'. Oele (1971a) also presented seismic evidence for buried channels, up to 25 m in relief, incised into meltwater clay beneath the southeast Dogger Bank and attributed to a preceding (Saale) glaciation.

Jansen (1976) presented the results of a reconnaissance survey of the central and northern North Sea between 55-60° N, based on a broad grid (10-30 km) of acoustic profiles that yielded up to 50 m of penetration. He recognised a system of buried channels (3-6 km wide, over 50 m in relief) which he suggested were traceable over distances of up to 240 km. The southernmost were coincident with those of Dingle (1971), untraceable across a closer grid. The western channels extended northwards from the Devil's Hole and related deeps, with similar orientations and dimensions. Jansen recognised a connection, but suggested that while the deeps were tunnel-valleys, they marked the northern limit of ice, and the buried channels recorded proglacial drainage. He suggested three acoustic phases of fill, which he related to drainage along the same lines during successive glaciations.

In the mid-1970s, offshore exploration fostered an increase in the amount and quality of subsurface information and buried channels interpreted as last-glacial tunnel-valleys were reported from a number of sites in the central and northern North Sea (e.g. Caston, 1974; Eden, 1975; Milling, 1975; reviewed by Caston, 1979). Milling presented contours on buried channels from a 13 by 14 km site in the Norwegian sector (block 25/11), showing them to be up to 3 km wide and 80 m in relief, with an arborescent organisation convergent to the north. Kunst and Deze (1980) presented contours from a 17 by 7.5 km site in the central North Sea, using a 300-600 m grid of seismic lines. They recognised three levels of channel up to 130 m deep and 1-2 km wide, with irregular axial relief (the middle converging to the northwest), which they interpreted as tunnel-valleys formed during successive glacial cycles, following Holmes (1977).

Holmes (1977) and Thomson and Eden (1977) presented the results of a systematic seismic survey of the upper 200-300 m of the central North Sea. Three levels of channeling were recognised, the upper including those of Jansen (see Caston, 1979). They could not be traced in plan with profile spacing of 10 km (closer than that available to Jansen), but seabed deeps such as the Devil's Hole were seen to correspond to an incomplete fill of the upper channels. The lower channels, up to 200 m in relief, were contoured within two small areas of closer line spacing (≤1 km), South Fladen (11 by 18 km - Holmes, 1977) and the Forties field (6 by 7 km - Caston, 1977b, 1979) and characterised as branched or anastomosing depressions (0.5-3 km wide) of irregular axial relief. Holmes (1977) concluded that the morphology of the channels was consistent with successive levels of tunnel-valley incision. Thomson and Eden (1977) concurred, but suggested tidal scour may have enlarged the seabed deeps.

The fill of the lower and upper channels was seismically characterised by apparently structureless material, in places including an upper interval of reflectors. Two partial-recovery boreholes into the upper and lower channels (BGS 75/29 and BP-DB6, respectively) indicated a preponderance of fine sand, with marine microfossils. The bottom of 75/29 also reached the lower channels and proved barren sand. Holmes acknowledged the possibility of reworking, but proposed a complex genesis in which some of the lower channels were entirely filled by marine sediments while others may have been filled by glacial deposits. In contrast, Caston (1977b, 1979) interpreted the preponderance of fine silty sand in BP-DB6 to indicate glaciofluvial deposition during ice retreat, possibly "by the very streams which cut the tunnel valleys" (1977b, p. 21). A borehole through the lower channels to the west (BGS 74/14, Thomson and Eden, 1977) was dominated by diamict but included 2 m of basal sandy gravel also interpreted as a subglacial stream lag deposited during channel erosion.

Holmes (1977) had attributed the three levels of channeling to successive stages of the Weichselian glaciation based on radiocarbon dates. Stoker et al. (1985a, b) presented a revised stratigraphy for the central North Sea based on new borehole data in which they identified the

Brunhes-Matuyama magnetic boundary (*ca.* 0.73 Ma) in sediments below the channels. Three boreholes (81/29, 34 and 37) were said to penetrate the lower and middle channel deposits (now the Ling Bank and Coal Pit formations) and to contain glaciomarine muds, with interglacial foraminiferal assemblages in the lower fill indicating that each spanned a glacial-interglacial cycle. Stoker et al. correlated the successive levels of channeling with the three continental glaciations of northern Europe (see Fig. 1.3). However, this was accompanied by an argument that ice sheets had not been extensive in the North Sea, due to the lack of associated till sheets (Stoker and Bent, 1985), and so the channels were not of subglacial origin. A proglacial origin for the tunnel-valleys was subsequently developed in summary papers on the North Sea by Long and Stoker (1986), Cameron et al. (1987) and Long et at. (1988).

Long and Stoker (1986) reviewed work on the open and buried channels of the central North Sea, and of the southern North Sea where regional mapping had shown buried channels up to 400 m in relief (section 1.3). Cameron et al. (1987) and Long et al. (1988) extended the review to the northern North Sea (see Fig. 1.3). Three seismic fill types were suggested to be common to all the channels: basal structureless, middle layered, and upper inclined or variable. (Seismic profiles presented by Bent (1986) from buried valleys in the central North Sea include other arrangements of structureless and layered seismic facies.) The limited borehole information was used to infer geographical variability in fill character. Large (Elsterian) valleys in the southern North Sea were unsampled, but suggested to have a subglacial to lacustrine fill, while the three generations to the north were mainly filled by proximal to distal glaciomarine sediments, based on up to six boreholes in the central North Sea. A few boreholes of the lower and upper channels in the northern North Sea were said to contain mainly sandy fills, in the case of the upper channels of possible glaciofluvial character (Cameron et al., 1987, p. 54). Reference was not made to the sandy channel fills described as glaciofluvial by Caston (1977, 1979) in the central North Sea.

It was concluded that the Elsterian channels of the southern North Sea were tunnel-valleys, recording meltwater drainage through the ice sheet margin, and that a similar origin might apply to other channels within the mapped limits of offshore till (Fig. 1.3). However, the majority of channels, of all generations, lay beyond the till limits and so were interpreted as fluvial channels modified and locally overdeepened by proglacial processes, including 'cataclysmic outbursts' of glacial meltwater and later tidal scour (Long and Stoker, 1986a; Cameron et al., 1987).

Wingfield (1989, 1990) addressed the origin of the multiple generations of buried channels in the North and Irish Seas and the English Channel, referring to them as 'major incisions' (distinct from 'minor incisions' of less than 100 m relief). He suggested the term incision was non-genetic and applied it to features incised into a wide range of sedimentary and non-sedimentary substrata. Nonetheless, he regarded their similarity of form and (seismic) fill character as evidence of a common glacial genesis. He proposed repeated outbursts of subglacial lakes through successive ice sheet margins during deglaciation, a modification of the flood model of Wright (1973). Incisions recorded the upstream migration of plunge-pools, backfilled by the eroded sediment, while open features were giant kettle holes, formerly filled with ice. Wingfield (1989) applied this model to the glaciations of the North and Irish Seas, inferring ice margin lines from the distribution of tunnel-valleys. His broad definition of 'incision' implied glaciation of the English Channel (see below).

Ehlers and Wingfield (1991) compared the Weichselian morphology and stratigraphy of the North Sea with that of northern Germany and Denmark and argued that partly-filled 'incisions' in both areas are equivalent features formed by meltwater processes. They noted that tills in the terrestrial areas tend to be thin and discontinuous near ice margins, and eroded by successive glaciations, and argued that the lack of till in the North Sea stratigraphy is not a reliable indicator of limited ice extent. They proposed that the tunnel-valleys themselves are indicative of meltwater erosion beneath former ice margins, be it in the diachronous manner proposed by Valentin (1957) in the southern North Sea, or in the catastrophic manner envisaged by Wingfield (1990) for successive glaciations of the continental shelf.

Whittington (1977) had presented seismic profiles from the central Irish Sea and showed enclosed deeps to be the unfilled portions of a more extensive, southward convergent system of valleys up to 100 km long, 1-3 km wide and over 100 m in relief, incised through a last-glacial till sheet. He suggested a subaerial fluvial origin, but allowed the possibility of subglacial initiation as tunnel-valleys. Regional mapping, summarised by Wingfield (1989), subsequently identified at least three levels of 'incision', the upper coincident with a number of enclosed deeps. They were attributed to the Elsterian to Weichselian glaciations. Mapping in the southern Irish Sea has shown the presence of four generations of incision, the lower considered to pre-date the Elsterian (BGS/GSI, 1991).

1.2.3 b) German/Danish North Sea, the Baltic

C. Schwarz (written communication and unpublished manuscript, 1994) compiled seismic reflection data from the German sector of the North Sea and recognised the presence of tunnel-valleys with depths comparable to those mapped beneath northern Germany and the Netherlands (up to 400 m). It was not possible to differentiate generations of channeling and broad line spacing limited interpolation to plan.

Salomonsen and Jensen (1994) presented reflection and borehole evidence from the Danish sector of the North Sea for buried valleys 0.5-3.5 km wide, extending up to 200 m below the sea floor, which contained discordant reflectors interpreted to record four successive erosional events along the same lines. It was not possible to trace the channels in plan and there was no downhole information on their fill. The westernmost features coincide with the Weichselian 'incisions' of Wingfield's (1990) case study in the UK sector and the general similarity to tunnel-valleys was acknowledged. However, they were interpreted as a system of fluvial valleys incised during glacial lowering of sea levels since the Saalian, modified by proglacial meltwater drainage.

An interpretation of proglacial meltwater erosion was also applied to buried valleys of Weichselian age in the Kattegat, between Denmark and Sweden, by Lykke-Andersen et al. (1993) and Glydenholm et al. (1993). The authors presented seismic reflection evidence (5-10 km grid) of buried valleys 1-2 km wide and up to 100 m in relief, of irregular axial relief, with a complex plan form. The valleys were incised into a sequence of lower to mid-Weichselian deposits and their fill inferred to be waterlain. However, the authors suggested that glacial ice had never extended across the area, describing evidence for the last glaciation of Jutland (Fig. 1.4) as 'an assumption based on relatively weak evidence' (Lykke-Andersen et al., 1993). The channels are obviously offshore equivalents to the tunnel-valleys of Jutland originally described by Ussing (1903).

Sviridov et al. (1977) presented contours on the bedrock surface of the Baltic Sea and adjoining areas, based on a broad grid of seismic profiles (Sviridov, 1976, Figure 1). Buried valleys up to 200 m in relief and 4 km wide were recognised, including offshore extensions of the buried tunnel-valleys of Poland mapped by Mojski and Rühle (1965).

1.2.3 c) English Channel

Bathymetric deeps and buried valleys in the English Channel have been suggested to be tunnel-valleys. However, there is evidence against ice sheets having reached the Channel (see Fig. 1.2) and other mechanisms have been proposed to explain the valleys. They have nonetheless contributed to doubt over the origin of tunnel-valleys in glaciated regions of the offshore (e.g. McCave et al., 1977; Hamblin, 1991). I briefly review the Channel setting in order to show that the buried valleys do not require a subglacial formation and are not tunnel-valleys.

A network of mainly buried valleys, incised into Mesozoic-Cenozoic strata, extend 200 km westwards from the Dover Strait with a sinuous anastomosing pattern and culminate in the Hurd Deep (Fig. 1.2). The latter is up to 100 km long and 6 km wide and seismic reflection profiles show it is partly filled with sediment and up to 150 m in basinal relief (Hamilton and Smith, 1972). Hamilton and Smith proposed the Deep to result from a combination of fluvial processes and tidal scour during successive glacial cycles, with fluvial drainage enhanced by meltwater supply through the Dover Strait from both former ice sheets to the north and from the glaciers of the Alps to the south. Dingwall (1975) extended this explanation to the buried valleys to the east, noting that abandonment of successive valleys would have contributed to a complex, composite network. These ideas were supported in a regional synthesis by Auffret et al. (1980), who also noted that many of the valleys extend seaward from the modern river systems.

Kellaway et al. (1975) argued that the deeps and buried valleys formed by subglacial drainage during one or more glaciations of the Channel. Destombes et al. (1975) argued that buried channels in the Strait of Dover represent tunnel-valleys excavated by two different glaciations, an earlier one which extended up the Channel from the west and a later (Weichselian) one from the North Sea. Wingfield (1989, 1990) argued for a Weichselian glaciation of the western Channel based on the presence of bathymetric deeps. There is no depositional evidence for glaciation of the coasts of either southern England or northern France (Fig. 1.2) and the proposed glaciations of the Channel have been refuted (Kidston and Bowen, 1976; see Hamblin et al., 1990).

Smith (1985) postulated that the channels of the Dover Strait and English Channel were the result of a catastrophic flood from a proglacial lake dammed between an ice sheet to the northeast and a chalk barrier across the Straits, possibly as early as the Anglian/Elsterian glaciation. This model of extraordinary fluvial erosion was supported by analogy to the erosional channels and scour basins formed by floods from glacial lake Missoula in the western United States. Gibbard (1988) noted that the latter were the result of repeated floods and suggested that although the overspill of such an ice-dammed lake may have initiated the Dover Straits, the bedrock channels were thereafter eroded by a combination of enhanced fluvial erosion and tidal scour as proposed by Hamilton and Smith (1972).

The channels cut into the bedrock surface in the English Channel represent a polygenetic system that has developed over successive glacial-eustatic cycles by a combination of (glacio-) fluvial and marine (tidal) processes. They are quite distinct from the multiple channels of the North and Irish Seas, which record subglacial drainage during successive continental glaciations.

1.2.3 d) other shelves

Tunnel-valleys may be more common within glaciated continental shelves than has been recognised, as suggested by their recent identification offshore both eastern Canada and Antarctica.

On the Novia Scotian shelf, buried valleys were identified on regional seismic profiles by King et al. (1974) and attributed to glacial modification of pre-existing fluvial valleys. Boyd et al. (1988) used close grids of seismic data to demonstrate their anastomosing plan form and irregular axial relief, with depths up to 300 m and widths of 2-3 km, and proposed them to be tunnel-valleys. Downhole information from overlying sediments indicated incision prior to, or early in, the last glaciation, but no samples were available of the valley fill. Boyd et al. suggested erosion during the waning stages of a catastrophic sheet flood from subglacial lakes.

Barnes and Piper (1978) had proposed that channels of irregular axial relief cut through till on the innermost Scotian Shelf represented *Rinnentäler*. Loncarevic et al. (1992) subsequently interpreted complex seabed relief farther offshore as a complex system of partly filled tunnel-valleys, some of which may be correlative to those of Boyd et al. on the outer shelf and others (unfilled) which may be younger. Complex cross-cutting plan forms were inferred to record multiple drainage episodes under different ice sheet configurations. They noted the presence of equivalent (buried) features elsewhere on the Scotian Shelf and Grand Banks. They suggested that more information was required on the fill in order to distinguish between catastrophic or diachronous drainage processes.

On the Antarctic shelf, Alonso et al. (1992) reported buried erosional surfaces at multiple levels within the Pliocene section of the eastern Ross Sea shelf, including cross-cutting features up to 2 km wide which they suggested to be tunnel-valleys. Lien et al. (1989) described 'canyon-like' depressions on the floor of the Weddell Sea, up to 60 m deep, 1 km wide, incised into unlithified sediments in 300-350 m of water, and suggested an origin in subglacial meltwater activity.

Wingfield (1990) suggested examples of 'major incisions' on other glaciated continental shelves, although some are much broader than tunnel-valleys (10-30 km) and may be better described as transverse troughs. extending seaward from fjords (e.g. Herzer and Bornhold, 1982 - western Canadian shelf; Carlson et al., 1982 - Alaskan shelf). Some of the open and buried valleys of the northern Eurasian shelf described by Lastochkin (1977) appear similar in depth and form to tunnel-valleys, although he did not mention their width. Astakhov (1991) also referred to offshore extensions of the buried glacial valleys of western Siberia.

In summary of previous research, over the course of the century since their initial formulation as meltwater landforms tunnel-valleys have come to be recognised as characteristic elements of the glaciation of sedimentary lowlands and continental shelves. Their physical characteristics have repeatedly been shown to indicate an association with drainage beneath ice margins during deglaciation, but these studies have not resulted in a consensus regarding their origin. In northern Europe, tunnel-valleys have been recognised as buried features related to at least three glaciations prior to the last. The largest known are those of the mid-Pleistocene Elsterian glaciation which extend across the southern North Sea (Fig. 1.3).

1.3 SOUTHERN NORTH SEA SETTING

In this section, I describe the Quaternary stratigraphy of the southern North Sea (1.3.1), and the palaeogeography at the time of the Elsterian glaciation (1.3.2).

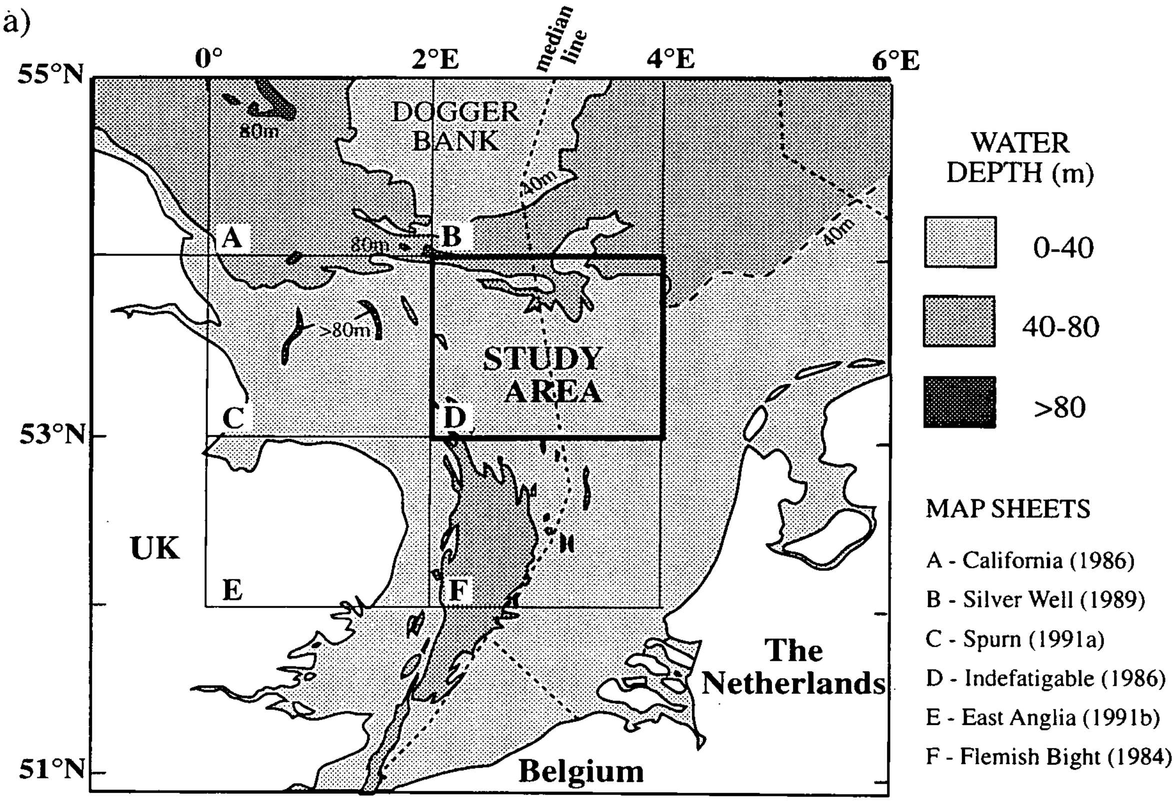
1.3.1 Quaternary Stratigraphy

The Quaternary geology of the southern North Sea has been mapped over the last decade by the British and Dutch Geological Surveys (BGS/RGD) and presented in a series of 1:250,000 maps (Fig. 1.6a) and related publications (Laban et al., 1984; Balson and Cameron, 1985; Cameron et al., 1984; 1987, 1989a, 1992; 1993; Long et al., 1988; Balson and Jeffery, 1991). The thesis study area corresponds to the Indefatigable map sheet (BGS/RGD, 1986). Water depths across the area are mainly between 20-40 m bsl and Quaternary sediment thicknesses increase to over 500 m (Fig. 1.6).

BGS/RGD mapping methods were outlined by Cameron et al. (1989a,b; 1992). The stratigraphical framework is based on seismic 'elements' transformed or subdivided into sedimentary formations using downhole data (Figs. 1.7, 1.8). Chronostratigraphy is based on correlation to the pollen-zone climatic stratigraphy established for the Netherlands and dated by palaeomagnetic and radiometric means (Zagwijn, 1985, 1989). The base of the Pleistocene is considered to be about 2.3 Ma (Cameron et al., 1992). The Brunhes-Matuyama palaeomagnetic boundary (0.73 Ma) has been identified within pre-glacial sediments in several boreholes (Cameron et al., 1992), and represents the only absolute age reference within the offshore Pleistocene.

Mapping has distinguished thick pre-glacial from overlying (and incised) glacial formations (Fig. 1.7). The latter are dominated by buried valleys of the Elsterian stage (Swarte Bank Formation).

a)



b)

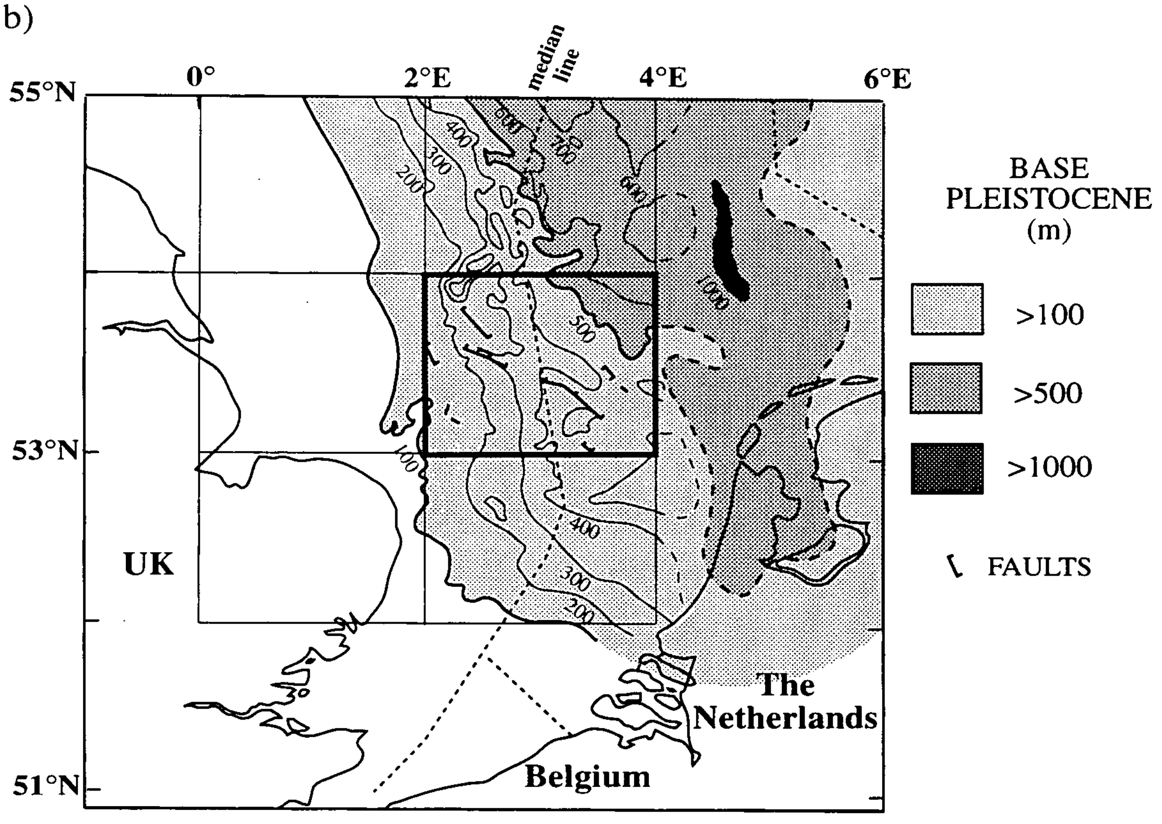


Figure 1.6 - a) southern North Sea bathymetry, and the grid used for 1:250,000 maps by the British and Dutch Geological Surveys (see references under BGS/RGD for authorships); b) depth to the base of the Pleistocene (west of 4°E from BGS/RGD map sheets, east of 4°E modified from Caston, 1977), and faults offsetting the Pleistocene in the study area (BGS/RGD, 1986).

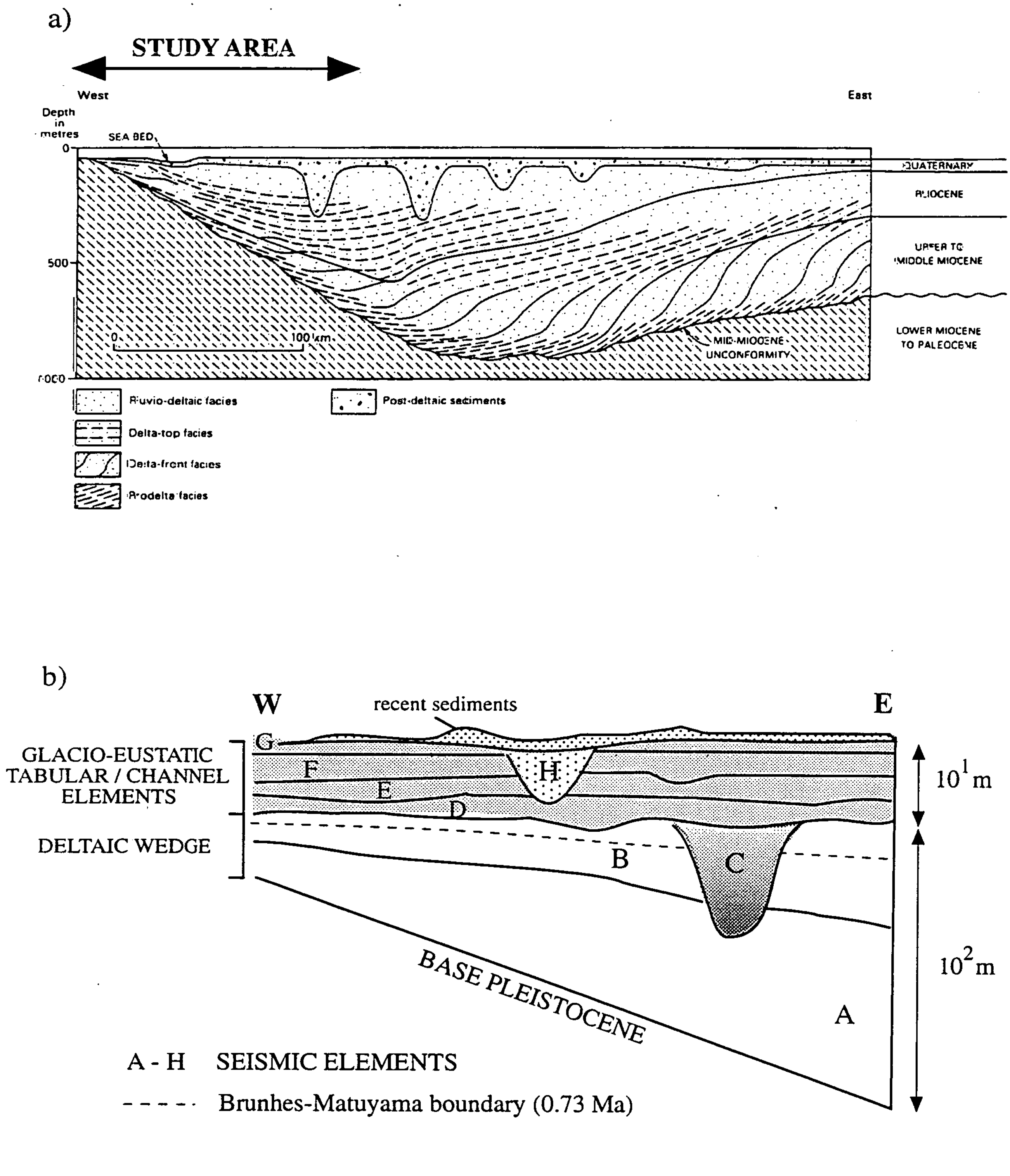


Figure 1.7 - a) schematic cross-section of the southern North Sea, showing the (north)westwards expansion of a Late Cenozoic delta complex (from Cameron et al., 1993); b) schematic Quaternary seismic stratigraphy of the study area (after Cameron et al., 1992).

Seismo-stratigraphic element and lithogenetic divisions		Formation	Depositional environment	Inferred chronostratigraphy
NONDELTAIC DIVISION	U	Various formations	MARINE	HOLOCENE
NONDELTAIC DIVISION	H	SUNDERLAND GROUND (SG)[1]	SUBGLACIAL TO PROGLACIAL : GLACILACUSTRINE TO GLACIMARINE	UPPER WEICHSELIAN
NONDELTAIC DIVISION	H	BOTNEY CUT (BCT)[1]	SUBGLACIAL : GLACILACUSTRINE TO GLACIMARINE	UPPER WEICHSELIAN
NONDELTAIC DIVISION	G	KREFTENHEYE (KR)[2]	PERIGLACIAL : FLUVIAL	UPPER WEICHSELIAN
NONDELTAIC DIVISION	G	TWENTE (TN)[2]	PERIGLACIAL : AEOLIAN	UPPER WEICHSELIAN
NONDELTAIC DIVISION	G	WELL GROUND (WLG)[2]	PROGLACIAL : FLUVIAL	UPPER WEICHSELIAN
NONDELTAIC DIVISION	G	DOGGER BANK (DBK)[3,2]	PROGLACIAL : GLACIMARINE TO GLACILACUSTRINE	UPPER WEICHSELIAN
NONDELTAIC DIVISION	G	BOLDERS BANK (BDK)[3,2]	SUBGLACIAL : TERRESTRIAL	UPPER WEICHSELIAN
NONDELTAIC DIVISION	F	BROWN BANK (BNB)	MARINE TO LACUSTRINE	LOWER WEICHSELIAN
NONDELTAIC DIVISION	F	EEM (EE)	MARINE	EEMIAN
NONDELTAIC DIVISION	E	TEA KETTLE HOLE (TKH)[4]	PERIGLACIAL : AEOLIAN	SAALIAN
NONDELTAIC DIVISION	E	CLEAVER BANK (CLV)[4]	PROGLACIAL : GLACIMARINE	SAALIAN
NONDELTAIC DIVISION	D	EGMOND GROUND (EG)	MARINE	HOLSTEINIAN
NONDELTAIC DIVISION	D	SAND HOLE (SH)	MARINE (LAGOONAL)	HOLSTEINIAN
NONDELTAIC DIVISION	C	SWARTE BANK (SBK)	SUBGLACIAL : GLACILACUSTRINE TO GLACIMARINE	ELSTERIAN
DELTAIC DIVISION	B[5]	YARMOUTH ROADS (YM)	NON-MARINE (FLUVIAL) TO INTERTIDAL	LOWER PLEISTOCENE TO MIDDLE PLEISTOCENE (Cromerian Complex)[6]
DELTAIC DIVISION	A[5]	AURORA (AA)	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	OUTER SILVER PIT (OSP)	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	MARKHAM'S HOLE (MKH)	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	WINTERTON SHOAL (WN)	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	IJMUIDEN GROUND (IJ); CRANE (CX): ? LITTORAL ? TO MARINE ?	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	SMITH'S KNOLL (SK)	MARINE	LOWER PLEISTOCENE
DELTAIC DIVISION	A[5]	WESTKAPELLE GROUND (WK)	MARINE	LOWER PLEISTOCENE
		RED CRAG (RCG)	MARINE	PLIOCENE

Abbreviated formation names are used in subsequent diagrams in this chapter.

1. Partly time equivalent
2. Partly time equivalent
3. Laterally equivalent
4. Partly time equivalent
5. Partly time equivalent and laterally equivalent elements
6. Possibly extends locally into the earliest Elsterian

Figure 1.8 - Transformation of seismic elements into sedimentary formations, and their correlation to the chronostratigraphical framework of northwestern Europe. The Elsterian Swarte Bank Formation corresponds to the tunnel-valleys of this study (see Figs. 1.7, 1.9). From Cameron et al. (1992).

1.3.1 a) pre-glacial and glacial formations

The base of the Pleistocene descends ENE across the area from 100 m to over 650 m bsl (Fig. 1.6b). The Pleistocene base is unconformable on Cretaceous chalks and Paleogene sediments in the west of the study area, but eastward into the basin it is conformable within a thicker upper Miocene to mid-Pleistocene succession resting on a mid-Miocene unconformity (Fig. 1.7a). The succession records the northwestwards expansion of a late Cenozoic delta complex locally up to 1.4 km thick (Zagwijn and Doppert, 1978; Cameron et al., 1993). Seismic facies analysis calibrated by boreholes suggests an upward-coarsening mega-sequence (Fig. 1.7a), from pro-delta through delta-front to delta-top sediments (Cameron et al., 1984, 1987, 1992; Pegler, 1995).

The delta complex in the study area corresponds primarily to Pleistocene delta-front and delta-top facies (Fig. 1.7a). These are represented by two seismic elements (A and B, Fig. 1.7b), the lower of which has been divided into several regional formations (Fig. 1.8). The upper element (B) corresponds to the mid-Pleistocene (<0.7 Ma) fluvio-deltaic sediments of the Yarmouth Roads Formation, which extend across the southern North Sea with thicknesses locally up to 180 m in the Dutch sector (Cameron et al., 1989a). The formation records a vast delta plain across the southern North Sea, subject to transgression and regression so that "a ready-made, low-relief surface was available for the spread of Scandinavian and British ice sheets" (Cameron et al., 1992, p. 108).

The deltaic succession is overlain and incised by seismic elements with tabular and channel geometries (Fig. 1.7b) interpreted to record three mid- to late Pleistocene cycles of glaciation and regression/transgression (Fig. 1.8). They are dominated by the lowermost Swarte Bank Formation, which fills a series of channels up to 400 m in relief correlated to the Elsterian glacial stage (see Fig. 1.3). These are unconformably overlain by tabular units related to successive glacial and marine interglacial stages. A second generation of channels up to 100 m in relief records partly-filled tunnel-valleys formed beneath the late Weichselian ice margin (Botney Cut Formation, Figs. 1.7, 1.8).

1.3.1 b) Swarte Bank Formation

The distribution of Elsterian tunnel-valleys within the Indefatigable sheet was presented as a 1:1,000,000 contour map of estimated trough-to-shoulder relief, drawn from a seismic grid of 5-15 km spacing (Fig. 1.9a). It approximates an isopachyte map of the Swarte Bank Formation in part of the UK sector (Balson and Cameron, 1985, Figure 13), but elsewhere the formation includes an upper part that extends across the valley shoulders (Fig. 1.9b). The contours describe a complex system of anastomosing valleys which trend mainly NNW-SSE, with widths of 0.5-12 km and relief locally over 400 m (Balson and Cameron, 1985; Cameron et al., 1987, 1992; Balson and Jeffery, 1991). A depth below sea level of 510 m near the centre of the area corresponds to an estimated maximum relief of 450 m (Long et al., 1988).

Samples are only available from the upper part of the Swarte Bank Formation in the study area. In the Dutch sector, shallow borings (<100 m) indicate lacustrine muds beneath Holsteinian marine sands (Oele, 1971b; Laban et al., 1984; BGS/RGD, 1986; Laban, 1995). In the UK sector, a single

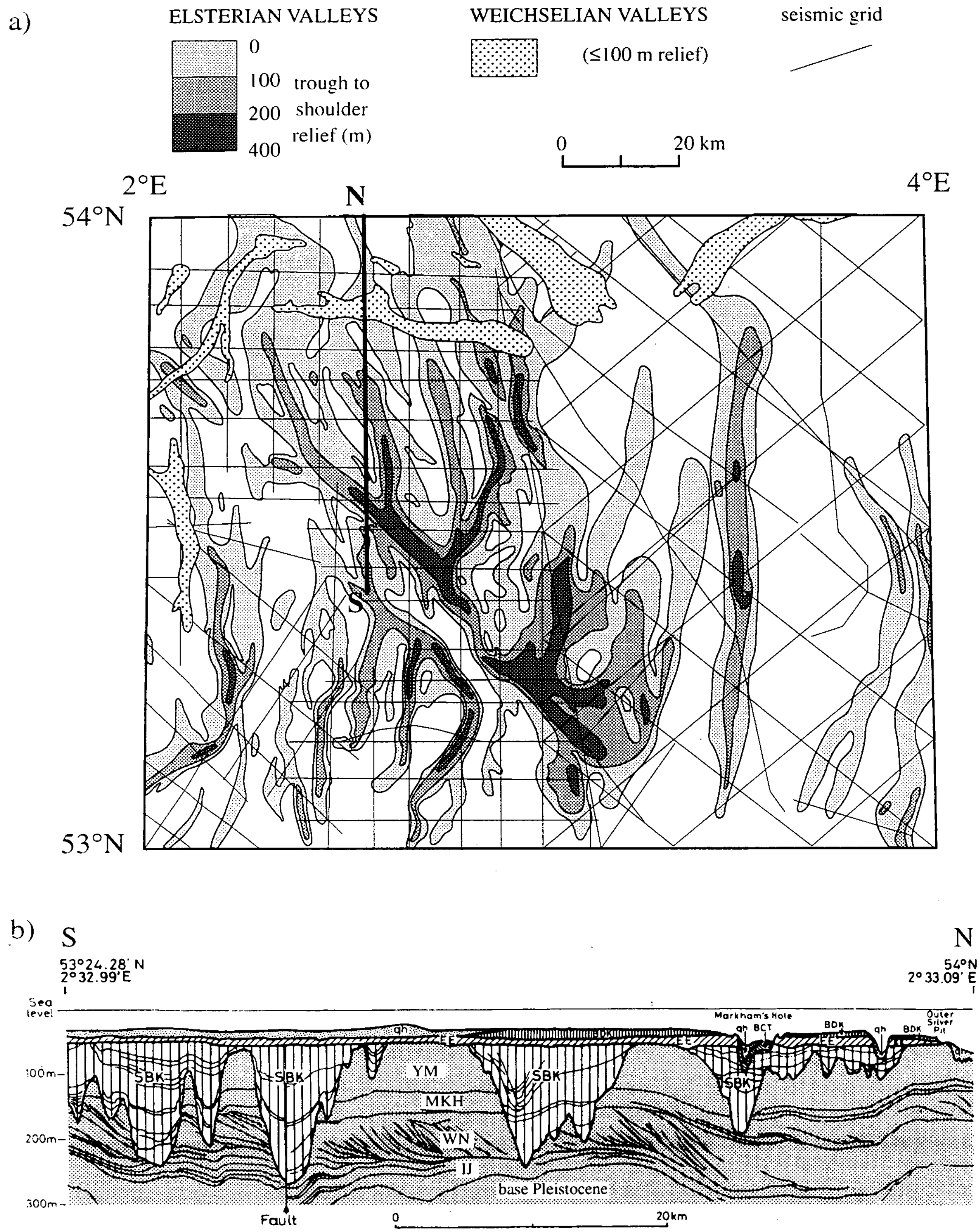


Figure 1.9 - a) contoured relief of Elsterian tunnel-valleys (Swarte Bank Formaton, SBK), and distribution of late Weichselian tunnel-valleys (Botney Cut Formation, BCT) in the study area (from 1:1 million insets on Indefatigable map sheet - BGS/RGD, 1986). b) interpreted airgun seismic reflection profile along the line shown in bold above, showing incision into Pleistocene formations (from Cameron et al., 1989a). For formation abbreviations, see Fig. 1.8; qh=Holocene sediments.

borehole is described as containing up to 30 m of glacigenic mud from the edge of a valley fill, as are two other boreholes from elsewhere in the UK sector (Balson and Jeffery, 1991). The majority of the Swarte Bank formation is thus of unknown composition.

Seismic facies analysis of individual vertical profiles has been used to generalise three fill units: 1 - lower structureless or chaotic facies, 2 - middle layered facies and 3 - upper complexly layered facies. These are suggested to correspond to an upward transition from subglacial or glaciofluvial to lacustrine and marine environments, by comparison with Elsterian tunnel-valley fills beneath northern Germany (BGS/RGD, 1986; Cameron et al., 1987; 1989a; 1992; Long et al., 1988; Balson and Jeffery, 1991; Ehlers and Linke, 1989).

Holsteinian marine sands overlying lacustrine muds in the Dutch sector are mapped as the Egmond Ground Formation (Oele, 1971b; Laban et al., 1984; BGS/RGD, 1984, 1986). Marine sediments sampled in boreholes to the west of the study area have been suggested to be Holsteinian (Hoxnian) on the basis of pollen content (Fisher et al., 1969) and foraminiferal content (Balson and Cameron, 1985) and have been attributed to the Egmond Ground Formation (BGS, 1991; Balson and Jeffery, 1991; Cameron et al., 1992). The putative marine seismic facies at the top of the Swarte Bank Formation in the UK sector of the study area has also been suggested to belong to the Egmond Ground Formation and to be extensive across valley shoulders, but has not been sampled (Cameron et al., 1988b, 1992).

The Swarte Bank Formation has been mapped in the offshore to the west and north of the study area (Fig. 1.10) as the fill of generally smaller (<200 m relief) buried valleys, although an isolated feature over 300 m relief occurs to the north in the Dutch sector. East of the study area, the distribution of 'Elsterian valleys' was presented by Laban (1995, Figure 23), without reference to their depths (Fig. 1.10). Other work in the area suggests relief of over 250 m (Cameron et al., 1993; Pegler, 1995, Figure 2.21). The Swarte Bank Formation is correlative to the Peelo Formation of the northern Netherlands (Fig. 1.10), which comprises glaciolacustrine muds extensive between underlying valleys (up to 350 m bsl) filled with sands (ter Wee, 1983). Smaller buried Elsterian (Anglia) tunnel-valleys beneath East Anglia (Fig. 1.10) are incised up to 100 m into Cretaceous chalk and filled with sands and muds (Woodland, 1970).

The Swarte Bank Formation appears to terminate to the south along 53°N and this has been suggested to record a former ice margin line (Cameron et al., 1992). However, the termination in part reflects the sequence of offshore mapping (Fig. 1.6a), as the formation was not identified during prior mapping of the Flemish Bight sheet to the south (BGS/RGD, 1984). Small-scale channelised relief (<50 m) occurs at the top of the Yarmouth Roads Formation on the Flemish Bight Sheet to as far south as 52°30' N and has been retrospectively suggested to include outliers of the Swarte Bank Formation in the UK sector (Cameron et al., 1992, Figure 105). Boreholes in the Dutch sector of the Flemish Bight sheet confirm that the fill of equivalent channels includes Holsteinian sediments (BGS/RGD, 1984). My results also show that some tunnel-valleys extend south of 53°N (Chapter 2), although none so large as in the study area.

1.3.2 Elsterian Palaeogeography

The study area lies near the southern limits of the former Elsterian ice sheet. Two aspects of the geographical setting of the period are of interest: a) the maximum southern extent of the ice sheet and b) the probable existence of a proglacial lake dammed between the ice-margin and the Dover Strait.

1.3.2 a) southern extent

The southern limit of Elsterian glaciation in the Southern Bight of the North Sea is generally drawn by interpolation between onshore limits based on depositional evidence (Fig. 1.10). Subsurface evidence onshore and offshore suggest that these limits are conservative.

The regional limit of till of the Anglian glaciation is shown in Fig. 1.10 along with the subsurface distribution of valleys studied by Woodland (1970) and mapped by BGS (1991b). Several studies indicate the occurrence of buried channels and valleys of apparent subglacial origin beyond the limit of till. Mathers et al. (1991) presented geophysical and borehole evidence for a system of small buried channels (up to 200 m wide, 8 m deep) on the east coast, just beyond the limit of till. The channels were filled with glaciofluvial sands and gravels interbedded with diamictons and were attributed to a sub-marginal origin. Lake et al. (1977) presented borehole evidence from the northern coast of the Thames estuary, well south of the till limit, for relatively shallow (≤20 m) buried valleys up to 1 km wide, oriented oblique to the coast (shown schematically in Fig. 1.10). The valleys contained lower sands and upper muds, the latter containing a lacustrine to marine transition, and were argued to record meltwater erosion and deposition beneath the margin of the Anglian ice sheet. D'Olier (1975) presented contours on the bedrock surface in the Thames Estuary based on seismic data and suggested that two elongate depressions of irregular axial relief were tunnel-valleys. These studies thus suggest Elsterian ice extent at least as far south as 51°50'N offshore.

The till limit extends off the coast of East Anglia but till is not recognised offshore, presumably due to later erosion (Long et al., 1988). However, buried outliers of the Swarte Bank Formation may extend south to at least 52°30'N as noted (Fig. 1.10; Cameron et al., 1992). In addition, buried glaciotectonic disturbances are recognised as far south as 53°15'N (Fig. 1.10) and have been attributed to the Elsterian glaciation by reference to overlying Holsteinian sediments in boreholes (BGS/RGD, 1984; Laban, 1995).

The Elsterian maximum line in the northern Netherlands is drawn along the southern limit of the glaciolacustrine muds of the upper Peelo Formation (Fig. 1.10), which extend between and beyond buried tunnel-valleys (ter Wee, 1983). Geomorphological evidence for the extent of the Elsterian glaciation is unavailable as the Saalian glaciation was more extensive in this area (Fig. 1.10). The Elsterian glaciation may have extended as far south as the Saalian ice limit, although no stratigraphical evidence for this has been recognised.

There is thus evidence to suggest that the Elsterian ice margin extended to at least 52°N in the southern North Sea, i.e. over 100 km south of the southern margin of the study area (Fig. 1.10).

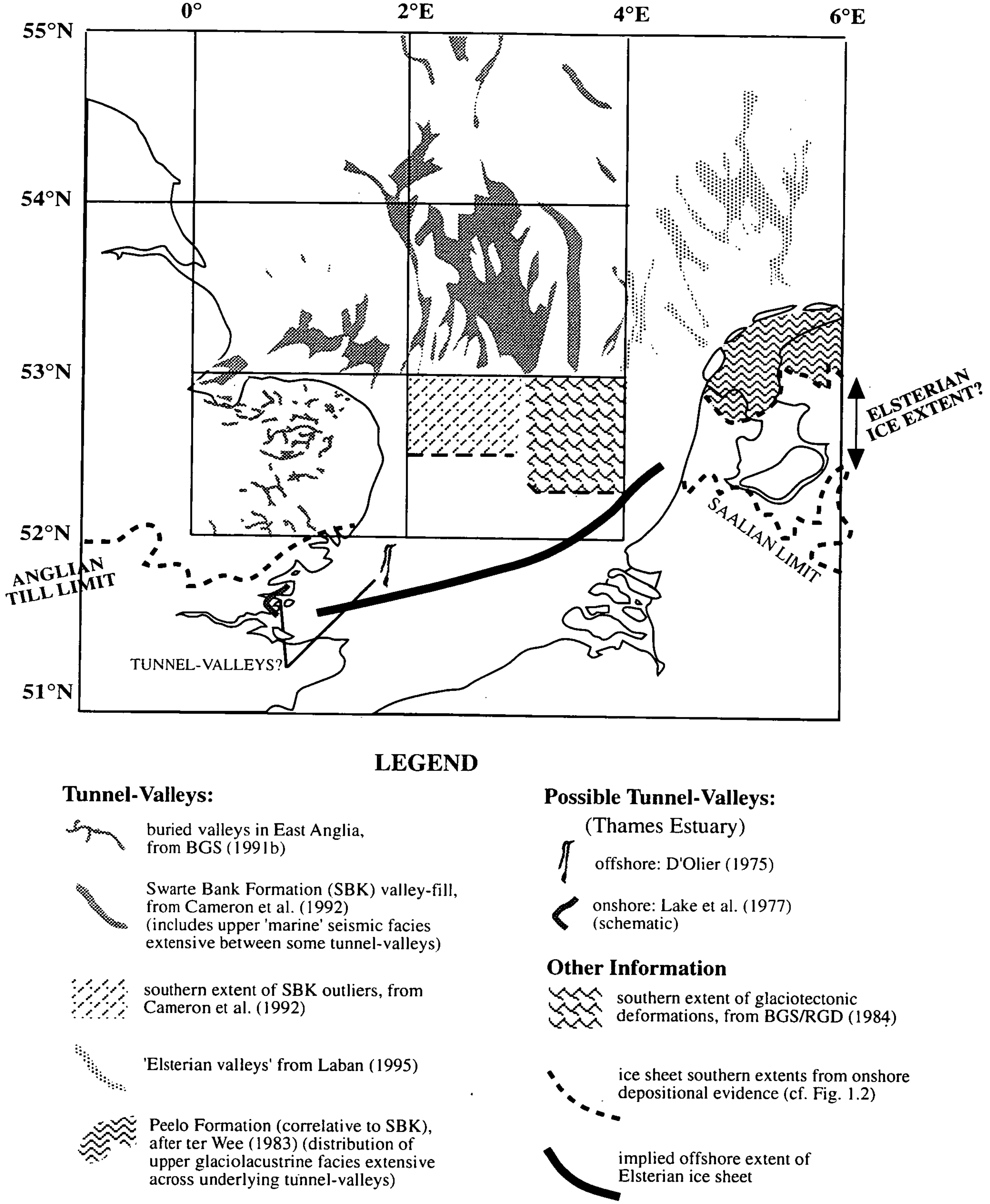


Figure 1.10 - Summary of different forms of evidence for the presence of the Anglian/Elsterian ice sheet in the southern North Sea and a line (bold) delimiting their southern extent. Buried tunnel-valleys are recognised from East Anglia to the northern Netherlands. Their offshore extent has been mapped (within the 1:250,000 grid) as the Swarte Bank Formation, outliers of which occur south of 53°N (Cameron et al., 1992) along with ice-push features of inferred Elsterian age (BGS/RGD, 1984). Possible tunnel-valleys occur beyond the till limit along the northern coast of the Thames estuary and offshore. The extent of the Peelo Formation in the northern Netherlands bounds the minimum extent of the Elsterian ice sheet beneath the younger landscape of the Saalian glaciation.

1.3.3 b) proglacial lake

Stamp (1927) first suggested the likelihood of a large lake dammed by the Elsterian ice margin, overflow from which would have initiated the Dover Straits. Roep et al. (1975) presented evidence for the overflow of such a lake, in the form of deposits along the southern shores of the Pas de Calais which record a former fluvial system that flowed south from the present offshore and west into the English Channel. The deposits consist of distally (northern) derived sands and locally derived gravels, up to 33 m above present datum, deposited during two episodes of overflow attributed to one or possibly two glacial stages prior to the last.

Gibbard (1988) observed that the overflow channel described by Roep et al. may have been one of several, another of which became dominant along the present Dover Straits. He suggested that the lake would have been a stable feature of the Elsterian glaciation, whereby runoff from the ice sheet and from western European rivers would have been directed seaward along a former Channel river. He reviewed evidence for lakes dammed along the receding ice margin in East Anglia, as well as the more extensive evidence of the Peelo Formation in the northern Netherlands (Fig. 1.10). He suggested that depositional evidence exists for a lake across the intervening southern North Sea, in the form of the Swarte Bank Formation. However, the Swarte Bank Formation is not regionally extensive (Fig. 1.10), and seismic facies interpreted as lacustrine deposits are largely confined to the upper parts of tunnel-valleys, except in parts of the Dutch sector (Laban, 1995).

It is probable that during the Elsterian occupation of the southern North Sea the immediate proglacial environment would have consisted of a large lake, perhaps tens of metres in depth, with an outlet through the present Dover Strait. In any case, the elongate tunnel-valley depressions themselves would have constituted proglacial lake basins during the withdrawl of the Elsterian ice margin.

CHAPTER 2
SEISMIC REFLECTION MORPHOLOGY AND STRATIGRAPHY

2.1 INTRODUCTION

The purposes of this chapter are:

- to present remotely-sensed evidence of the form of the valley base unconformity and the three-dimensional stratigraphical organisation of surfaces within the fill
- to interpret the results in terms of basal erosion to the south and fill progradation to the north within successive cut and fill elements overlapping to the north

The study area tunnel-valleys have previously been examined using a broad grid (5-15 km) of shallow seismic reflection profiles totalling about 4000 line-km (BGS/RGD, 1986). Contours on relief suggested a complex anastomosing system of elongate, subparallel basins up to 450 m deep and 12 km wide (Fig. 1.9). Seismic facies analysis indicated three fill units: lower structureless, middle layered and upper complex (Cameron et al., 1987; Balson and Jeffery, 1991).

I re-examine the tunnel-valleys using over 40,000 line-km of digitally processed exploration seismic data, supplemented by about 2000 line-km of analog BGS profiles in the UK sector (Table 2.1). Grids of ≤1 km spacing cover large parts of the 100 x 150 km area and a 3D-seismic volume provides continuous information (25 m grid) over a 22 x 39 km area (Fig. 2.1). Depth contours on the valley base yield a new picture of its morphology. Contoured fill reflecting surfaces along eight larger valleys allow a seismic sequence analysis based on discordant reflector relations.

The *valley base* defines elongate basins up to 6 km wide and locally up to 500 m in relief. In plan, the basins converge to the south as a series of arborescent elements of rectilinear morphology, a pattern complicated locally by anastomosing. Stratal and structural trends in the incised Cenozoic sedimentary strata have only locally influenced cross-sections (benches following stratal contrasts) and plan form (changes across faults). However, valley size and spacing increase eastwards across the area in parallel with substrate thickness.

The *valley fill* is divisible into three seismic sequences bounded by basinal relief: **I** - axially downlapping clinoform surfaces up to 3.5° in gradient and 3-20 km in extent, which dominate the fill with thicknesses up to 400 m and record northward progradation; the discordant upper boundary defines axial basins up to 120 m in depositional relief; **II** - onlapping surfaces up to 150 m thick that fill the axial basins and extend throughout the upper parts of adjacent tunnel-valleys; **III** - complex reflectors up to 100 m thick that fill axial basins up to 70 m in erosional relief and extend thinly (<25 m) across some valley shoulders in the northern UK sector. The sequences correspond to the seismic facies of BGS/RGD analysis, however the lower fill is not structureless.

Table 2.1 - Seismic reflection data types used.

DATA TYPE	BAND-WIDTH Hz	SEPARABLE RESOLUTION *m* *	S/N RATIOS	LINE SPACING	LINE-KMS
BGS PROFILES	100-1000 (analog)	1-2	high above 60-110 ms	5-15 km	2200
RGD PROFILE	32-640 (digital)	2-3	high below 100 ms	(1 line)	160
EXPLORATION PROFILES	5-100 (digital)	5-15	high below 100-200 ms	≤1 km	12,000
EXPLORATION 3D-VOLUME	8-65 (digital)	5-10	high below 150 ms	25 m	34,000

* half wavelength at 2 km/s TWT (maximal), measured from profiles; see also Fig. 2.2.

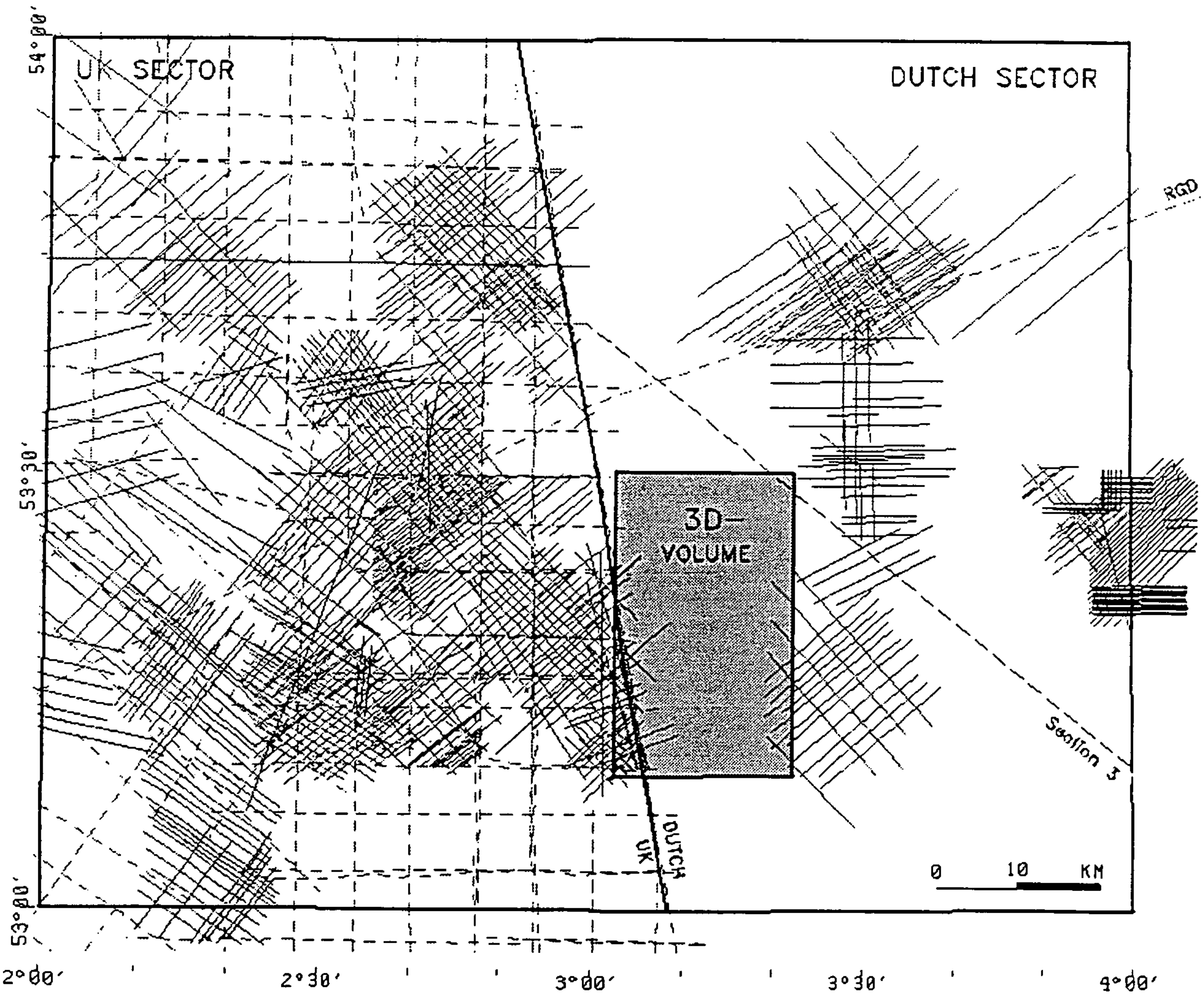


Figure 2.1 - Distribution of reflection data types: BGS/RGD profiles (dashed lines), exploration profiles (solid lines) and 3D-seismic volume. RGD and Section 3 refer to published profiles (respectively Cameron et al., 1993 and BGS/RGD, 1986). See also Enclosure.

Erosional overlap (a form of cross-cutting) by younger cut and fill elements to the north is recorded by surfaces of discontinuity, which bound northwards reduction in clinoform gradient and coincide with plan form complexity (apparent anastomosing, angular offsets). The valley base thus records erosion convergent to the south, but younger to the north, while the fill records progradation to the north. This morpho-stratigraphical architecture is explicable by drainage south to an ice margin that receded north. A similar erosional morphology has been proposed for other tunnel valley networks but evidence for fill progradation has not previously been presented.

I begin by discussing reflection data and its interpretation (section 2.2). I then present evidence of basal morphology (section 2.3), fill stratigraphy (section 2.4) and erosional overlap (section 2.5). Finally, I interpret the results in terms of drainage to receding ice margins (section 2.6).

2.2 REFLECTION METHODOLOGY

Seismic reflection data provide remote information on physical properties, subject to fundamental limitations of resolution and noise content. Exploration data are of low vertical resolution relative to BGS profiles but of superior signal to noise ratio (Table 2.1). In addition, their close line-spacing affords enhanced spatial resolution, optimised in a 3D-seismic volume (Fig. 2.1).

Seismic stratigraphical analysis is a means of establishing material units (sequences) based on reflector relations, interpreted by analogy with stratal surfaces. Geomorphological analogy also plays a role in seismic interpretation, notably in the glacial basins of this study.

I discuss the basis of reflection profiling (2.2.1), the data types available (2.2.2), seismic interpretation (2.2.3) and the mapping of the study area (2.2.4).

2.2.1 Geophysical Basis

The acquisition and processing of seismic reflection data have been discussed in detail by Yilmaz (1987) and more generally by Badley (1985) and McQuillin et al. (1984).

The reflection method is based on the transmission of a pulse of sound of given frequency content (bandwidth) from a source to an offset receiver via incident reflecting interfaces. The recorded seismic trace (x) represents a convolution (time-series superimposition) of the outgoing wavelet (w) with the earth's impulse (impedance) response (e), plus noise (n), that is:

$$x(t) = w(t) * e(t) + n(t)$$

The goal of reflection profiling is to evaluate recorded traces in terms of the earth's response e. It is therefore necessary to address both a) convolution (resolution) and b) noise (digital processing).

2.2.1 a) convolution (resolution)

Assuming the earth's impulse response to consist of discrete interfaces, their resolution is limited by the source wavelet dimensions. Resolution can be expressed as fractional source wavelengths (λ), both vertically and horizontally (Sheriff, 1984). Vertically, separable resolution describes two wavelets at limits of separation (1/2 λ). Closer spacing results in constructive interference, to an amplitude maximum near a resolvable limit (1/4 λ). Horizontal resolution is described in terms of the radius of the first Fresnel zone, the area circumscribed during passage of a quarter wavelength of the spherical wavefront within which all reflected energy interferes constructively.

Resolution is thus a measure of the closest vertical or horizontal spacing of interfaces that can contribute to reflection without interference. Vertical interfaces of closer spacing than half the source wavelet (<1/2 λ) will result in a recorded reflection that represents a complex convolution, or an interference pattern. Such complex reflections may appear to be discrete, as illustrated by Badley (1985, Figures 2.12-13). Similarly, horizontal interference may result in the apparent lateral continuity of surfaces which contain gaps, as illustrated by Sheriff (1984, Figure 13).

Resolution in the shallow sedimentary materials of interest is a function of possible source frequencies in the range of 1-1000 Hz and velocities in the range of 1.5 km/s (water) to 2 km/s. Assuming a linear velocity profile within the upper second, vertical resolution is seen to vary mainly with frequency while horizontal resolution decreases more rapidly with depth (Fig. 2.2). This neglects a natural decrease in resolution with depth due to earth filtering of higher frequencies. The actual resolution of broadband source wavelets depends in a complex manner on their pulse lengths and bandwidths, such that comparison of their frequency content with Fig. 2.2 yields only a maximum range of resolution. The alternative is to measure resolution directly from profiles, as wavelet dimensions which increase with time (Badley, 1985, Figure 2.15). Both approaches yield shallow separable resolution in the range of 1-15 m, depending on source characteristics (Table 2.1).

2.2.1 b) noise (digital processing)

Noise is described as random or coherent. Coherent noise results from secondary reflections (multiples) within a succession of interfaces. The seabed multiple is a particular problem of marine reflection profiling and results in decreased signal to noise (S/N) ratios below twice the water depth. Noise reduction may be achieved through the acquisition and digital processing of multiple fold data, which are amenable to three primary processing methods: stacking, deconvolution and migration (e.g. Yilmaz, 1987). The first two methods have both benefits and disadvantages for the upper second.

Stacking refers to the combination of multiple traces, collected at increasing offset from the same common mid-point (CMP) to generate a single trace of multiple 'fold'. This results in the constructive reinforcement of primary reflections and the destructive interference of both coherent and random noise due to differential normal move-out (NMO). NMO is the non-linear upward shift in the time-series necessary to bring each trace to zero-offset, from a 'normal' hyperbolic TWT versus offset relationship that is a function of interval velocity. In the shallow section, NMO correction results in a shift to lower frequencies that increases both with proximity to source and with offset, while farther offset traces may be distorted due to low angles of incidence. The solution to these problems is to remove, or mute, the shallow portions of traces at depths increasing with offset. The result of a muting ramp is that data fold decreases upwards to one (or less) at source.

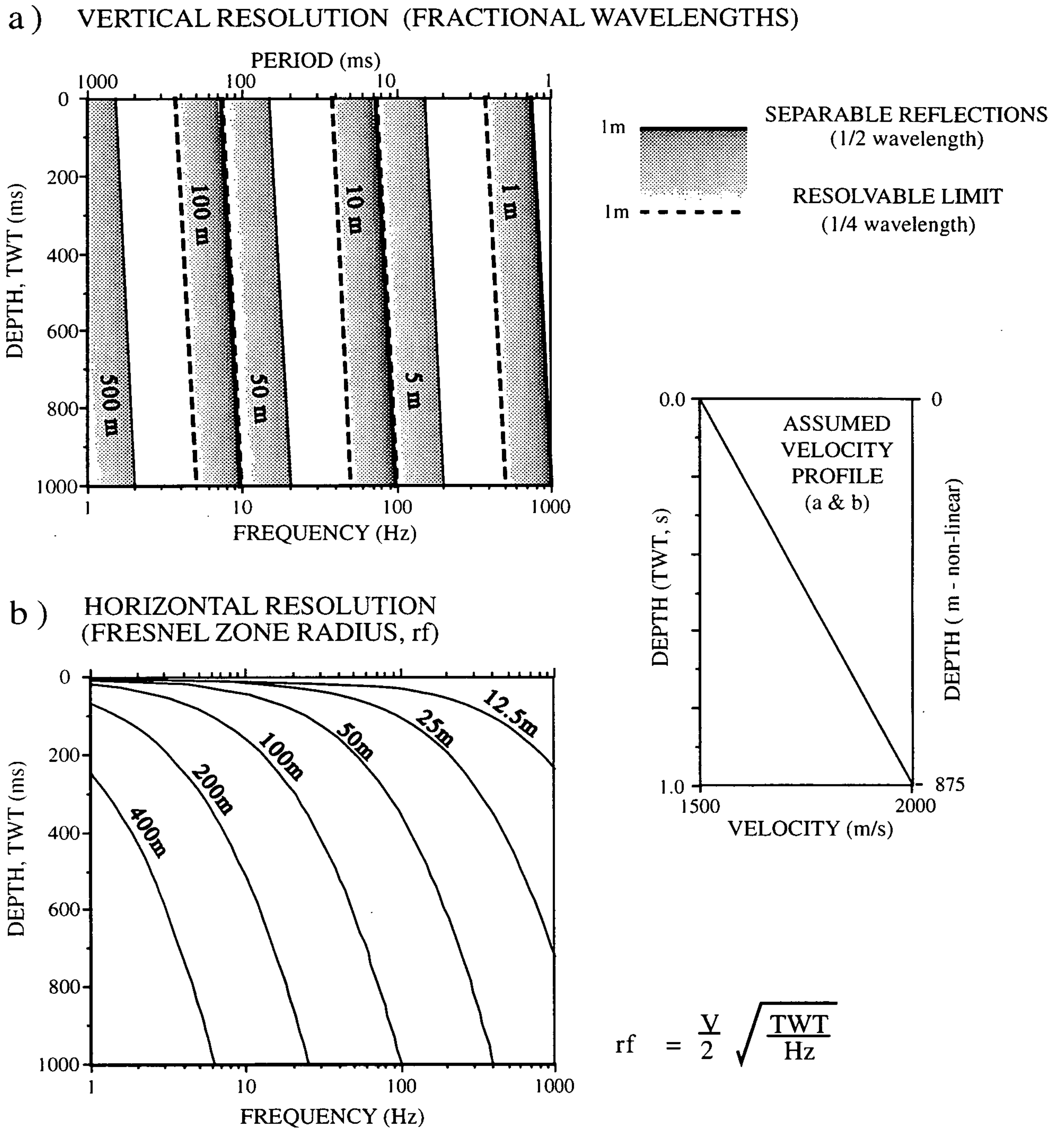


Figure 2.2 - Theoretical measures of seismic resolution for frequencies and velocities representative of the shallow section (upper second) of interest.
a) vertical resolution: waveforms interfere constructively at spacings closer than 1/2 wavelength, to a maximum at about 1/4 wavelength.
b) horizontal resolution: Fresnel zone (1/4 wavelength) dimensions distinguish apparently continuous reflectors (>rf) from point source (<rf) reflectors, for unmigrated data.

Deconvolution acts to compress the source wavelet and to remove multiple energy, thus enhancing both resolution and S/N ratios. Deconvolution may be applied before and after stack and uses algorithms that seek to statistically determine the source wavelet characteristics from the recorded traces. The algorithms tend to define the operator 'window' using the first seabed multiple, resulting in its enhancement.

Migration is a form of spatial deconvolution that acts to improve horizontal resolution, by moving dipping events to their 'correct' location in the plane of section and focusing energy spread out over Fresnel zones (diffraction collapse to points). Migration has minimal effect on the shallow section as displacements decrease with proximity to source. In any case, shallow velocity control is poor due to the low data-fold constraining shallow NMO velocities. Migration of multiple fold data across the upper second of the study area resulted in no perceptible changes (J. Bulat, BGS, personal communication, 1993; *cf.* Cameron et al., 1993).

2.2.2 Available Data

The three available reflection data types vary in resolution (frequency content), S/N ratio and line-spacing (Table 2.1). Examples are provided of a) BGS/RGD profiles and b) exploration profiles (Figs. 2.3, 2.4) and of c) 3D-seismic volumetric data (Fig. 2.5). Water depths across the study area mainly in the range of 25-40 m (33-53 ms) imply multiple effects below 65-110 ms.

2.2.2 a) BGS/RGD profiles

BGS profiles in the study area were collected from 1979-82 along about 2200 km of lines (Table 2.1; Fig. 2.1). Shallow survey methods used by BGS have been described by McQuillan et al. (1984) and Fannin (1989). Profile lines were run using broadband, compressed wavelet sources, recorded on hydrophones up to 30 m long, filtered to minimise random noise and printed as paper records taken to approximate zero-offset sections. Sources were fired periodically (1-2 seconds) rather than by distance so that horizontal shot separations depend on survey speed. Navigation was mainly by radio positioning with an estimated accuracy within hundreds of metres.

I obtained photocopies of the original paper records held in BGS archives, inclusive of field information. Most lines were run using several sources concurrently: depths up to one second were examined using 5-10 in^3 airguns or 4-5 kJ sparkers (100-500 Hz), while information on the upper 500 ms was obtained using higher frequency 0.5-1 kJ sparker sources (200-1000 Hz). The frequency bandwidths imply vertical separable resolution of better than 5 m and wavelets measured on profiles (2-4 ms) indicate separable resolution of 1-2 m. However, this nominally high resolution is impaired by low S/N ratios due to multiple noise below 65-110 ms (e.g. Fig. 2.3C). Interpretation below the first seabed multiple depended on data quality (notably survey speed) and subsurface characteristics.

A single digitally processed shallow profile across the study area was collected by RGD in 1987 (Fig. 2.1, Table 2.1) and published by Cameron et al. (1993). I obtained paper copies of the processed records courtesy of J. Bulat (BGS). The profile was designed to investigate the upper second of section, using a high frequency broadband source (32-640 Hz) deconvolved and stacked to obtain data fold of up to 12 at depth. Wavelets measured on profiles (4-6 ms) indicate vertical separable resolution of 2-3 m. Signal to noise ratios are high due to multiple suppression, save in the upper 100 ms (Fig. 2.4A).

4.2.2 b) exploration profiles

Exploration profiles were acquired and digitally processed as multiple fold records using standard methods (e.g. Yilmaz, 1987). 'Tuned' airgun source arrays were fired by distance (shot separations of 12.5-25 m) and recorded on hydrophones up to 3 km long. Navigation was by radio and satellite positioning with an accuracy within metres. I obtained copies of the upper second of about 500 profiles collected mainly between 1980-90 along about 12,000 km of lines (Table 2.1, Fig. 2.1). Records were supplied by exploration companies as black and white plastic and/or paper profiles, most as migrated sections. Data fold decreased toward source in a manner dependent on muting ramps. I estimate fold of less than 10 in the upper second, corresponding to offsets of less than 1000 m. Digital sampling of 2-4 ms resulted in filtered frequency content of less than 100 Hz and so separable resolution no better than 5 m. Deconvolved wavelets measured directly from profiles vary from 10-30 ms corresponding to separable resolution of 5-15 m.

Vertical resolution is thus one tenth that of BGS profiles (Table 2.1). In addition, exploration target depths of generally over 1 km meant that little attention was given to optimising the shallow section. Nonetheless, S/N ratios are low and data quality high below 100-200 ms (Figs. 2.3, 2.4). The uppermost data were generally muted due to a combination of i) deconvolution enhancement of the first seabed multiple (65-110 ms), ii) limited multiple suppression due to low data fold (<2) and iii) low frequency stretching. The upper limit of data depended on acquisition parameters, notably the separation between the source and near-offset receiver which varied from 75-200 m. Shorter separations (<100 m) and careful processing yielded information to seabed in some cases (Fig. 2.3 A,B) but in general the seabed reflector (33-53 ms) was not represented (Fig. 2.4 B-E).

Profile data included paper extracts from two 3D-seismic volumes in Dutch block K12 on the eastern margin of the study area. Grid spacing (0.5-1 km) was similar to that of adjacent exploration profiles (Fig. 2.1), with comparable spatial benefits for interpretation.

2.2.2 c) exploration 3D-volume

Exploration 3D-seismic methods and analysis were discussed in detail by Brown (1988). Exploration profiles are collected at line spacings which approach in-line shot separations (10-100 m). CMPs are reattributed to an orthogonal grid of 'bins' and deconvolution and stacking within bins is accompanied by 3D-migration. The resulting matrix of elements represents a continuous volumetric sample of the earth. It is 'explored' by examining planar samples, or slices, along the three orthogonal axes (in-lines, cross-lines and horizontal or time-slices) or along arbitrary planes.

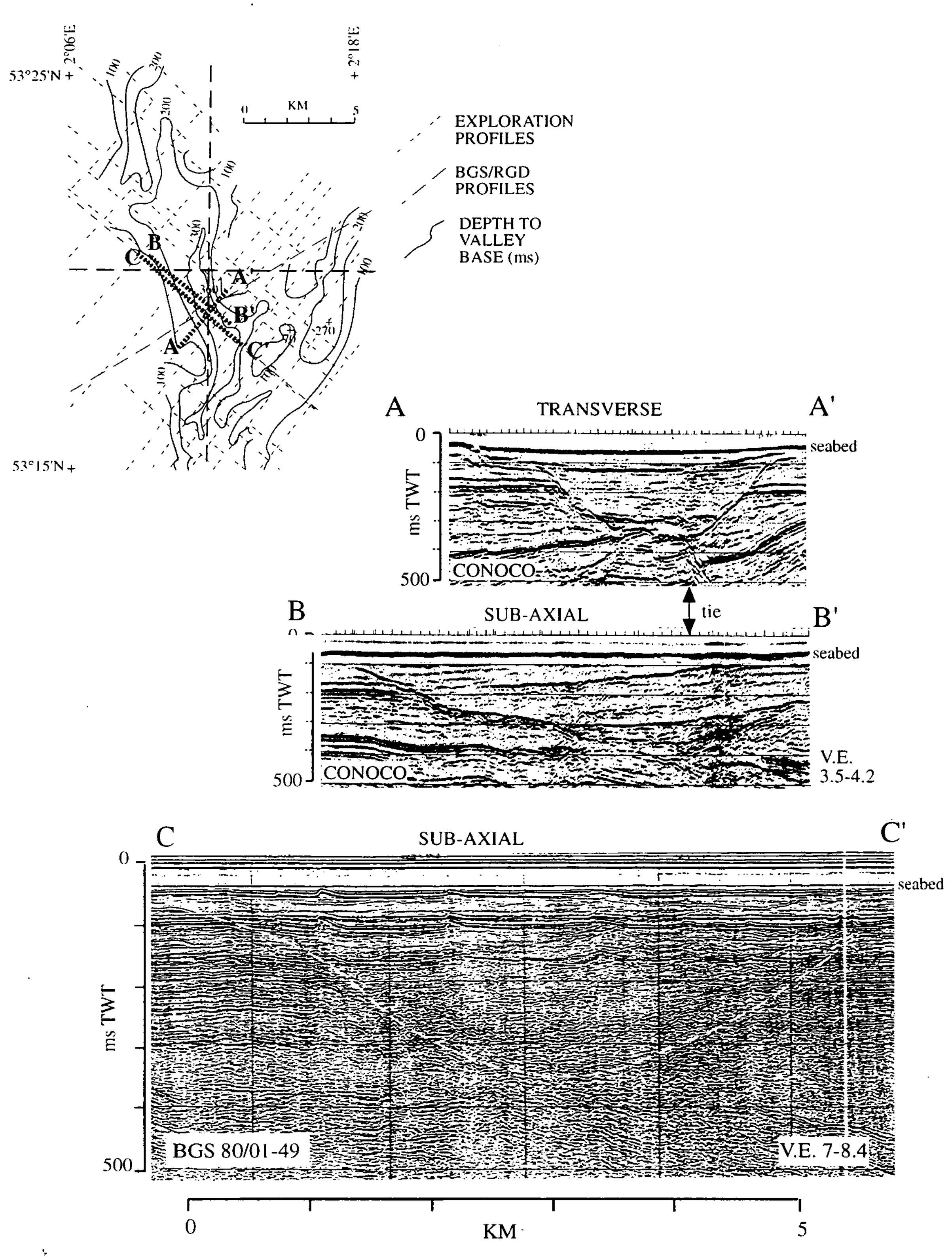


Figure 2.3 - Examples of exploration profiles (A,B) and a BGS profile (C) across a tunnel-valley in the southwestern UK sector (see Fig. 2.7). The exploration profiles are of exceptional shallow quality: the seabed is resolved, although the upper 100 ms are noisy (cf. Fig. 2.4). In contrast, the BGS profile is dominated by multiple noise below about 100 ms. Profiles B and C are subparallel and nominally 400 m apart, within the navigational accuracy of BGS data. Note downlapping clinoform reflectors on profiles A and B (fill sequence I).

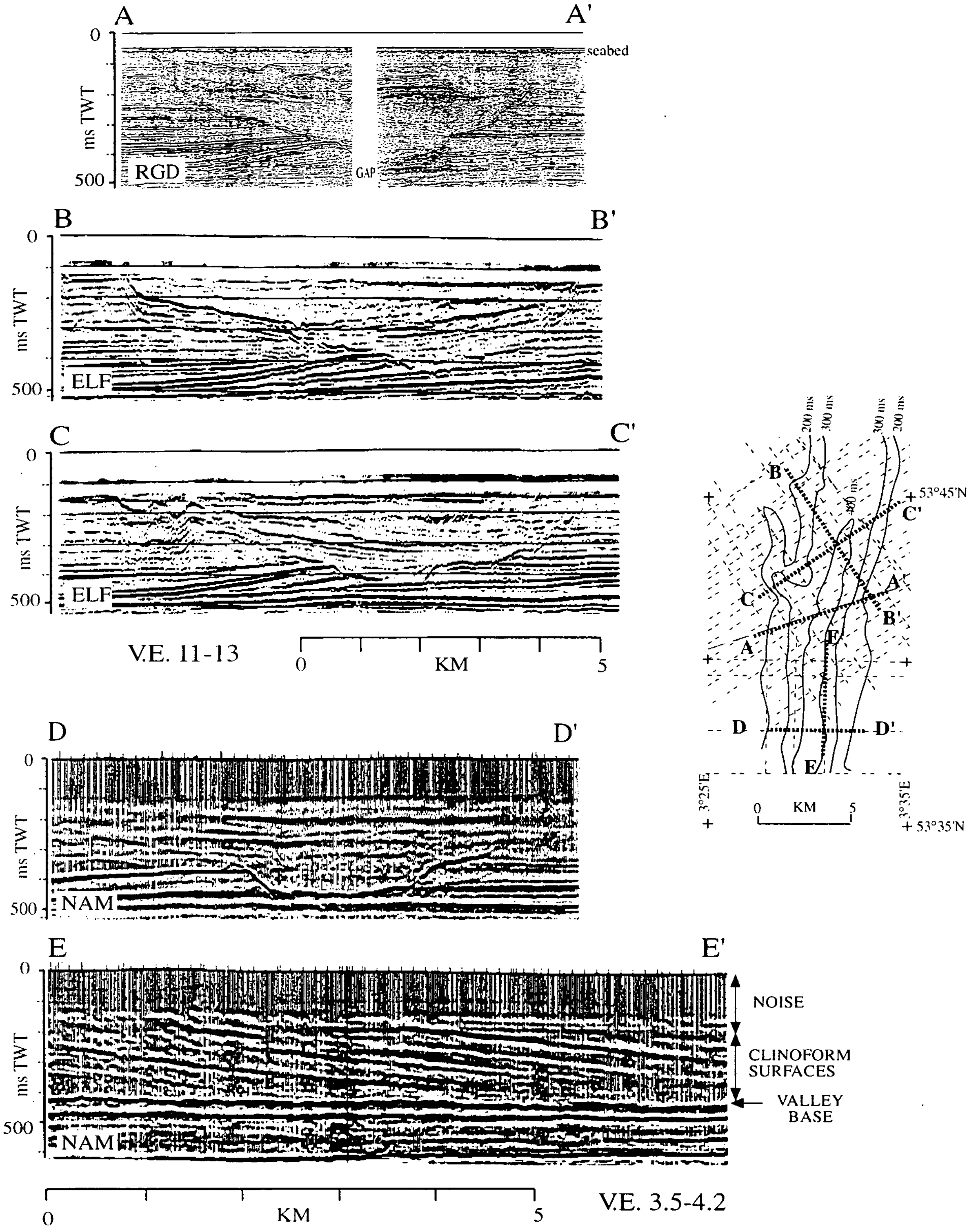


Figure 2.4 - Examples of processed reflection profiles across and along a tunnel-valley in the north-central Dutch sector (see Fig. 2.7). An RGD profile (A) is of high vertical resolution and signal to noise content (see Table 2.1). Exploration profiles (B to E) are of much lower vertical resolution, and the upper 150 ms are lost to noise. However at greater depths, oblique, transverse and axial profiles provide information on the valley base, and on axially downlapping clinoform surfaces within the fill (sequence I). The area corresponds to about 55-75 km along axial profile F (Fig. 2.15). Note different horizontal scales of A-C versus D-E.

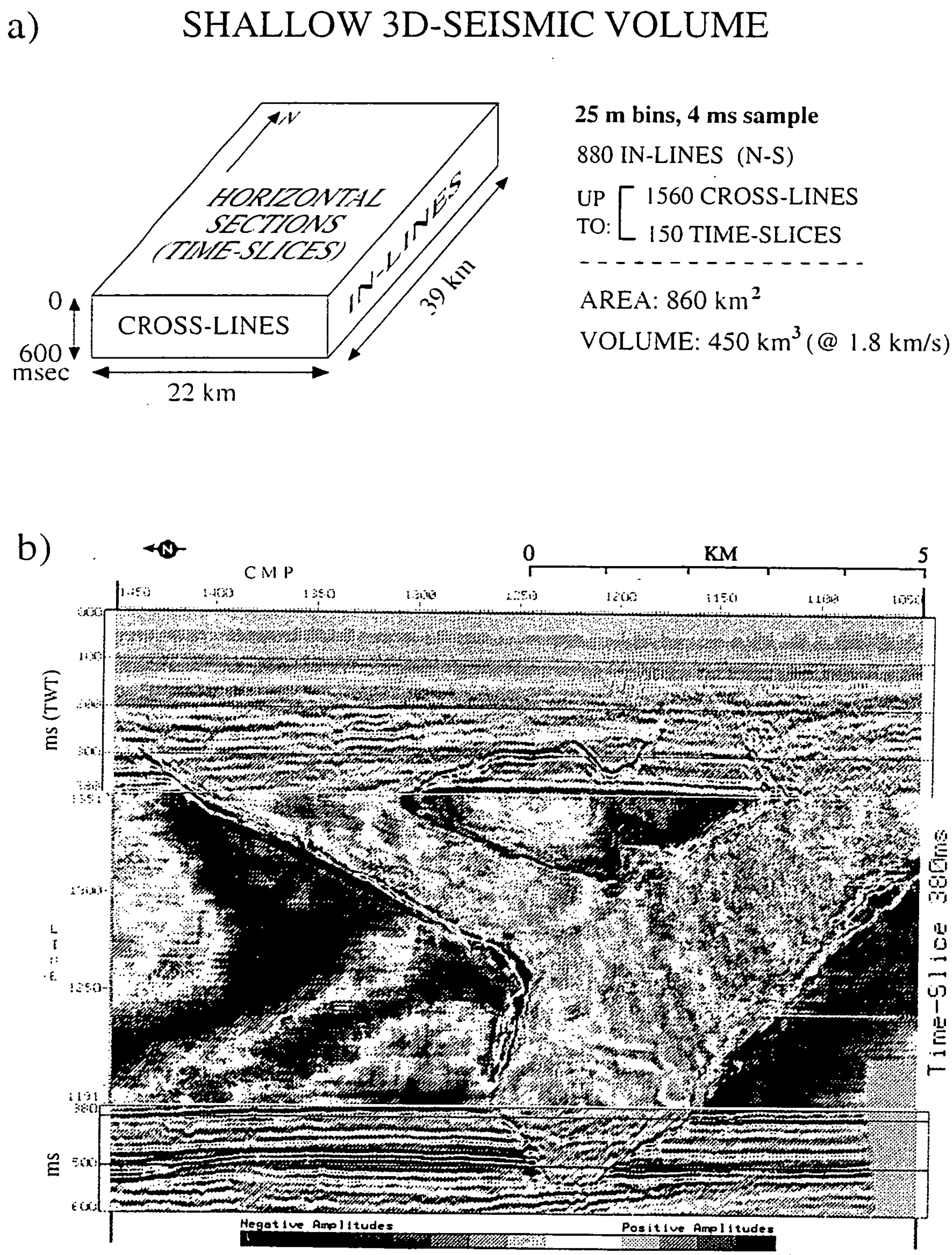


Figure 2.5 - a) exploration 3D-seismic volume characteristics; b) example of data in vertical+horizontal display: time-slice at 380 ms (3.8 x 10 km), bounded by vertical in-lines ('view' to east). The upper 150 ms are lost to noise; seabed is at about 50 ms. Note truncation of incised reflectors in both vertical and horizontal sections, and southward convergence of tunnel-valleys. Location on Fig. 2.16. Dual-polarity colour original. Courtesy of Wintershall Noordzee.

I obtained access to the upper 600 ms of a 3D-seismic volume collected in 1989-90 from Dutch blocks K10 and K13 (Figs. 2.1, 2.5). The volume covers a 22 x 39 km (860 km^2) area and comprises 25 m bins at a 4 ms vertical sample rate. Bandwidth and vertical resolution are within the range for exploration profiles, discussed above (Table 2.1). Data fold decreases towards source from a maximum of five at 600 ms and the upper 100-150 ms are lost to noise (Fig. 2.5b).

3D-seismic data were obtained on digital tape and explored using *Geoquest* interpretation software (version 8) on a *Sun* workstation. The program allowed a variety of display options including flexible scales, combined vertical and horizontal images and dual-polarity colour schemes. From 880 north-south oriented in-lines (25 m spacing), I generated 156 west-east cross-lines (250 m spacing) and up to 30 time-slices (8-20 ms intervals). In addition, vertical 'reconstruction' profiles (CMP gathers) were generated along defined lines (e.g. valley axes).

2.2.3 Seismic Interpretation

Seismic reflectors are interpreted by implicit analogy with stratal surfaces, to limits of resolution. Seismic sequences defined from reflector relations are material stratigraphical units, independent of abstract concepts such as time or genesis (Owen, 1987). Geomorphological analogy plays an important role in the interpretation of both profile and volumetric seismic data, particularly in the glacial basins of this study.

I discuss a) the nature of reflectors, b) sequence analysis and c) geomorphological analogy.

2.3.1 a) nature of reflectors

Seismic <u>reflection</u> is a response to impedance contrasts, such as those due to vertical lithological variability. Seismic <u>reflectors</u> (in profile) or <u>reflecting surfaces</u> (in three dimensions) suggest the lateral continuity of such variability. The relation of individual reflections to vertical variability is a complex inverse problem addressed within the field of seismic lithology modelling (e.g. Cross and Lessenger, 1988). In contrast, the nature of lateral continuity in relation to reflectors has seldom been formally addressed. Rather, an interpretive model which transforms seismic profiles into geological sections of temporal significance has diverted interest from the physical origin of reflectors to their relation to time (e.g. Christie-Blick et al., 1990). Evaluations of the physical nature of reflectors show that their relation to both chrono- and litho-stratigraphy may depend on seismic resolution.

Vail et al. (1977b) proposed a chronostratigraphical model for shallow marine reflectors that has received universal application in seismic interpretation (reviewed by Cross and Lessenger, 1988). It is argued that while vertical lithological variability causes reflections, lithofacies boundaries are laterally diffuse and will not produce reflectors. Instead, lateral continuity is attributed to unconformable to conformable surfaces (sequence boundaries, see below) which may transgress lithofacies boundaries. The discordant to concordant relations of reflectors observed on seismic profiles are found to be analogous to such surfaces observed in outcrop. The surfaces are argued to form in direct response to hierarchical sea-level fluctuations and therefore be of temporal significance. The apparent regional continuity of many reflectors is offered as empirical evidence for their contemporaneous formation.

The model rests on the physical assumption that reflectors correspond to discretely resolved stratal surfaces. Tucholke (1981) and Cartwright et al. (1993) challenged this assumption, showing that seismic resolution limits may result in apparently discrete reflectors from what are in fact closely-related families of surfaces deposited over time (Fig. 2.6a). Cartwright et al. (1993) observed that the apparent lateral continuity of such complex reflectors may be a function of resolution and in any case renders temporal correlations problematic.

Tucholke (1981) addressed the origins of three regionally extensive deep-sea seismic reflectors in relation to lithology and time. He presented evidence for two diachronous reflectors, one an erosional unconformity related to current activity, the other a conformable lithological boundary attributed to deposition. A third reflector was of diagenetic origin and approximately synchronous. Tucholke emphasised the importance of a clear distinction between seismic, litho- and chronostratigraphies, although he noted the likelihood of "notable coincidences among the three" (p. 23). His work serves to illustrate that a single genetic model is unlikely to embrace all reflectors, in time or space.

Seismic reflectors appear broadly analogous to stratal surfaces in that they record the accumulation of sediment over time and thus they may have a temporal significance in the vertical sense of the principle of superimposition (*cf.* Cross and Lessenger, 1988). However, their physical significance and lateral relation to time will depend on seismic resolution, in a given geological context. Ager (1993) observed that diachroneity is a concept closely related to the temporal scale of investigation (and so to resolution).

2.3.1 b) sequence analysis

Seismic sequence stratigraphy was formalised by contributions in Vail et al. (1977a) and reviewed by Cross and Lessenger (1988). The methodology includes both sequence and facies analysis.

Sequence concepts were advanced (e.g. Sloss, 1963) as a means of correlation and subdivision based on bounding stratal surfaces (unconformities) rather than intervals (lithostratigraphy). Mitchum et al. (1977a, b) extended the concepts to reflection surfaces, emphasising the objectivity of relationships in which "discordance is the main physical criterion used" (1977b, p. 57). Discordant reflector relations are described as toplap, onlap, downlap or trunction and bound reflector configurations such as (sub)parallel, divergent or clinoform (Fig. 2.6b).

Seismic facies analysis refers to the interpretation of vertical or horizontal variations in reflector parameters within or between sequences: parameters include configuration and form, as well as continuity, amplitude, or any distinctive difference (Mitchum et al., 1977a). There are no objective criteria to define a seismic facies boundary, which may be diffuse, and no physical basis for a direct relation with lithofacies. Facies analysis seeks to estimate potential lithofacies variation by analogy with stratal patterns and plan form of inferred geological settings (Cross and Lessenger, 1988).

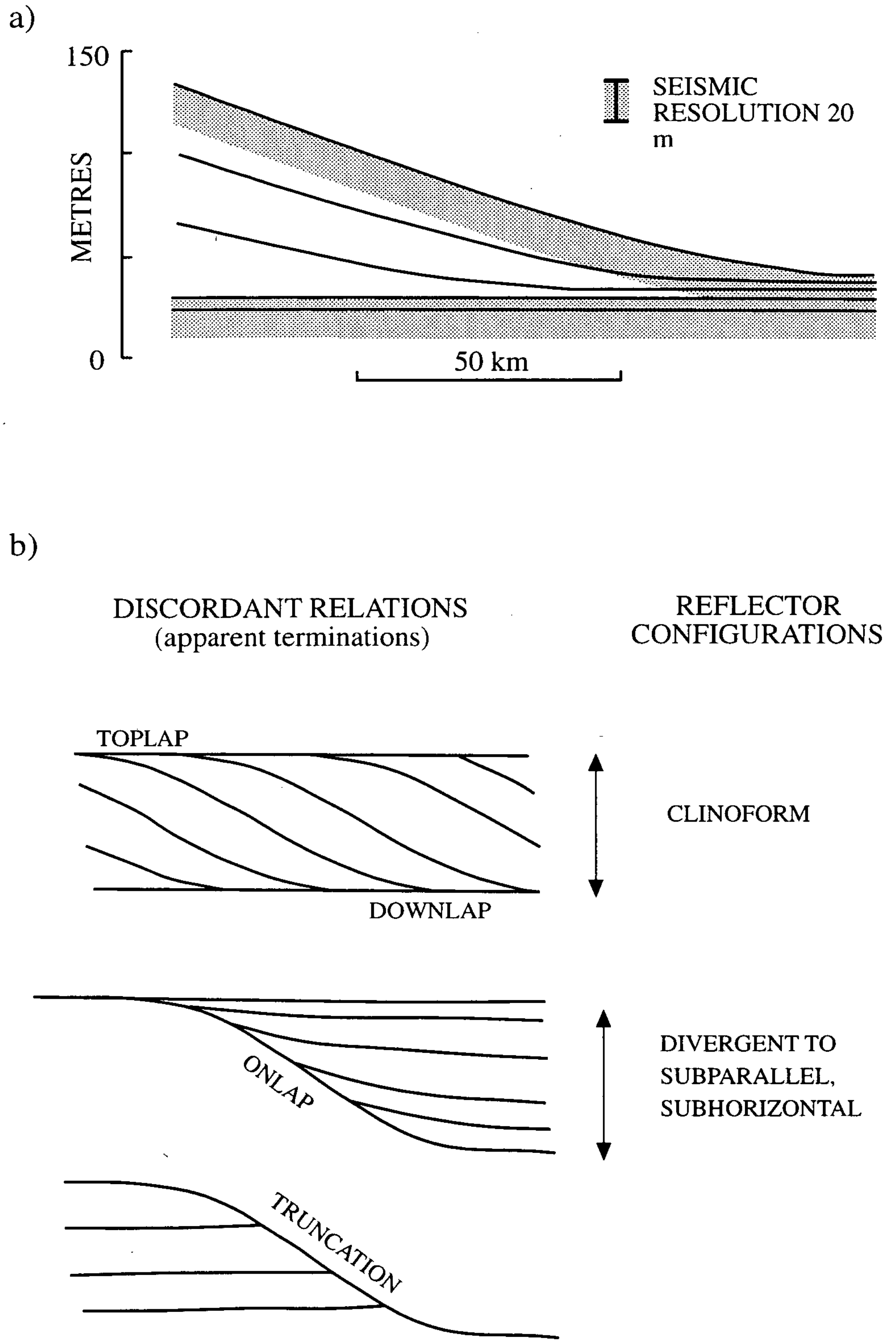


Figure 2.6 - Seismic stratigraphical concepts and terminology:
a) apparent versus true stratigraphical relationships due to seismic resolution limits; from Tucholke (1981) for downlap, also applicable to onlap and toplap.
b) sequence definition from discordant (and correlative concordant) reflector relations.

Seismic resolution limits on the perception of stratal relationships were acknowledged by Mitchum et al. (1977a,b) in a distinction of two kinds of upward discordance, erosional truncation versus toplap due to depositional thinning. They noted that the distinction may be "quite subjective, depending on the angularity of the reflections" (1977a, p. 118) and that for toplap "resolution may be such that reflections appear to terminate abruptly against the upper surface at a high angle" (1977b, p. 59). The same problem was recognised by Tucholke (1981) for downlap (Fig. 2.6a) and generalised by Cartwright et al. (1993) for all reflector relations used to define sequence boundaries (toplap, downlap, onlap). The latter authors concluded that apparently discrete terminations may embrace closely-related families of surfaces and recommended a clear distinction between apparent (seismic) and true (stratigraphical) relationships. I use terms such as toplap to refer to observed reflector discordance (Fig. 2.6b) and erosion and deposition for interpreted stratigraphical relationships.

Seismic sequence analysis provides a methodology to delineate stratal intervals to limits of seismic resolution. Seismic sequences in this sense are informally accepted as material stratigraphical units, comparable to other stratal-bound units such as those of allo-stratigraphy (Owen, 1987).

2.3.1 c) geomorphological analogy

Seismic interpretation was until recently applied mainly to profiles and sequence analysis emphasised regional continuity of reflectors. The proliferation of 3D-seismic data has been accompanied by a renewed appreciation of complex spatial relationships (e.g. Martinsen, 1993) and the interpretive significance of form in influencing the perception of reflector continuity.

Brown (1988) discussed the significance of geomorphological analogy, or physical plausibility, in the interpretation of horizontal seismic sections. In the absence of significant structural variation, amplitude variability alone may allow the recognition of stratigraphical relationships. Channels present the classic example, continuously traceable in plan by amplitude and form even in the absence of reflector discordance (e.g. Fig 2.5b; Brown, 1988; Davies et al., 1992; Praeg, in press - a,b). Stratigraphical discordance may be recognised from analogy of form on horizontal sections, for example cross-cutting buried channels (Praeg and Long, in press) and buried iceberg scours (Newman, 1990; Long and Praeg, in press).

Physical plausibility also operates in vertical profile, where geological surfaces are interpreted that may not necessarily correspond to reflectors. Channels are again a good example. Faults are another, recognised through reflector discordance and continuity of form (by analogy with geological sections) as opposed to reflector continuity. Analogy of form is thus an integral aspect of seismic interpretation where the surfaces of interest exhibit appreciable vertical or horizontal variability.

The tunnel-valley basal unconformity in the study area was traced through a combination of reflector identification and discontinuity of incised reflectors, as well as by tie-line correlation in the case of profile data. It is not a continuous reflector but it is an objectively identifiable geological surface, on both vertical and horizontal sections (Figs. 2.3-2.5).

2.2.4 Study Area Mapping

The grid of seismic data (Fig. 2.1) was used to trace and contour both a) the tunnel-valley base and b) surfaces within the fill. Results are presented as a 50 ms contour map of the depth to the valley base (Fig. 2.7) and as a series of axial and transverse profile lines (A to F) along several larger features (Fig. 2.8). A 1:250,000 map is also provided as an enclosure, showing the basal contours superimposed on the seismic database and profile locations A-F.

All results are presented in milliseconds two-way time (TWT). Depths or thicknesses in metres are given in the text based on a velocity of 1.8 km/s, estimated from identification of the valley base on well data (Chapter 3, Table 3.11). Contour and profile diagrams were not depth-converted, as this would merely have served to linearly change values without accommodating vertical or lateral variations in the velocity of the fill. Vertical exaggerations on all profile diagrams thus vary by up to 20%, due to velocities from sea level in the range of 1.8 km/s (valley base) to 1.5 km/s (seawater).

Exploration data processed by different companies were calibrated to different source levels. I found the variation to be generally <20 ms but locally up to 40 ms. This was not important across relatively steep valley walls, where contours are closely spaced, but posed problems in places for the correlation of gently inclined basal or fill surfaces. Where necessary, recalibration to sea level was effected during contouring by reference either to BGS profiles or (arbitrarily) to existing contour lines.

2.2.4 a) valley base

The valley base was contoured within mapping limits determined mainly by the upper bounds of interpretable exploration data, which varied from 100-200 ms (Fig. 2.7). The mapping limit generally lies along the upper valley walls, as inter-valley 'shoulders' are of poor angular contrast with shallow multiple noise (e.g. Figs. 2.3, 2.4). BGS profiles provide information on the valley shoulders in the UK sector, but the broad line spacing limited interpolation to plan form (see Enclosure). Tunnel-valleys were traced north and south of the exploration data coverage using BGS profiles, but a combination of smaller features and noisy data limited this to between 52°55'N and 53°55'N (*cf.* Figs. 2.1, 2.7). In the Dutch sector, shallow information was limited to two published profiles (Fig. 2.1), the processed RGD line (Cameron et al., 1993) and Section 3 on the 1:250,000 map of the area (BGS/RGD, 1986).

Contour levels (50 ms or better) were transferred directly from profiles to 1:50,000 base-maps and contoured by hand. I did not digitise interpreted reflectors for automated contouring due to the irregular seismic grid spacing and the relatively steep tunnel-valley walls. Also, contouring was an useful means of line-to-line correlation and verification of interpretations. The contour lines were digitised and combined at larger scales using a mapping package (ZYCOR).

In contrast, 3D-seismic data were digitally contoured on a workstation using *Geoquest* interpretation software. Contour lines were drawn directly on successive horizontal seismic sections (Brown, 1988) at 50 ms intervals. Contour levels not divisible by the 4 ms vertical sample interval (e.g. 250 ms) were approximated by the next lowest level (e.g. 252 ms). Contours drawn from time-slices were exported to ZYCOR for combination with other contours (Fig. 2.7). Time-slice contours are shown on Fig. 2.16, along with the locations of horizontal and vertical extracts used in figures.

2.2.4 b) fill profiles

Profiles A to F were constructed along eight larger tunnel-valleys (Fig. 2.8) from both contoured surfaces and 3D-seismic data. Profiles A to C, central D and F represent axial and transverse intersections with contoured reflecting surfaces (Figs. 2.9-2.12, 2.15). Profiles D and E were generated along valley axes within the 3D-seismic volume and reflectors were interpreted directly (Figs. 2.12 - 2.14). Reflector relations along the profiles were then used to define seismic sequences (section 2.4).

Contoured fill reflecting surfaces were traced by hand across the grid of seismic profiles (e.g. Fig. 2.4) and their intersections plotted on the axial and transverse profiles. The intersection profiles were useful interpretive tools, both to correlate surfaces across gaps in the seismic grid (e.g. due to wider line spacing or noisy data) and to clarify the spatial relations of different surfaces. They were also useful to incorporate shallow information from widely spaced BGS profiles - along profiles B to D they contributed to the identification and contouring of reflecting surfaces above 250 ms and were the only source of information on fill surfaces above 150 ms.

BGS profiles were also used along profiles A to D to interpolate the valley-top unconformity, a subhorizontal reflector across the UK sector between depths of 50-70 ms (20-40 ms below the seabed) that records erosional truncation of underlying valley fill reflectors (Fig. 1.9b; BGS/RGD, 1986). It locally includes concave relief along the north of profiles C3 and C4 (Fig. 2.11b) that corresponds to the base of Markham's Hole, a partly-filled Weichselian tunnel-valley with seabed expression. The seabed was interpolated along the profiles from BGS profiles and where necessary from depths recorded at regular intervals along exploration profiles.

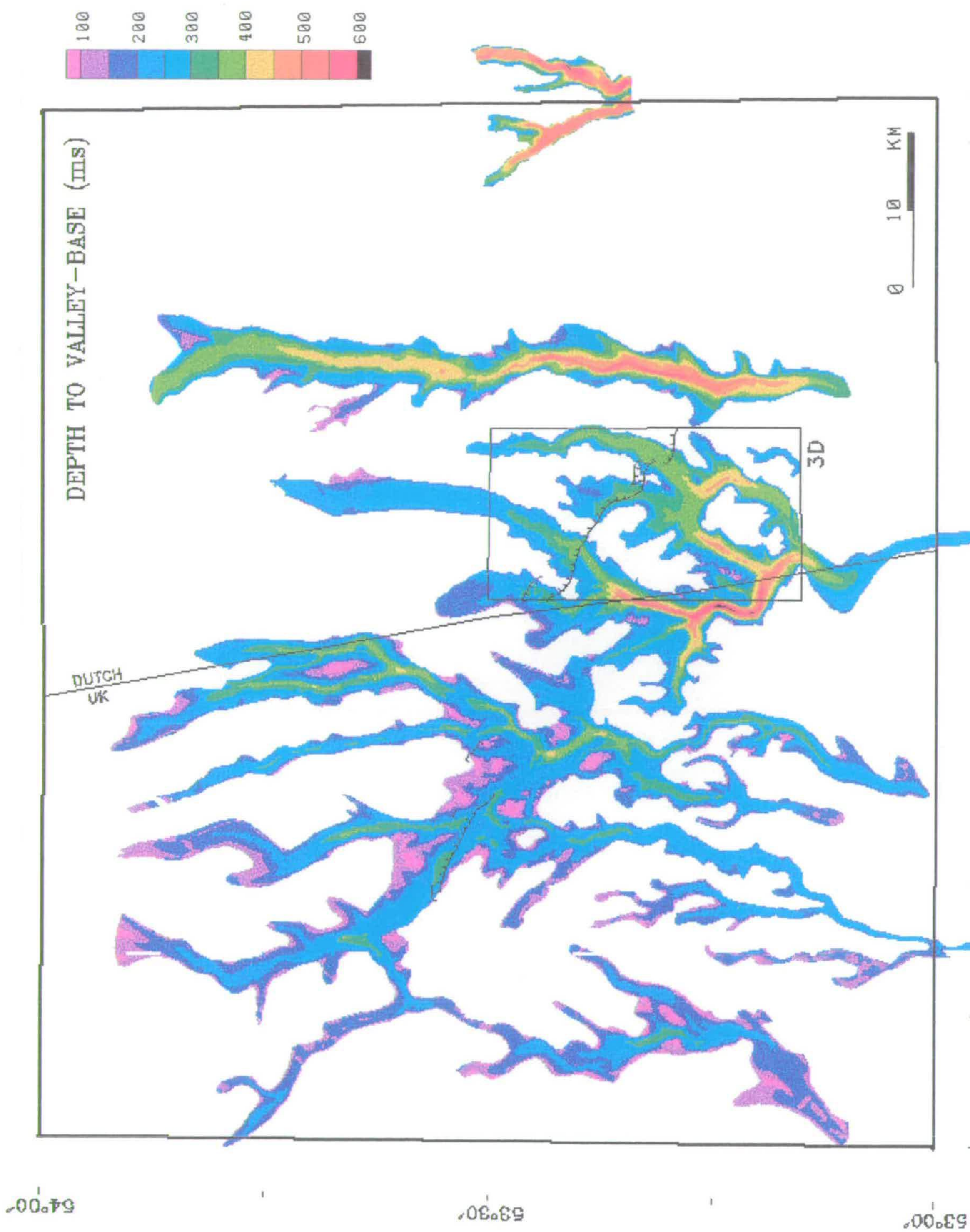


Figure 2.7 - Depth to valley base contours (50 ms interval), within mapping-limits, from data in Fig. 2.1 (see also Enclosure). Hachured lines represent fault traces, which include offset of the tunnel-valleys by up to 10 ms (see Fig. 2.8, profiles C2, D and E). Time-slice contours (3D) shown on Fig. 2.16.

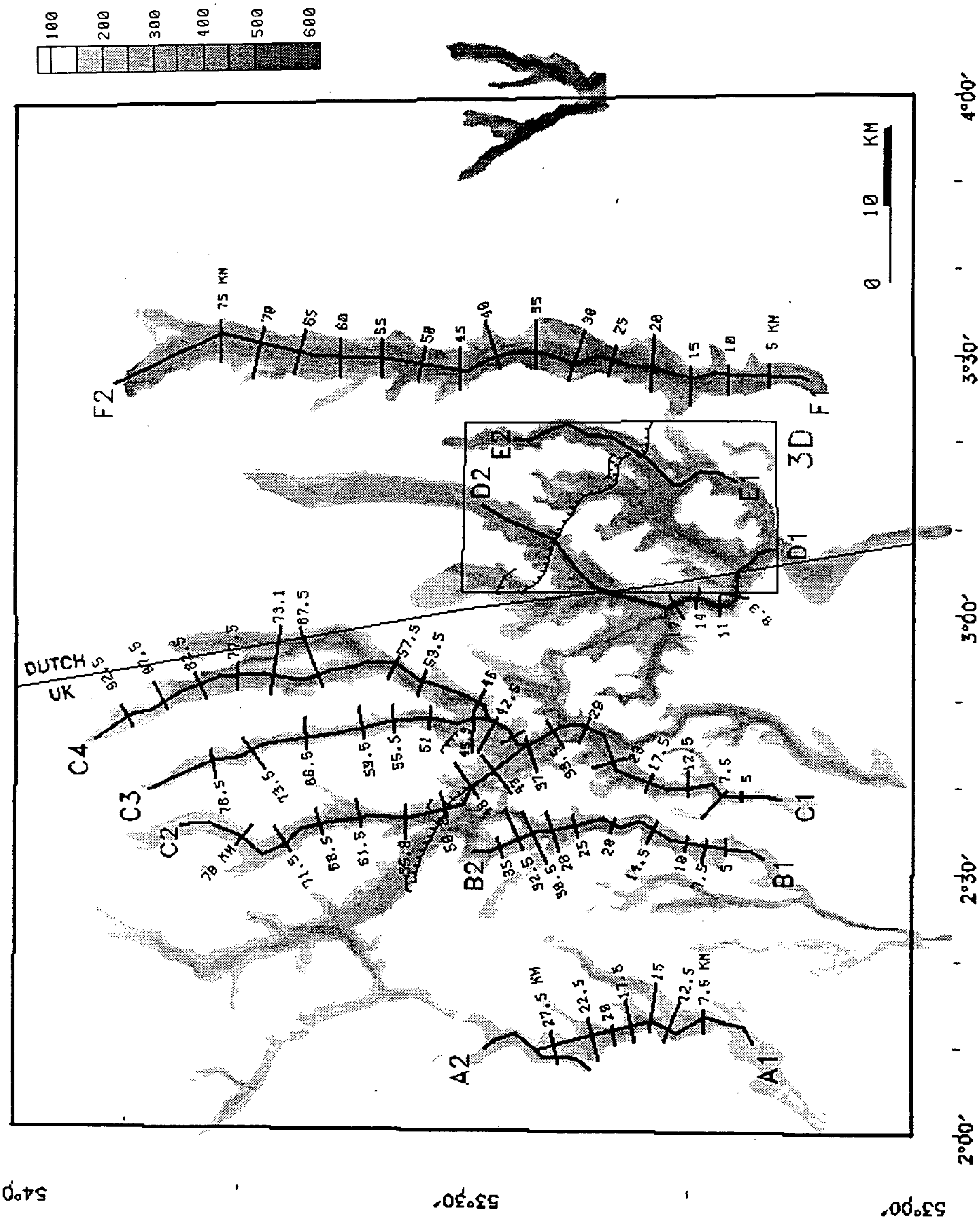


Figure 2.8 - Location of axial and transverse profiles (Figs. 2.9-2.15), superimposed on valley base contours and fault traces (see Fig. 2.7). Numbers are axial distances in kilometres. Time-slice contours and 3D profile locations shown on Fig. 2.16.

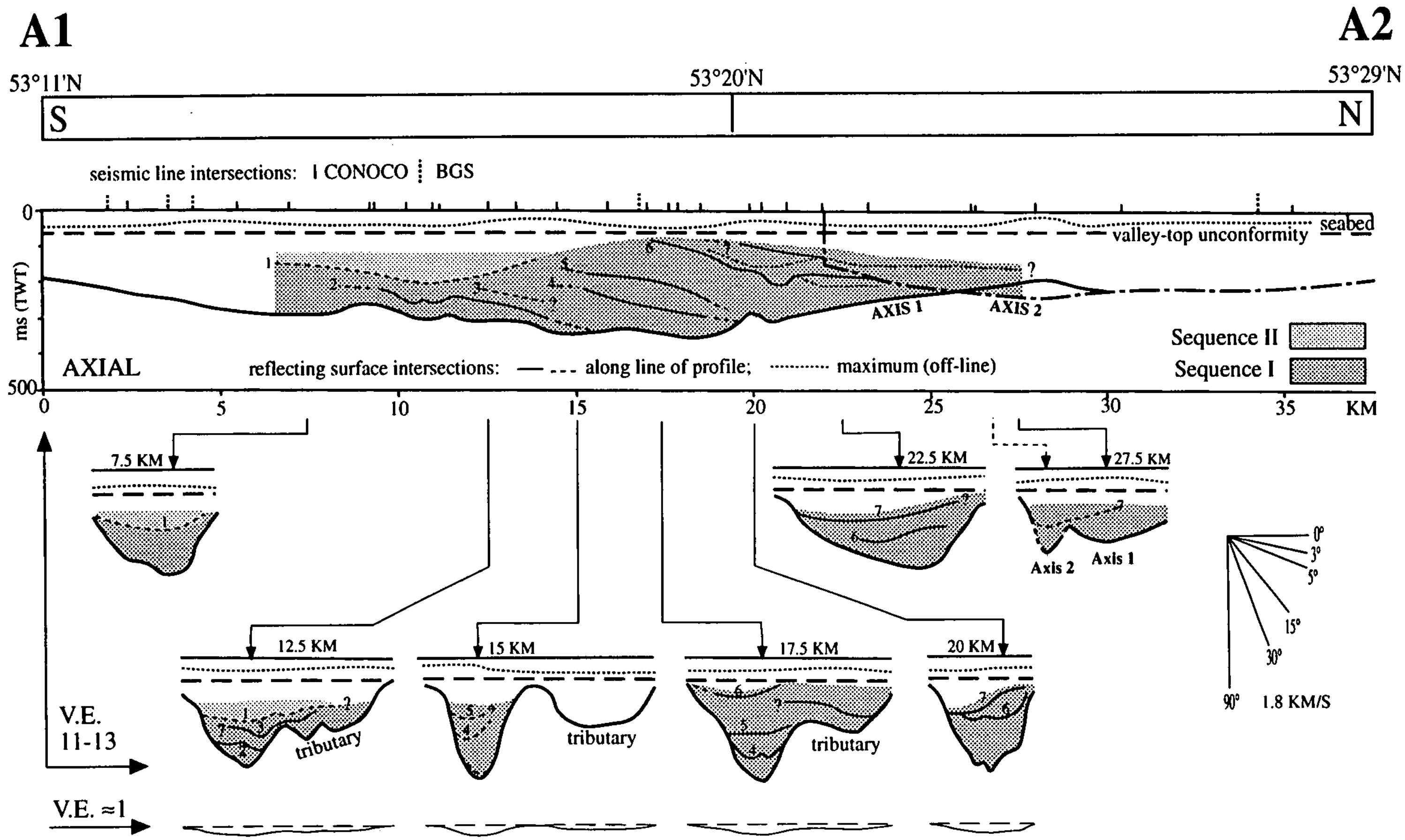


Figure 2.9 - Axial and transverse profiles along tunnel-valley A (Fig. 2.8) showing contoured reflecting surface intersections and interpreted fill sequences. The fill is dominated by axially downlapping clinoform surfaces (sequence I). Subhorizontal surfaces (sequence II) are inferred to occupy the axial basin from 7-14 km and may overlie sequence I in another basin to the north. The axial profile is drawn along the lines of two convergent axes to the north (e.g. cross-section at 27.5 km). Valley-top unconformity interpolated from BGS profiles. See Fig. 2.3 for examples of exploration and BGS reflection data from about 15-20 km along the profile.

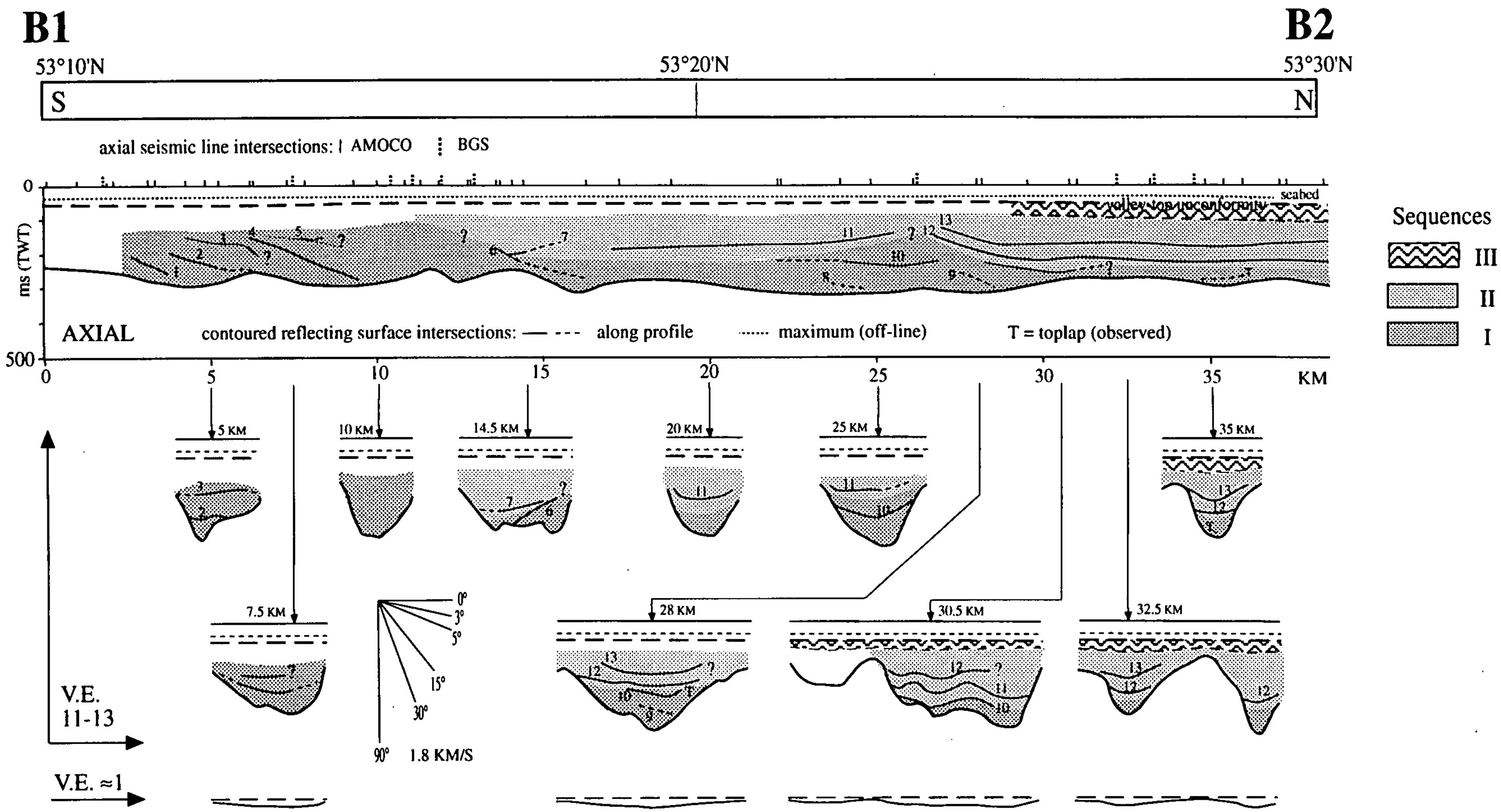


Figure 2.10 - Axial and transverse profiles along tunnel-valley B (Fig. 2.8) showing contoured reflecting surface intersections and interpreted fill sequences. Sequence I thins north of 14 km and is overlain by sequence II. The I/II boundary has been defined in part from BGS profile intersections, but is uncertain south of 15 km. Sequence III corresponds to complex reflectors on BGS profiles above the indicated boundary, which rises to the south. Valley-top unconformity interpolated from BGS profiles.

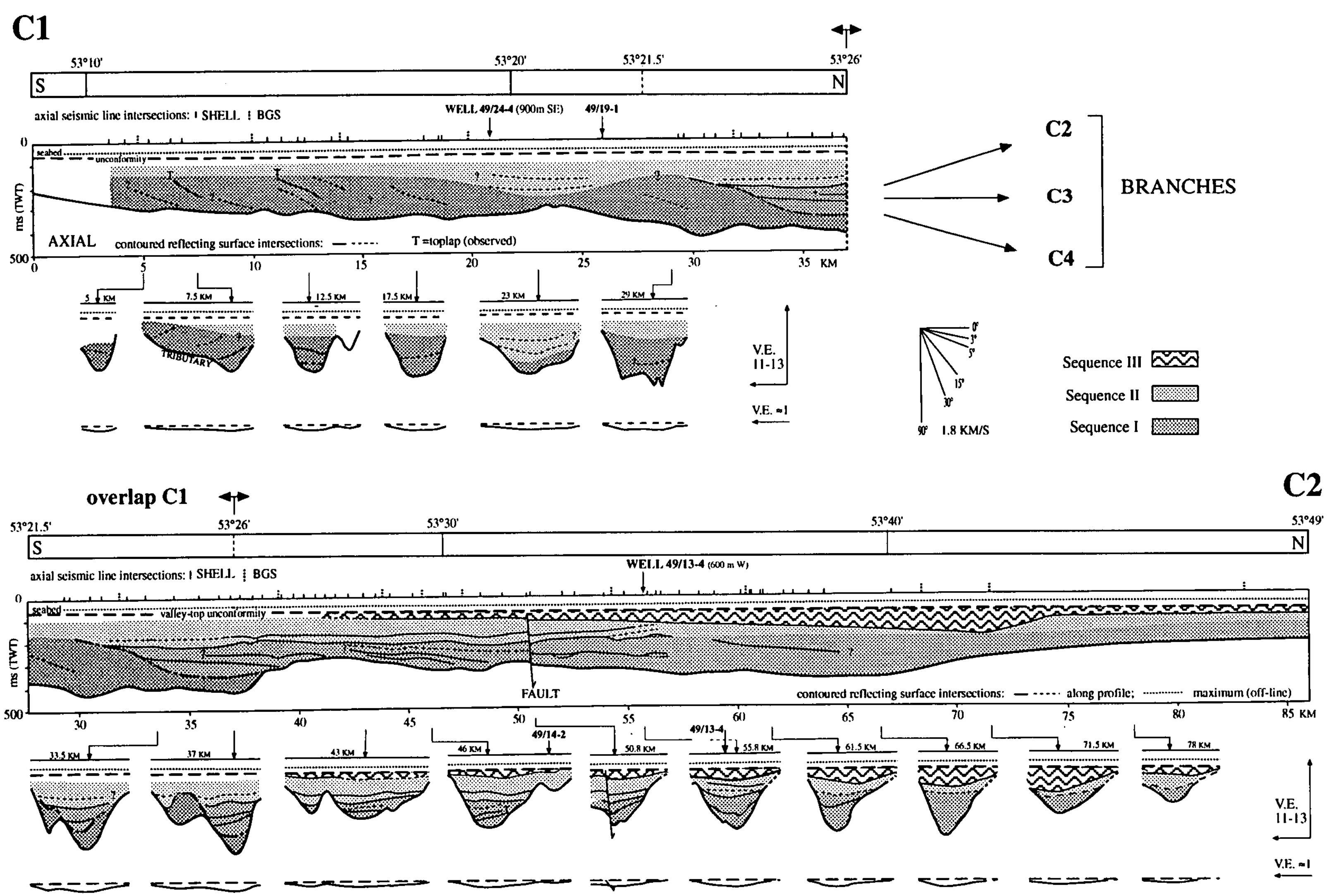


Figure 2.11a - Axial and transverse profiles along tunnel-valley C1 and the first of three northern branches C2 (see Figs. 2.8 amd 2.11b) showing contoured reflecting surfaces and interpreted fill sequences. Branch C2 is cut into C1 along an onlapped surface of discontinuity from 30-38 km (compare Fig. 2.11b). Sequence I clinoform surfaces are of reduced dip in the younger valley. Sequence II (onlapping) includes two surfaces at about 55 km that are upwardly discordant with the base of overlying sequence III, the lower of which downlaps to the south. Sequence III (complex surfaces on BGS profiles) thickens in a basin to the north on profile C2. Sequence boundaries north of 65 km interpolated from BGS profiles. Well locations referred to in Chapter 3.

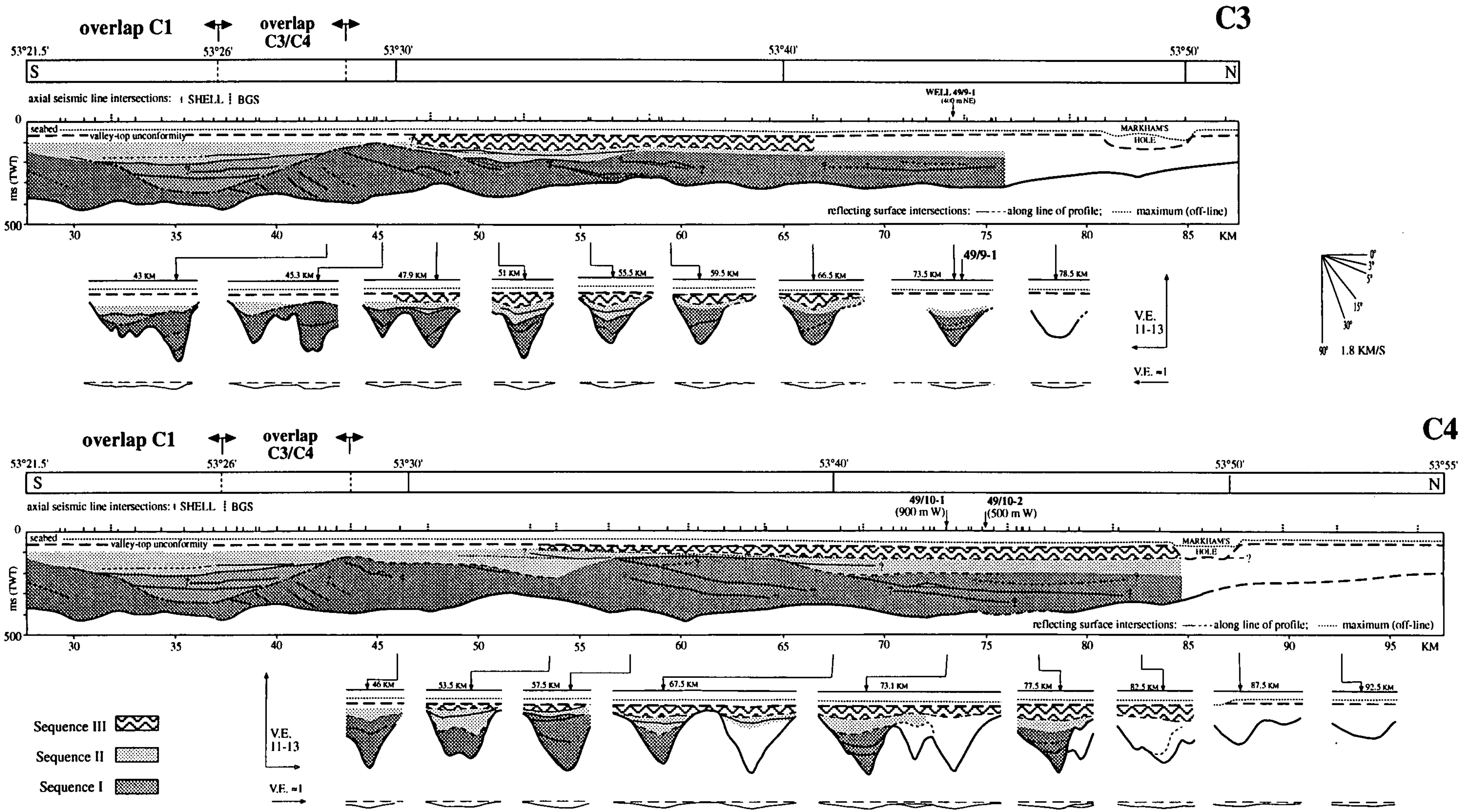


Figure 2.11b - Axial and transverse profiles along tunnel-valley branches C3 and C4 (see Figs. 2.8 and 2.11a). From 30-42 km a younger branch (profile C2) erosionally overlaps along a discontinuity surface (compare Fig. 2.11a). Branches C3 and C4 diverge north of 43 km and in both sequence I clinoform surfaces are of reduced gradient north of the divergence (also profile C2). Sequence II (onlapping surfaces) thickens in basins at similar locations on profiles C2-C4 (see Fig. 2.27). Sequence III (complex reflectors on BGS profiles) thickens to the north in basins in both valleys. Sequence boundaries north of 65 km (C3) and 75 km (C4) by interpolation between BGS profiles. Markham's Hole represents a partly filled Weichselian tunnel-valley at the northern end of both profiles (BGS/RGD, 1986). Well locations referred to in Chapter 3.

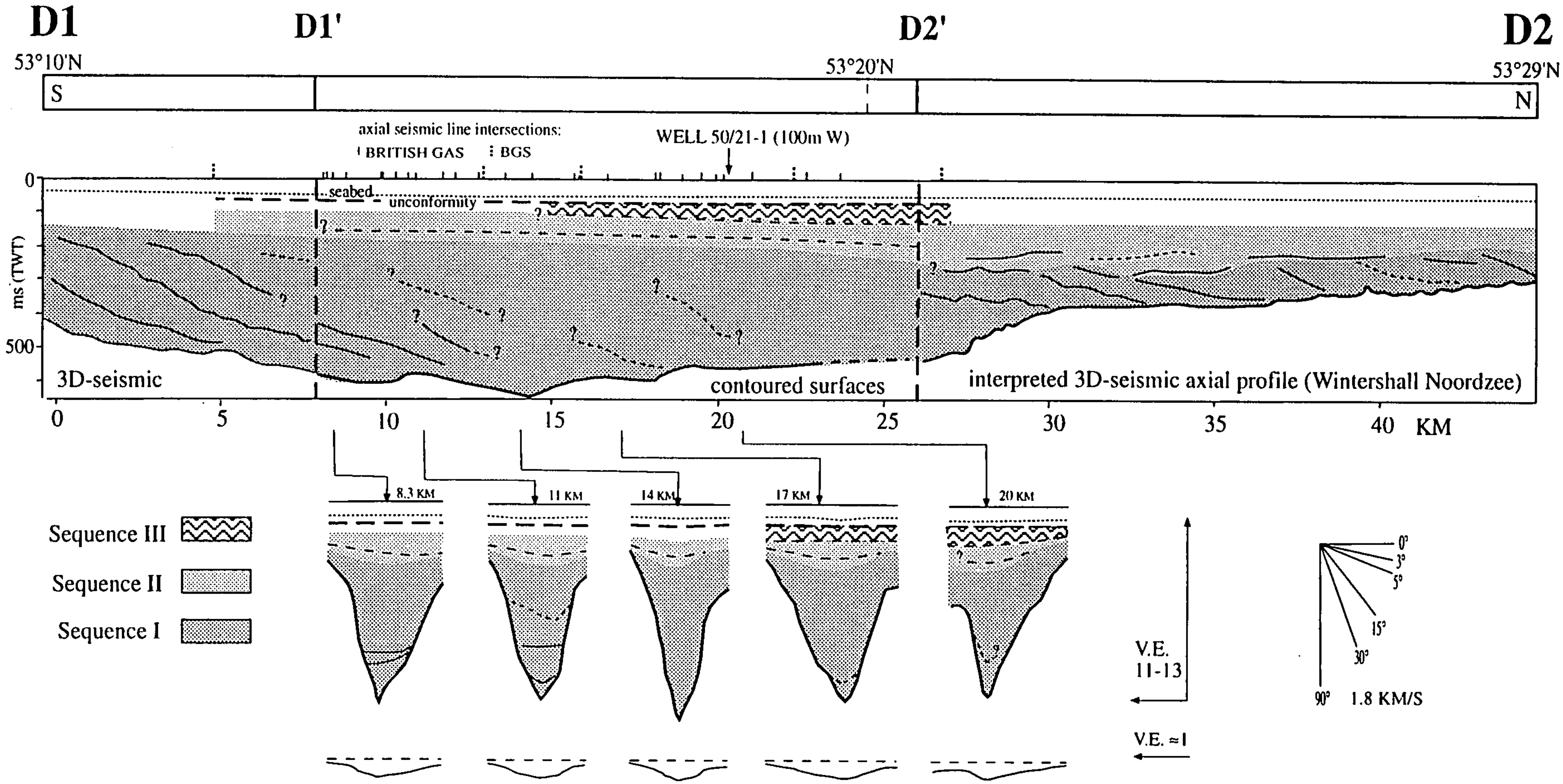


Figure 2.12 - Axial and transverse profiles along tunnel-valley D (Fig. 2.8). The central interval from D1' to D2' (and the cross-sections) show reflecting surfaces contoured from exploration profiles, whereas intervals to either side are interpretations of 3D-seismic profiles generated along the valley axis (see Fig. 2.14). The tunnel-valley includes the greatest depths in the study area (640 ms), and axially downlapping clinoforms (sequence I) are up to 400 m thick. The sequence I/II boundary was roughly contoured from BGS profiles in the central interval and inferred from reflector relations on the right hand 3D-seismic profile (Fig. 2.14). Location also shown on Fig. 2.16.

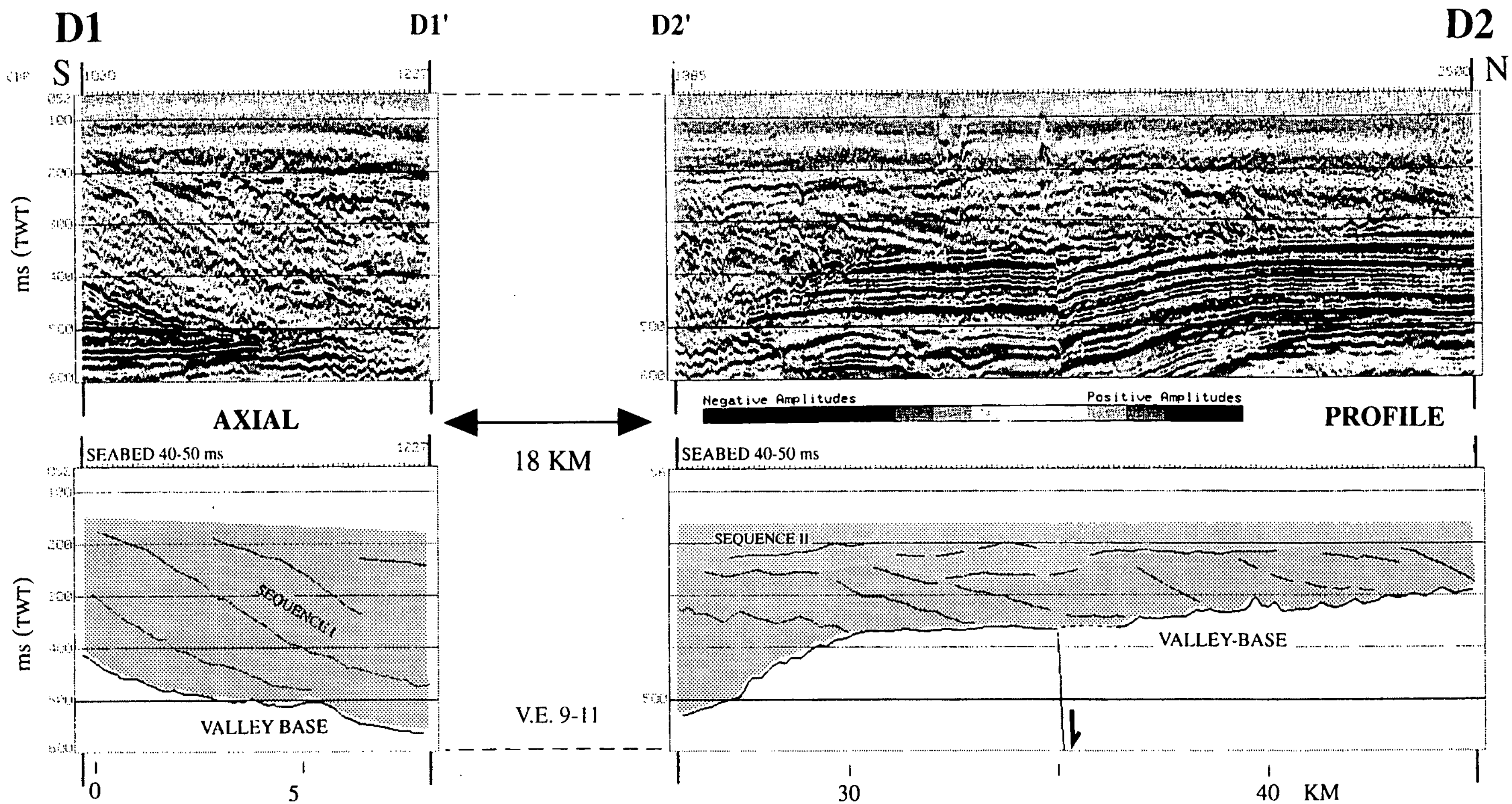


Figure 2.13 - 3D-seismic profiles generated along the axis of valley D (Figs. 2.8, 2.16), and interpretation (see Fig. 2.12 for the central 18 km of the profiles. The fault has offset the incised reflectors by up to 100 ms, and has influenced the valley plan form (see Fig. 2.16), but does not appear to have significantly offset the valley base or fill at this location (compare Fig. 2.14). Dual polarity colour original (see Fig. 2.19).

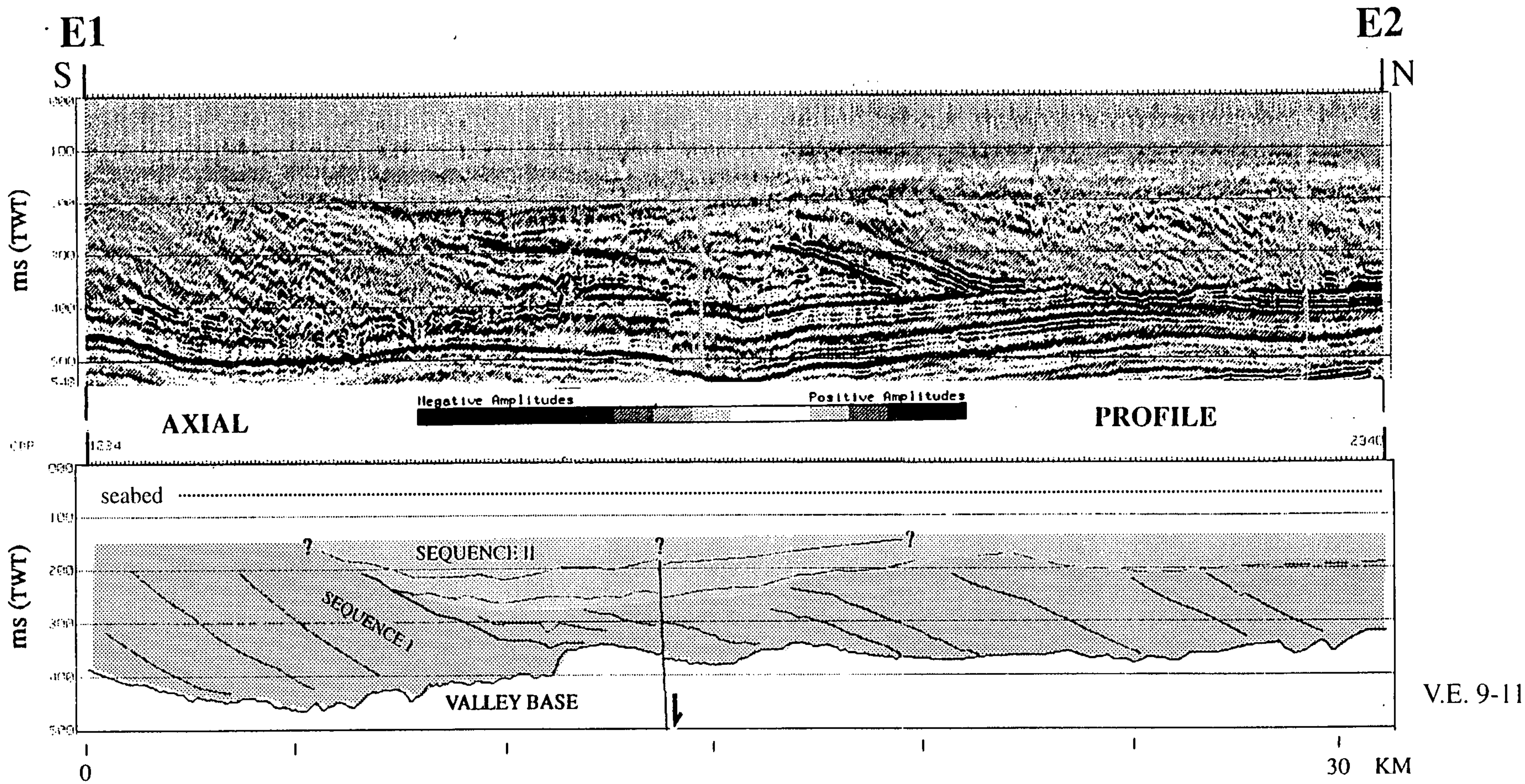


Figure 2.14 - 3D-seismic profiles generated along the axis of valley E (Figs. 2.8, 2.16), and interpretation. Note onlap surface from 6-11 km, which records erosional overlap by younger basal elements to the north (see Fig. 2.28). The I/II boundary does not correspond to a reflector, but is inferred from reflection relations. The upper 150 ms of data are lost to noise. The fault at 14 km has offset the valley fill by about 10 ms, and the underlying incised reflectors by about 30 ms at this location (compare Fig. 2.13). Dual polarity colour original (e.g. see Fig. 2.19).

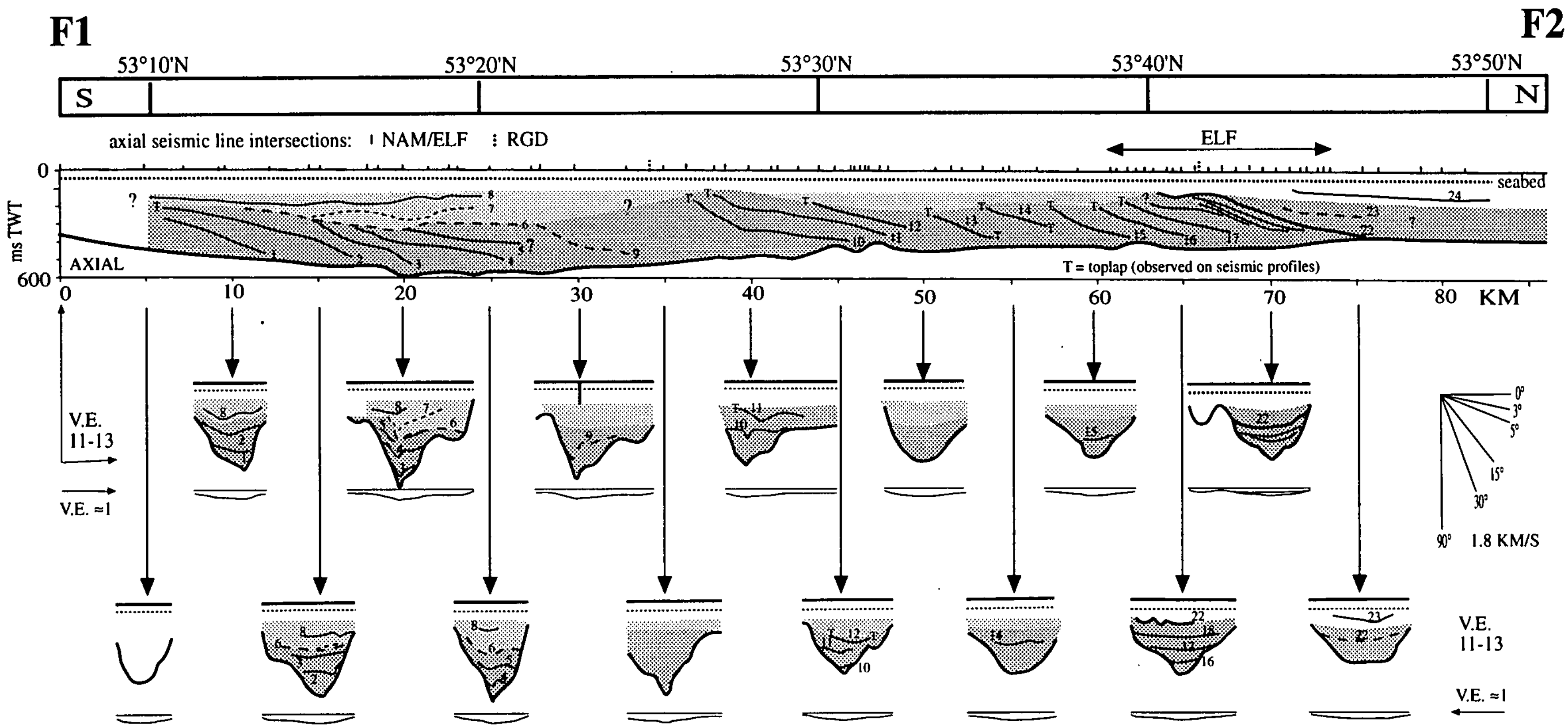


Figure 2.15 - Axial and transverse reconstruction profiles along tunnel-valley F (Fig. 2.8) showing contoured fill reflecting surfaces and interpreted sequences. Based entirely on exploration seismic data, sequence III may be present at the top of the fill succession but is unresolved. A published interpretation of an RGD seismic record (Section 3 - see Fig. 2.1) crosses the profile at 34 km and shows the subhorizontal valley-top unconformity at 70-80 ms. See Fig. 2.4 for seismic profiles from 55-75 km along the profile.

Sequence II

Sequence I

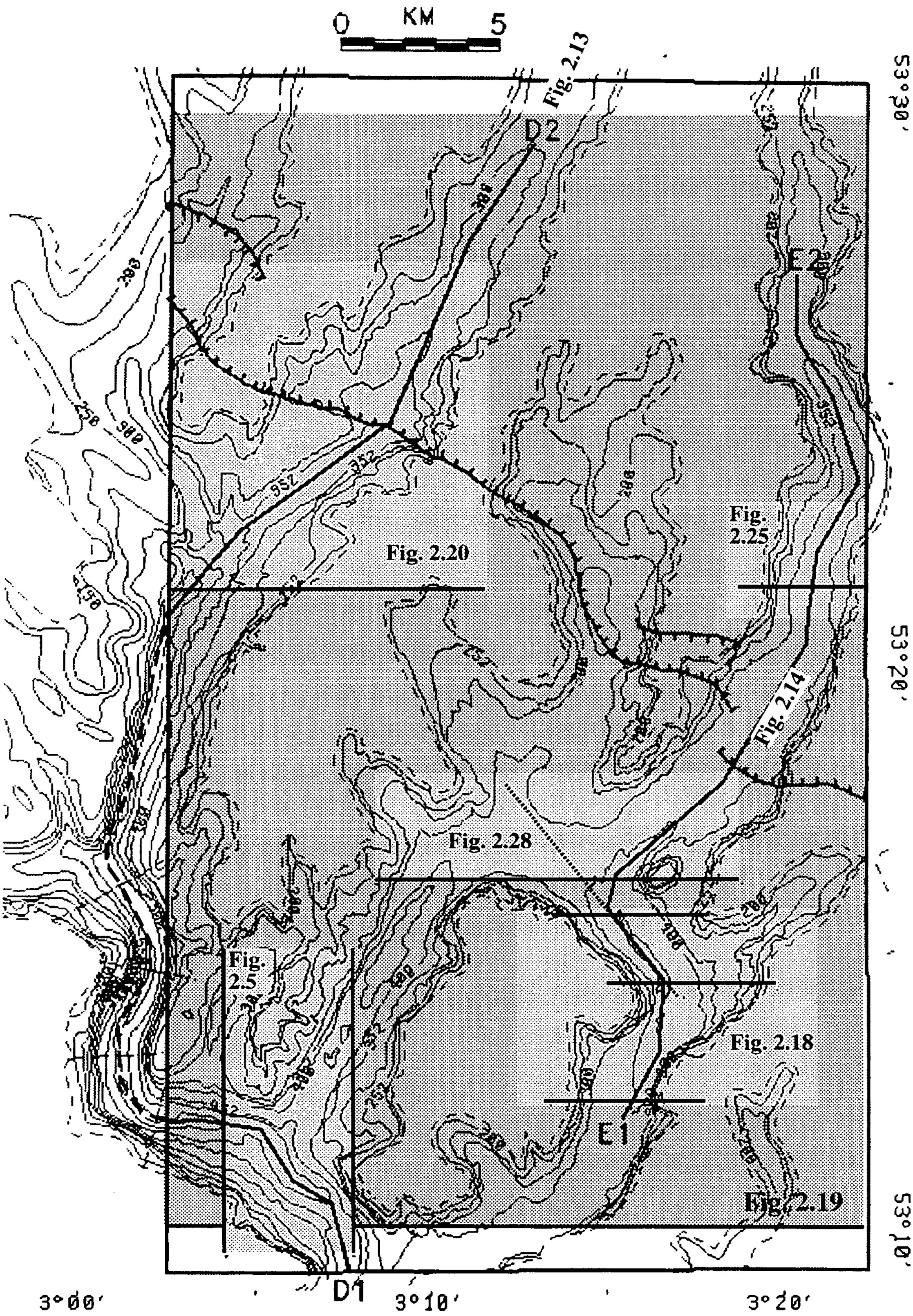


Figure 2.16 - Regional 3D-seismic time-slice contours drawn from eight horizontal sections at approximately 50 ms intervals (200, 252, 300, 352, 400, 452, 500 and 552 ms). Superimposed are the locations of 3D-seismic vertical (bold lines) and horizontal (shaded) extracts used in the indicated figures.

2.3 BASAL MORPHOLOGY

Contours on estimated relief previously suggested the study area tunnel-valleys to comprise a complex anastomosing system of elongate basins up to 450 m in relief and up to 12 km wide (Fig. 1.9; BGS/RGD, 1986; Cameron et al., 1987; Long et al., 1988). Depth contours, drawn from data of closer line spacing, yield a different picture of their dimensions and plan form (Fig. 2.7). The elongate basins are locally up to 500 m in relief (640 ms deep, or 575 m bsl) and include a range of axial sub-basins 10-170 m in relief and 1-45 km long. Cross-sections are up to 6 km wide, relatively steep-sided (5-40°) and variable in form. The basins converge to the south in a series of rectilinear arborescent elements, a pattern complicated in places by anastomosing patterns. Stratal and structural trends in the incised Cenozoic sedimentary substrate have locally influenced cross-sections (benches) and plan form (orientations across faults). An eastwards increase in valley size and spacing correlates with substrate thickness trends.

I describe the basin dimensions (section 2.3.1), their arborescent plan form (2.3.2) and their relation to the incised strata (2.3.3).

2.3.1 Basin Dimensions

The contoured valley base is shown in plan on Fig. 2.7 (also Enclosure) and in section along profiles A to F (Fig. 2.8). Cross-sections (A-D, F; Figs. 2.9-2.12, 2.15) are presented at vertical exaggerations of 11-13 times (sediment-water velocities of 1.8-1.5 km/s) and at a one-tenth scale corresponding to almost no vertical exaggeration (1.1-1.3 times). Contours and seismic profiles across portions of three basins are shown in Figs. 2.17 and 2.18.

Here I summarise information on a) axial relief and b) cross-sectional character.

2.3.1 a) axial relief

The tunnel-valleys extend north and south beyond the seismic data coverage, so that depth contours provide only minimum constraints on overall axial relief (basinal closure). However, regional non-continuity to south and north of the study area (BGS/RGD, 1984 and 1989) suggests closure comparable to maximum cross-sectional relief (trough to shoulder depth - BGS/RGD, 1986). Shoulder depths increase gently eastwards on BGS profiles, from 40-70 m in the UK sector to 70-90 m in the Dutch sector (BGS/RGD, 1986, Section 3).

The tunnel-valleys increase in depth and relief eastwards across the area (Fig. 2.7). West of 3°E, all include depths over 200 ms (180 m bsl) and locally over 400 ms (360 m bsl). East of 3°E, all include depths over 400 ms (360 m bsl) and locally over 600 ms (to 640 ms - profile D, Fig. 2.12). This corresponds to estimated cross-sectional relief in the west of over 120 m and locally over 300 m, versus relief in the east of over 300 m and locally to 500 m (640 ms or 575 m bsl). The latter figure compares with a previous estimate for the study area of 450 m, across the axis of the same valley but 3 km to the northwest (Long et al., 1988).

Undulating axial relief defines sub-basins within all tunnel-valleys over a range of scales, such that larger sub-basins include lesser depressions. The simplest (and largest) example is the valley of profile F, which is divided into two main sub-basins separated by a sill of 400 ms depth (at 47 km - Fig. 2.15), the southern up 590 ms deep, the northern up to 450 ms deep, corresponding to closure of 190 ms (170 m) over 45 km and 50 ms (45 m) over 25 km, respectively. The two main sub-basins each include lesser basins as little as 10 ms in relief and 2 km in length (Fig. 2.15). Axial relief is typically more complex, over vertical scales of 10-100 m and horizontal scales of 1-30 km. For example, the valley of profile C4 contains a sill of 290 ms (at 51.5 km - Fig. 2.11b) which separates several sub-basins to the south from one the north up to 34 km long and 430 ms deep (125 m relief); this northern sub-basin is divided into two roughly equal parts by a sill of 340 ms (at 66.5 km), the southern one 430 ms deep (80 m relief) and the northern one 405 ms deep (60 m relief); each of these two sub-basins in turn include lesser features as little as 2 km long and 10 m in relief.

2.3.1 b) cross-sectional character

The basins are shallow relative to their widths and have relatively steep walls (5-40°). Form and symmetry vary. Steeper gradients are recognised locally in association with cross-sectional benches.

Widths vary from 1-6 km, measured at depths between 150-200 ms. Narrower (0.5-1 km) features were traceable locally above depths of 100-200 ms (Fig. 2.7) and may be more common beyond mapping-limits. There is no simple correspondence between width and depth, neither across the study area nor along individual features. The widest tunnel-valley occurs in the Dutch sector (65 km along profile F - Fig. 2.15), where depths are greatest, but widths up to 5 km also occur to the west in the UK sector. The deepest and widest points in the study area are contrasted in Fig. 2.17 - the deepest point (640 ms) is about 3 km wide, while the widest point (6 km) is only 430 ms deep. The latter lies along northern profile F, the deepest point of which (590 ms) lies to the south and is about 4 km wide (20 km, Fig. 2.15).

Cross-sectional form varies from almost flat-floored ('U-shaped') to sharply concave ('V-shaped') and from symmetrical to asymmetrical, both between and along individual features. The valley of profile F is mainly flat-floored and symmetrical in its northern sub-basin, but sharply concave and asymmetrical in its southern sub-basin (Figs. 2.7, 2.15). However, such a correspondence of cross-sectional form with axial basins is atypical. Neither is there a consistent relationship between cross-sectional and plan form curvature. For example, steeper slopes occur on the concave (outer) side of bends (Fig. 2.17a), along straight sections of valley (Fig. 2.17b) and on the convex (inner) side of bends on opposite sides of a valley (Fig. 2.18).

Cross-sectional gradients (defined by 50 ms contour interval) range from 5-25°, locally up to 30° (e.g. 14 km on profile D, Fig. 2.12). Steeper gradients are recognised in places on seismic profiles, separated by subhorizontal 'benches' on the valley walls (e.g. Figs. 2.17, 2.18). Examples of such benches and intervening steep gradients were mapped in plan along portions of two valleys from seismic profiles (Fig. 2.17). In the western example (50 ms contour gradients 15-30°), gradients up

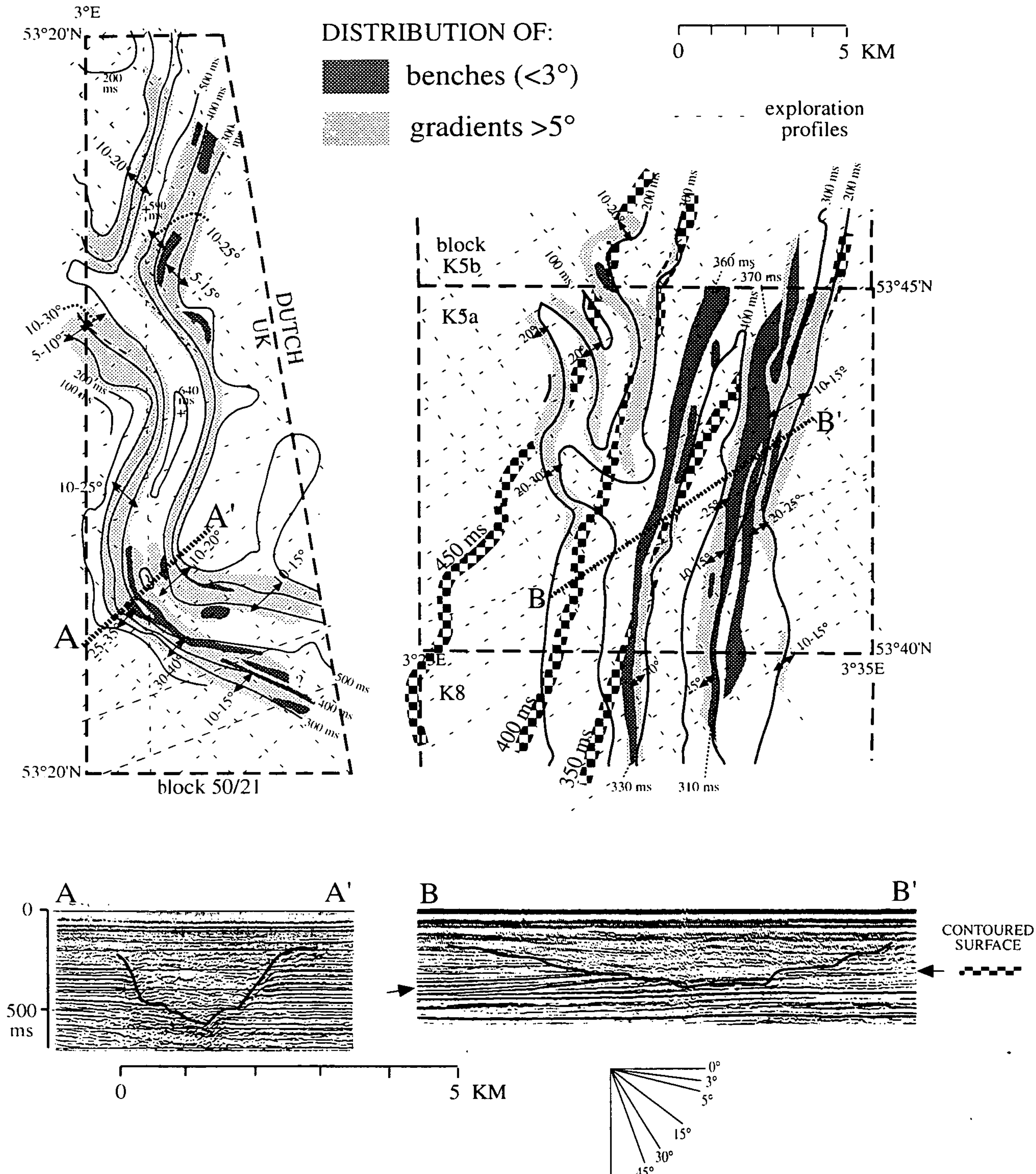


Figure 2.17 - Geomorphological mapping of two tunnel-valley sections, which correspond to the deepest basin in the study area (640 ms, left) and the widest (6 km, right). Both include benches and intervening steep (>5°) slopes in vertical profile, seen to be subhorizontal to discontinuous in plan. On the right, the benches are traceable over 10 km, and descend gently northwards, as do the incised reflectors (contoured at 50 ms interval). Seismic data courtesy of British Gas (A) and ELF (B).

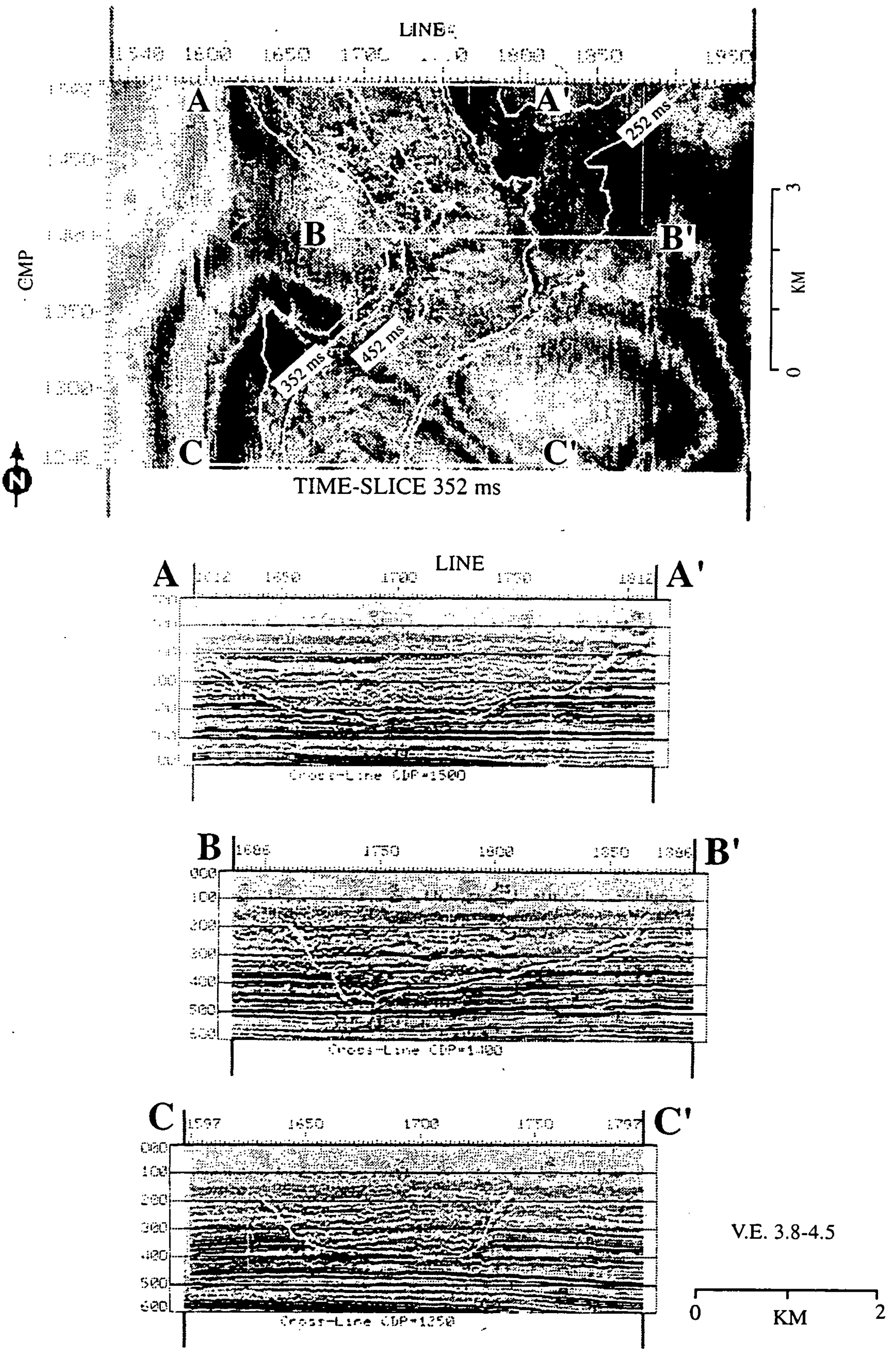


Figure 2.18 - 3D-seismic extracts showing, top, the arcuate plan form of a tunnel-valley on a time-slice at 352 ms (6.4 x 10.2 km area), with superimposed 100 ms contour lines; and below, the variable cross-sectional form on three transverse cross-lines - note varying symmetry, with steeper gradients on profiles B and C on the convex ('inside') of a bend in plan. Location on Fig. 2.16. Dual-polarity colour originals (see Fig. 2.19).

to 40° occur on the southwestern wall above benches traceable along distances of up to 7 km. In the eastern example (contour gradients 5-10°), gradients up to 30° occur on either side of the valley between benches which decline gently northwards (0.1-0.2°) over distances of at least 15 km. The benches are associated with reflectors in the incised sediments on seismic profiles (see section 2.3.3).

2.3.2 Arborescent Plan Form

The elongate basins define a series of rectilinear arborescent elements convergent to the south, complicated in places by anastomosing patterns. Plan form is shown on Fig. 2.8 with profiles A-F for reference. In addition, time-slice contours are shown on Fig. 2.16 and superimposed on two time-slices in Figs. 2.19 and 2.20.

The tunnel-valleys converge southwards to seven main axes along the southern part of the study area (those at 4°E are presumed to converge to a single axis - Fig. 2.8). The arborescent basinal 'elements' gathered by each axis are largely distinct in the Dutch sector, but interconnected to the north in the UK sector (possibly to a greater degree than shown, due to smaller features beyond mapping-limits). The interconnection is in part attributable to the lateral overlap of adjacent arborescent elements. However, anastomosing patterns occur in places, as divergence around isolated elevations or 'islands'. Within the area of 3D-data, anastomosing occurs around two 'islands' (one large and a second to the northeast much smaller) which lie to the south of two main convergent branches (Fig. 2.16). In the central UK sector, anastomosing occurs around several 'islands' along the southwest side of profile C2 (30-50 km, Fig. 2.8) and along the valley of profile C4 from 60-75 km (Fig. 2.8). These locations will later be shown to coincide with fill discontinuities recording overlap by younger cut-and fill elements to the north (section 2.5.1), the result of which is an 'apparent' anastomosing.

The convergent pattern can be described via a Horton ordering system in which 'trunk' order increases with the convergence of 'branches' of the same order (e.g. Selby, 1985, p. 298). All arborescent elements are then order 2 or 3. Order 2 describes trunks with individual branches (e.g. elements of profiles A, C1, C2-C4) while order 3 describes branches which themselves branch (e.g. element of profile B, that at 4°E). These values are likely to be minimal since many smaller tributaries may be present beyond mapping-limits. For example, the valley of profile F is order 3 by virtue of a single branched tributary locally resolved by shallow data (west of 60 km, Fig. 2.8). An ordering system also fails to adequately describe either the locally complex convergence of more than two axes to the same point (e.g. profile D, Fig. 2.16, 2.20) or the rectilinear morphology.

A rectilinear or angular-offset morphology is expressed both in the parallelism of adjacent basins prior to their convergence and in the segmentation of individual basins. Parallelism is apparent despite orientations varying from the overall north-south trend by as much as 90° (west-east). In the western half of the area, for example, basins in the south are mainly oriented SSW-NNE, while those in the north are mainly oriented SSE-NNW. Parallelism of adjacent tributaries is noted prior to their angular convergence with a larger trunk (e.g. elements of profile A, F - Fig. 2.8) or with each other (elements of profiles B, C1). Along southern profile D, two subparallel branches converge southwards to join the main valley at an angle of 90° (Fig. 2.16).

Segmentation is expressed in the angular offset of relatively straight axial segments 2-20 km long. Extreme examples are the basins of southern profiles D and E, where individual segments up to 5 km long meet at almost 90° (D) and 70° (E) (Fig. 2.16). In general, segments (and branches) converge at angles of less than 45°. An en echelon pattern is apparent locally, along northern profile A where an angular offset corresponds with a separate valley axis to the north (Figs. 2.8, 2.9). This and some other locations of angular offset also coincide with probable surfaces of discontinuity recording overlap by younger basal elements to the north (section 2.5.2).

2.3.3 Relation to Substrate

The tunnel-valleys are incised into subhorizontal Cenozoic sedimentary strata which thicken to the east and are halokinetically deformed and in places offset by faults (see Figs. 1.6b, 1.7, 1.9). Substrate influence on the morphology of the valley base is recognised at three successively larger scales: a) discontinuous benches on the walls which follow stratal (reflector) trends, b) valley orientations which locally follow fault trends and c) size and spacing which increase with substrate thickness. The first two represent local influences on cross-sectional and plan form, respectively, while the second is a regional influence on distribution.

2.3.3 a) stratal benches

Subhorizontal benches along the valley walls appear to follow reflectors in the incised strata on seismic profiles (Figs. 2.17, 2.18). The correspondence with reflectors is confirmed in the eastern of the two examples on Fig. 2.17, where prominent benches on either side of the basin descend gently northward (0.1-0.2°) over distances of 15 km. Contours on the incised reflector with which the benches are associated (within Plio-Pleistocene sediments between 330-380 ms) show that the trend of the valley is oblique to the strike of the surface, which dips 1-1.5° WNW (Fig. 2.17). The reflecting surface declines northward about 0.1° along the trend of the valley, a gradient comparable to that of the benches.

The Plio-Pleistocene fluvio-deltaic succession comprises semi-consolidated sandy versus muddy sediments (BGS/RGD, 1986; Pegler, 1995). The benches appear to record differential erosion of such stratal contrasts during excavation of the basins. They are secondary to the overall broad, shallow and steep-sided cross-sectional form of the tunnel-valleys (Figs. 2.17, 2.18).

2.3.3 b) orientations across faults

The plan form of the tunnel-valleys includes orientations that range from transverse to parallel to incised structural trends. This was noted in a general way during mapping and is illustrated by a 3D-seismic horizontal section over a 22 x 36 km area (Fig. 2.19) which reveals discordance with small-scale halokinetic trends. The section also provides evidence for a local influence on plan form by faults.

The horizontal section (valley base at 340 ms, plus superimposed contours) provides information both on tunnel-valley orientations and on structural trends within the incised sediments. Two main trends are apparent: i) sinuous to circular configurations, corresponding to halokinetic domes or basins and ii) a set of faults running NW-SE across the northern half of the area, which offset the gently westward-dipping incised strata by up to 100 ms (profile D, Fig. 2.13) and the tunnel-valleys by up to 10 ms (profile E, Fig. 2.14). The tunnel-valleys locally parallel the sinuous reflector configurations (e.g. valley of profile D1-D2, southernmost valley of profile E) but in the main cut across them. Similarly, the overall trend of the valleys is directly across the faults.

However, two tunnel-valleys change orientation across the fault, that of profile D and that immediately to its southeast between profiles D and E (Fig. 2.19). The latter shifts northwards from a NNE to NW orientation so that its steep southern wall follows the fault surface, whereas branches to the north remain NNE. The valley of profile D shifts northwards from NE to NNE and widens along the fault surface (Fig. 2.20). Down-faulting of the valley base to the north is negligible at this location (profile D - Fig. 2.13), so the widening is mainly related to lateral erosion along the fault surface.

In contrast, no change is observed along tunnel-valleys to the west or east where they cross the faults on the horizontal seismic section (Figs. 2.16, 2.19). Similarly, west of the 3D-seismic area tunnel-valleys are oblique to prominent faults (Fig. 2.8) which follow the overall NW-SE halokinetic structural grain (see also Fig. 1.6b). Faults have thus only influenced the orientations of some tunnel-valley segments.

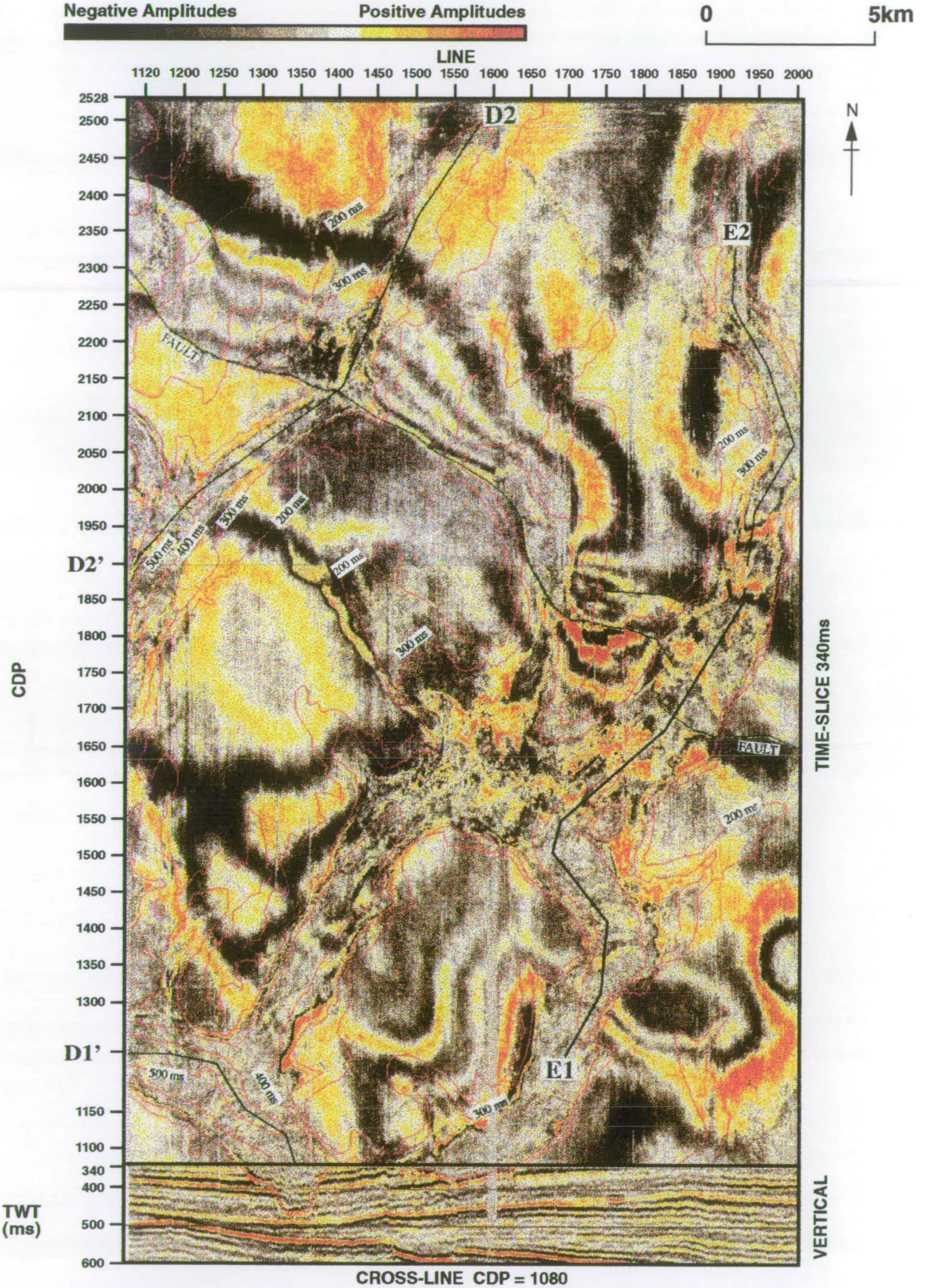


Figure 2.19 - Horizontal seismic section at 340 ms of a 22 x 36 km area with vertical cross-line at bottom (340-600 ms). Superimposed valley base contours (100 ms) were traced from successive time-slices. The tunnel-valleys cut across circular to arcuate reflector trends (halokinetic domes and basins) and across faults in the north which have locally influenced plan form. Axial seismic profiles D and E (Figs. 2.13, 2.14) shown for reference. Location on Fig. 2.16.

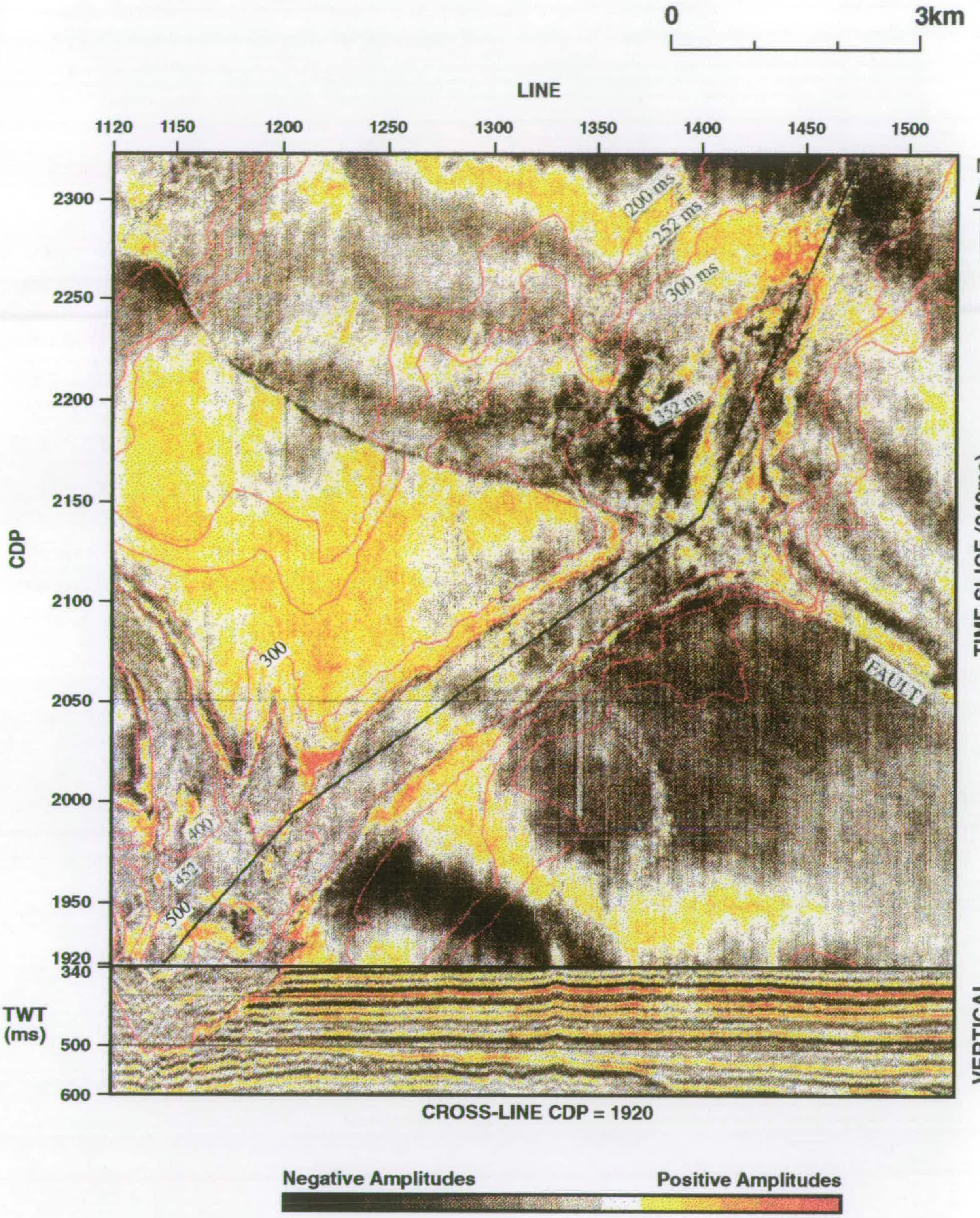


Figure 2.20 - Horizontal seismic section at 340 ms across a 10 x 10 km area with vertical cross-line at bottom (340 -600 ms). Superimposed valley base contours traced from successive time-slices (a strict 50 ms interval is precluded by the 4 ms digital sample rate). Several tunnel-valley branches converge south to a single axis over 500 ms deep. Black line is a portion of axial profile D (Figs. 2.12, 2.13). Note change in axial orientation and lateral expansion across the fault trace, which has not significantly offset the valley base at this location (Fig. 2.13). Location on Fig. 2.16.

2.3.3 c) size, spacing and thickness

A correspondence of tunnel-valley size and spacing with substrate thickness is illustrated by two west-east sections across the study area (Fig. 2.21). The Pleistocene base is shown for reference. It is conformable to the east within an upward-coarsening deltaic mega-sequence, which onlaps older Tertiary muds and Cretaceous chalks in the west. Its eastward descent thus approximates an increase in the thickness of sandy delta-front and delta-top sediments (see Fig. 1.7a). The tunnel-valleys extend to or below the base of the Pleistocene succession, locally into older Tertiary or Cretaceous strata along the western margin of the area (BGS/RGD, 1986, Section 3).

Tunnel-valley axial depths are shown as encountered along sections at 53°15'N and 53°45'N (and as maxima ±5° on either side). The basins increase eastwards in depth and so in relief. On the southern section, a marked increase in depth near the centre of the area (3°E) from less than 320 m (360 ms) to over 420 m (470 ms) closely corresponds with an increase in substrate thickness. On the northern section, the eastwards increase in depth is less pronounced although depths west of 3°E of mostly <360 m (400 ms) compare with a maximum of 400 m (440 ms) in the east. Tunnel-valleys in the western half of the area thus become slightly deeper to the north, while those in the east become shallower.

The basins also become more widely spaced to the east (Fig. 2.21). On the southern section, nine axes between 2-3°E (67 km) are 3-11 km apart (average 7 km/axis), while four between 3-4°E are 11-38 km apart (17 km/axis). On the northern section, six axes west of 3°E are 2-14 km apart (11 km/axis), while two axes in the east are 18 km apart (33 km/axis). The number of axes decreases to the north on both sides of the area.

An eastwards increase in tunnel-valley size and spacing is thus moderated by variations from north to south. The locally close correspondence with substrate thickness trends suggests a regional influence on the erosional morphology.

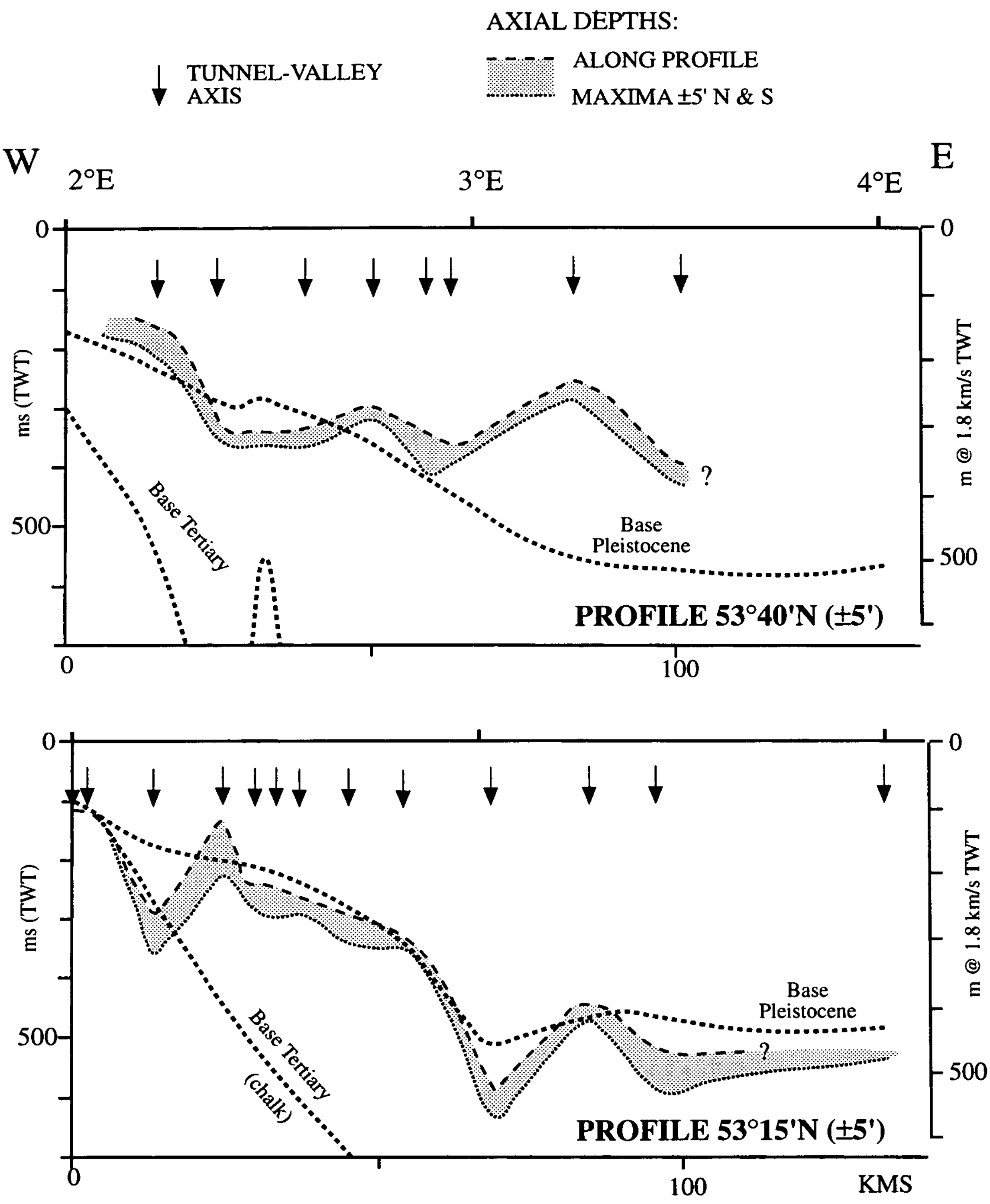


Figure 2.21 - W-E profiles across the northern and southern halves of the study area, showing tunnel-valley axial depths (and spacing) increasing in parallel with Cenozoic sediment thickness. Axial depths are shown both along the lines of profile, and as maxima in a ±5' zone 5' either side of the lines. The southern profile includes the deepest tunnel-valley in the study area (640 ms). Depths to base Pleistocene from BGS/RGD (1986 - reconverted from metres to TWT at 1.7 km/s), base Tertiary from BGS/RGD (1987 - converted to TWT at 1.8 km/s).

2.4 FILL STRATIGRAPHY

Seismic facies analysis of individual BGS/RGD profiles, mainly oriented obliquely to tunnel-valley axes, previously allowed the generalisation of three fill units (Fig. 2.22): 1 - lower structureless or chaotic, 2 - middle layered and 3 - upper complexly layered (Cameron et al., 1987; Balson and Jeffery, 1991). Here I present a seismic sequence analysis based on the axial and transverse relations of fill reflecting surfaces along profiles of eight larger basins (Fig. 2.8). Three sequences are identified, separated by discordant boundaries of basinal relief. They are shown by stipple on the profiles and schematically in Fig. 2.23: I - axially downlapping clinoform surfaces, II - onlapping surfaces and III - complex surfaces.

The sequences are correlative to the three BGS/RGD facies units but are in part of different character. They were defined using a combination of exploration data (at depths greater than 100-200 ms - Figs. 2.3, 2.4) and intersections with BGS profiles (depths above 250 ms in the UK sector - Fig. 2.22). Sequence I is seen to be structured rather than chaotic on exploration data. Sequence II is defined from both exploration data and BGS profiles and sequence III is recognised only on BGS profiles. The subhorizontal valley-top unconformity was interpolated from BGS profiles in the UK sector (Fig. 1.9b; see section 2.2.4b).

The sequences and their discordant boundaries are interpreted in terms of erosion and deposition by stratigraphical analogy. The lower fill (sequence I) records progradation to the north, beneath an upper boundary of depositional relief. The upper fill records deposition within the basins (sequence II), separated by an upper erosional boundary from a second interval of deposition in basins (sequence III). Sequences II and III are eroded against the valley-top unconformity.

I describe the three sequences and their two intervening boundaries in terms of reflecting surface relations (sections 2.4.1 to 2.4.5) and interpret them in terms of erosion and deposition (2.4.6).

2.4.1 Sequence I (Clinoform Surfaces)

Sequence I comprises clinoform reflecting surfaces 3-20 km long that axially downlap the valley base at gradients up to 3.5°, axially toplap against overlying sequence II (see section 2.4.2) and dominate the fill along profiles A to F with thicknesses up to 400 m (Figs. 2.8-2.15).

The sequence is described as structureless or of chaotic reflector configuration on BGS profiles (e.g. Fig. 2.22; Cameron et al., 1987). The lack of structure is attributable to low signal to noise ratios at depth (e.g. Fig. 2.3). On exploration profiles, the clinoform surfaces vary in continuity and amplitude but are recognised on transverse, oblique and axial profiles (e.g. Fig. 2.4). Prominent reflectors were traced and contoured as data quality and density permitted at axial intervals of 1-20 km (profiles A-D and F, 67 examples). Along 3D-seismic axial profiles they were traced directly at intervals of 2-5 km (profiles D and E, 21 examples). Examples of surfaces contoured from seismic profiles are shown in Fig. 2.24 and an example contoured from 3D-seismic horizontal sections in Fig. 2.25.

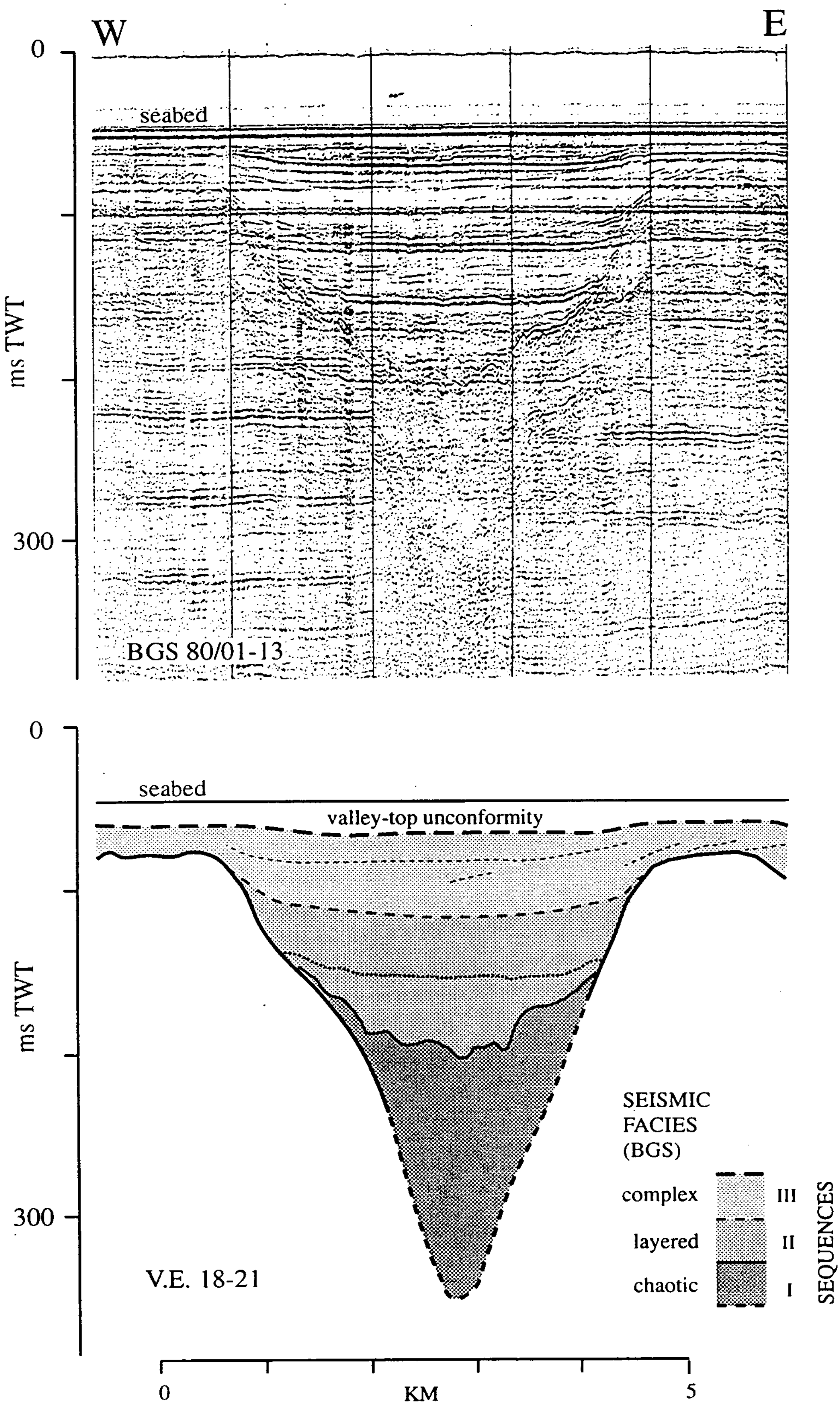


Figure 2.22 - BGS reflection profile of superior quality (cf. Fig. 2.3C), and interpretation of valley fill seismic facies/sequences. The profile transversely crosses the tunnel-valley of profile C4 (69 km - Fig. 2.11b). The lower (dashed) valley base has been interpreted by reference to contoured exploration data.

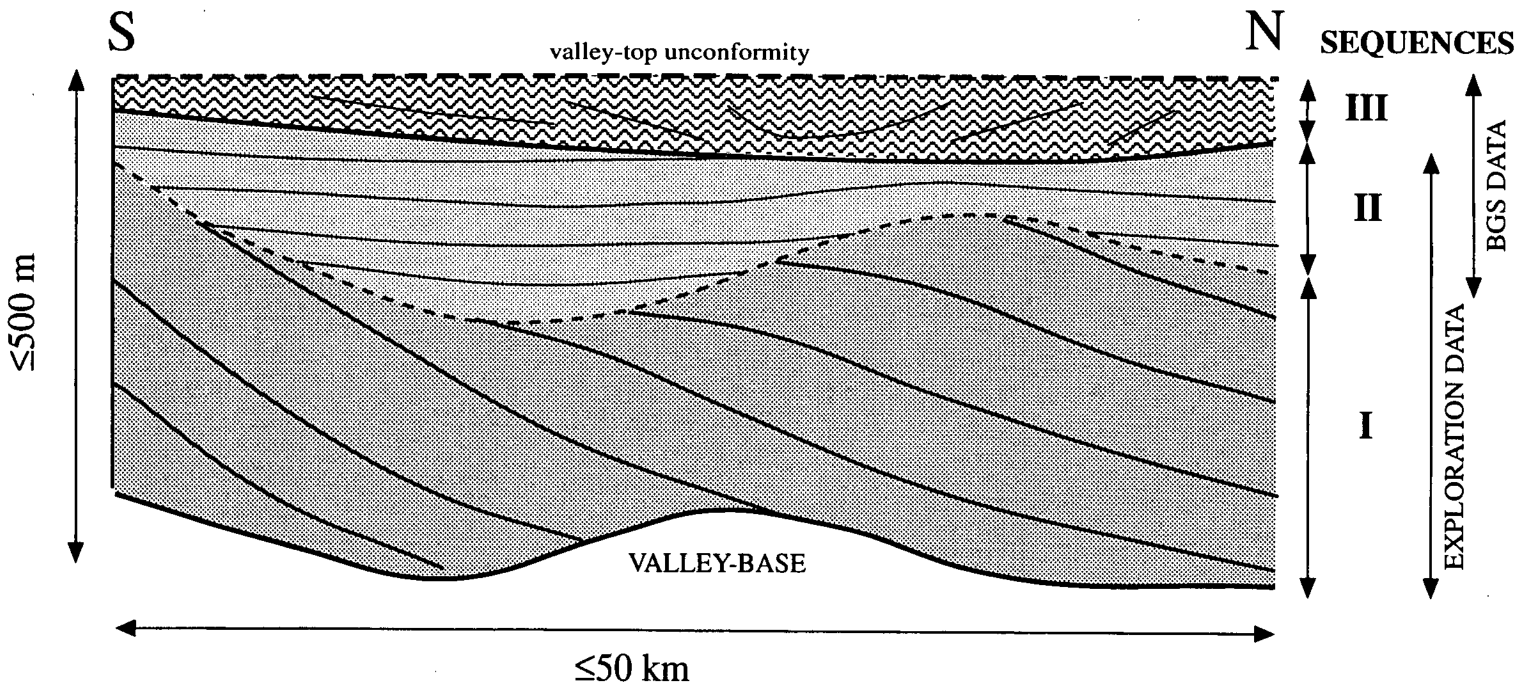


Figure 2.23 - Schematic axial reflector relations and interpreted fill sequences.

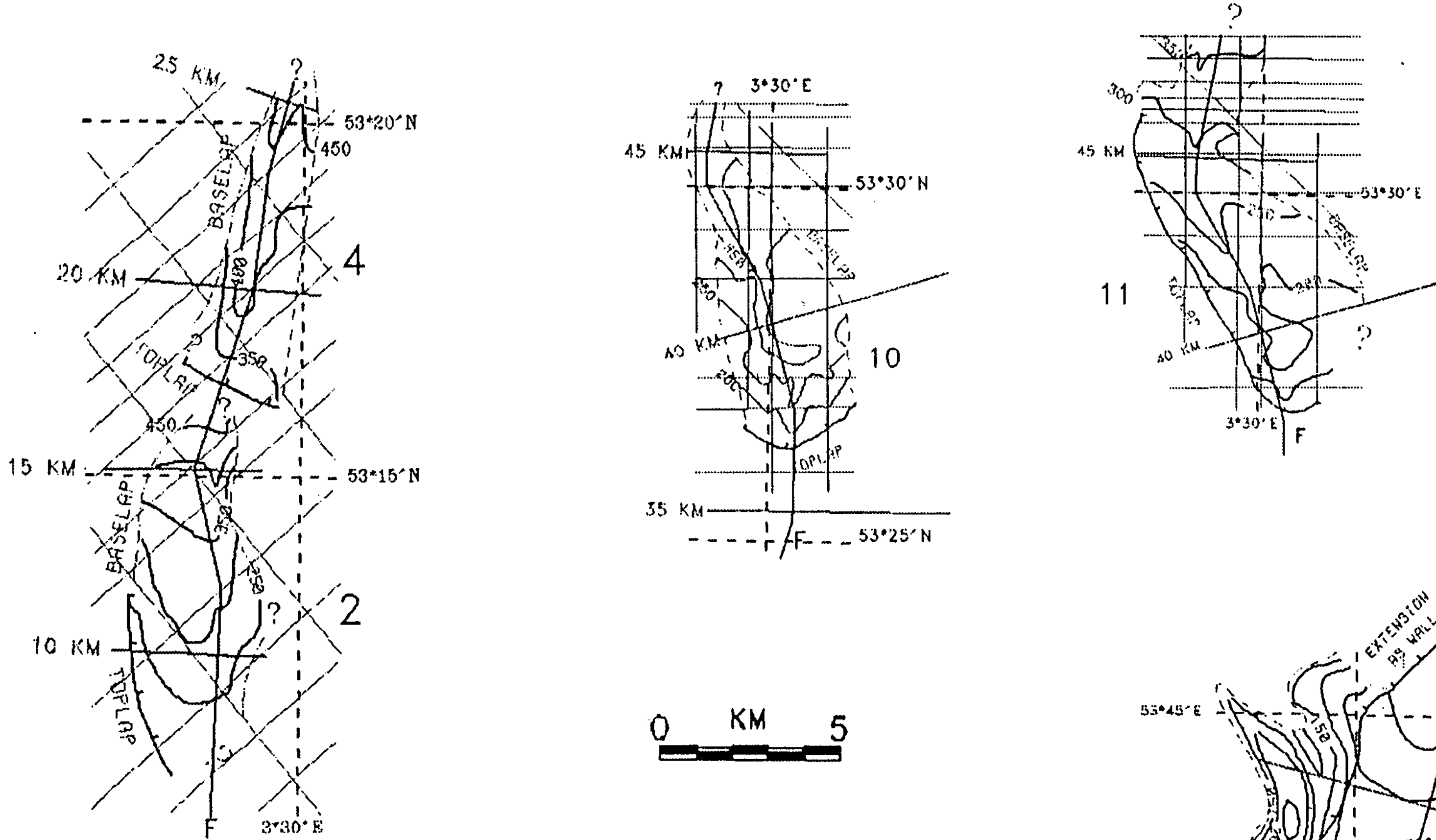
The surfaces are concave in cross-section, with relief across their widths (up to 5 km) of 20-70 ms, locally to 130 ms (30 km, profile F - Fig. 2.15). Cross-sectional gradients are generally less than 3° but gradients up to 10° occur locally across central depressions up to 50 ms deep (e.g. 20-40 km, profile F - Fig. 2.15). The depressions are associated with irregular relief, in places traceable in plan as depressions and ridges which appear to diverge up-slope (Figs. 2.24, 2.25). Contoured surfaces taper down-slope against the walls of the basins and are difficult to trace to downlap. Where axial downlap is observed directly the surfaces extend smoothly to the valley base, to the limits of seismic resolution (Fig. 2.25a).

The clinoform surfaces decline northwards at overall axial gradients of 0.5-3° and up to 3.5° along portions of some surfaces (e.g. 7 km on profile C1 - Fig. 2.11a; 19-20 km and 37-39 km on profile F - Fig. 2.15). Subhorizontal or reversed gradients also occur along some surfaces, for example on profiles C2 and C3 (Figs. 2.11a,b). Many surfaces are slightly concave but there is no typical axial form - steeper central portions are observed, as well as steeper upper and lower portions. Gradients are generally steeper in the south (1-3°) and reduce northwards at certain locations on profiles A-F (Fig. 2.8). Gradients reduce to less than 1° on northern profiles A (20 km), C2 (33 km) C3-C4 (50 km) and F (70 km) and to 0.5-2° on northern profiles D (25 km) and E (10 km). The northward reductions in gradient take place across discordant surfaces of discontinuity (section 2.5).

The vertical extent of the surfaces varies with the thickness of sequence I, which depends on the relief both of the valley base and of its irregular upper surface. Overall thicknesses increase from west to east in correspondence with the increase in basin depth and relief (Fig. 2.7). West of 3°E, thicknesses are only locally less than 100 ms (90 m) and in places are up to 250-300 ms (225-270 m) (16 km, profile A - Fig. 2.9; 44 km, profile C2 - Fig. 2.11a; 44 km and 56-33 km, profile C4 - Fig. 2.11b). East of 3°E, the valleys of profiles D to F include thicknesses over 300 ms (270 m), locally to 450 ms (400 m) along profile D (10-15 km - Fig. 2.12). In general, clinoform surfaces account for at least half of valley fill thicknesses.

The axial extent of the surfaces varies with both gradient and sequence thickness from 3-20 km. West of 3°E, steeply dipping surfaces in the south have estimated extents of 3-8 km along profiles A to C1 (Figs. 2.9-2.11) while gently dipping surfaces in the north have extents of 5-20 km (profiles C2-C4 - Figs. 2.11). In the eastern half of the area, relatively steeply dipping surfaces in the south are over 10 km in extent along profiles D and F due to greater sequence thickness (Figs. 2.12, 2.15).

Discordance of adjacent clinoform surfaces is observed in places on seismic profiles (e.g. Fig. 2.4E). It is recorded by contoured surfaces in two locations where data quality and line-spacing allowed the tracing of closely-spaced (<1 km) surfaces. At 43 km along profile C2 (Fig. 2.11a) and at 67 km along profile F (Fig. 2.15) lower surfaces converge towards overlying surfaces, in the latter case both up- and down- the clinoform surface slope. Given the separable resolution of exploration data (Table 2.1), the surfaces are within 15 m of each other vertically.



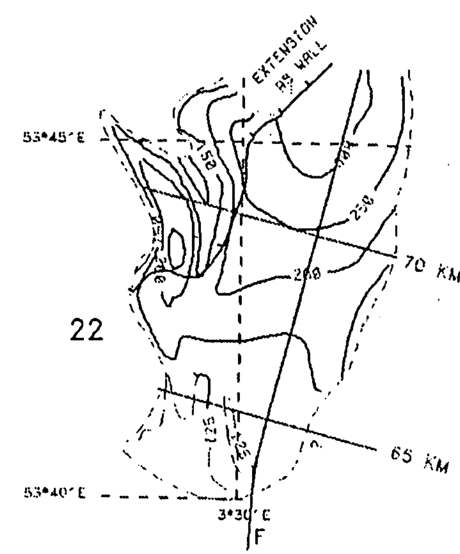


Figure 2.24 - Contours (50 ms intervals) on five clinoform surfaces within sequence I, from exploration data (thin lines) along profile F (thick lines, axial and transverse sections labelled). Surfaces are as numbered on profile F (Fig. 2.15). Bounding relations are indicated as baselap (on- or downlap against the valley base), and toplap (against overlying surfaces, see profile F). Surface 22 extends northwards into a smaller tributary to the main valley, with basinal relief, and merges with the valley wall across the hachured line, a relationship interpreted as erosional (see Fig. 2.15).

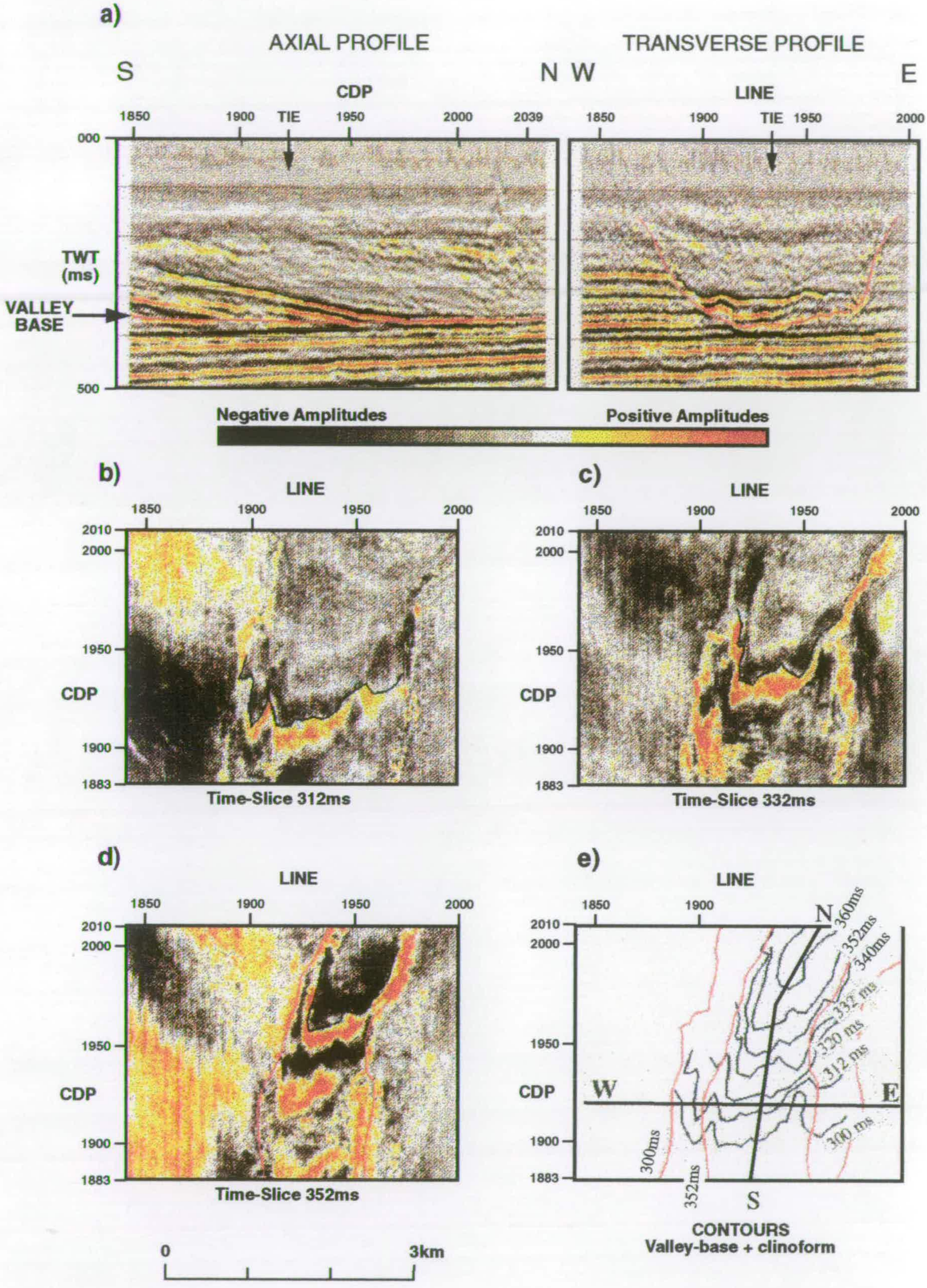


Figure 2.25 - Construction of time-slice contours on a clinoform surface within fill sequence I: **a)** axial and transverse profiles (locations at lower right) with superimposed interpretation of valley base (red) and prominent clinoform surface (blue) ; **b-d)** time-slices at 20 ms intervals with superimposed contours drawn on positive to negative amplitude contrast across the clinoform surface; **e)** successive time-slice contours on the clinoform surface (blue, 300-360 ms) and on the valley base (red, 300-352 ms). A decimal contour interval is precluded by the 4ms digital sample rate. Location on Fig. 2.16.

2.4.2 Boundary I/II

The boundary between sequences I and II represents the angular axial discordance between clinoform and subhorizontal surfaces (Fig. 2.23). It was not everywhere traceable as a reflecting surface on exploration data, although in places it is found to be coincident with contoured surfaces (Figs. 2.9-2.12, 2.15). Rather, it was inferred along profiles A to F from discordant relations, either between contoured surfaces or between reflectors on seismic profiles (where overlying surfaces were not contoured - recorded by 'T" on profiles B to F). It was also inferred, rather than traced directly, along axial 3D-seismic profiles D and E (Figs. 2.13, 2.14). Along profiles B-D, it was also interpolated from intersections with BGS profiles (of reasonable signal to noise content), on which the boundary may be distinct as a concave reflector of irregular small-scale (<10 m) relief onlapped by subhorizontal reflectors (Fig. 2.22).

The boundary is of irregular large-scale axial relief, defining basins separated by sills at the surface of sequence I. The basins are generally less than 100 ms (90 m) in relief and less than 15 km long (profiles A, C1-C4, E - Figs. 2.9, 2.11, 2.15), although a basin (with sub-basins) up to 130 ms (120 m) in relief and 30 km long is recognised on southern profile F.

The axial relief is correlative between some adjacent valleys. Along the three valleys of profiles C2-C4 (Fig. 2.11), a basin and a sill to its north occur in similar locations between about 45-60 km and are correlative along lines transverse to the orientations of the subparallel valleys (see Fig. 2.27). A basin also appears to be correlative between profiles D to F (Figs. 2.13-2.15 - 32 km on D, 11 km on E, 20 km on F; see Fig. 2.27).

2.4.3 Sequence II (Onlapping Surfaces)

Sequence II comprises subhorizontal reflecting surfaces that onlap the axial basins and the upper valley walls with thicknesses up to 150 m. One or two onlapping surfaces were traced along portions of profiles B to F (Figs. 2.10-2.15) from both exploration data (where thicker) and from BGS profiles in the UK sector. In some cases, the presence of the sequence was inferred from its discordant lower boundary but no surfaces were contoured (e.g. profiles A, F - Figs. 2.9, 2.15). An example of a contoured surface is given in Fig. 2.26.

Sequence II corresponds to facies unit 2 on BGS profiles (Fig. 2.22), distinct from underlying unit 1 by its "well defined, parallel to subparallel reflectors, draped over any underlying irregularities" (Balson and Jeffery, 1991, p. 247) and including "rare, high amplitude reflectors which occur at similar intervals and depths in adjacent valleys" (Cameron et al., 1987, p. 52). In contrast, on exploration profiles sequence II is of similar 'facies' character to underlying sequence I, distinguished only by discordant axial reflector relations (e.g. profile E - Fig. 2.14). Prominent reflectors could be correlated between the two types of data for contouring and are confirmed to extend throughout some adjacent valleys (e.g. Fig. 2.26).

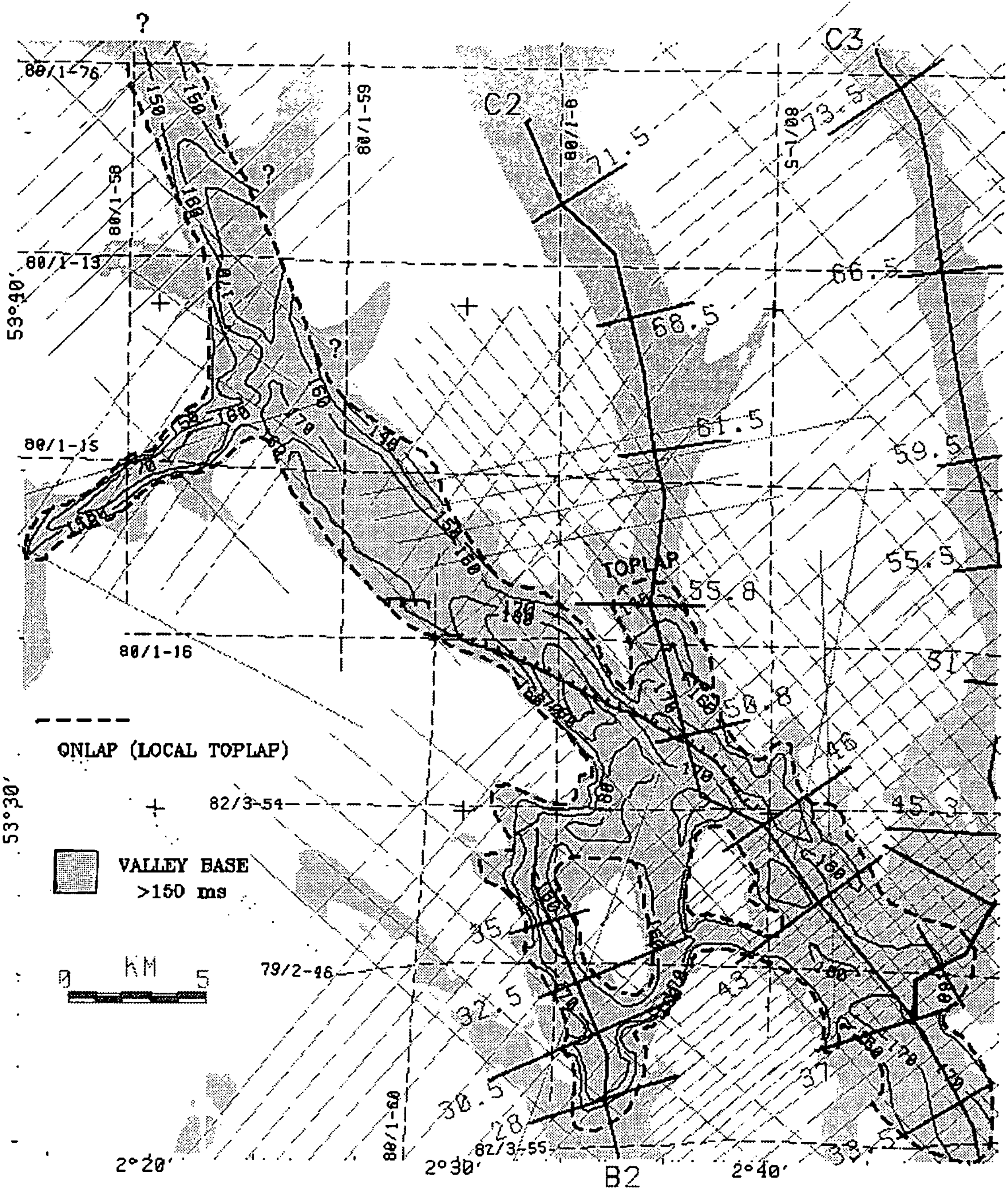


Figure 2.26 - Contours (140-190 ms, 10 ms intervals) on a subhorizontal surface within sequence II, drawn from indicated exploration (solid) and BGS (dashed, labelled) profiles. The surface is cut by faults (hachured lines). The bold dashed line represents the onlap of the surface against the upper valley walls, and axially against underlying sequence I to the south; to the north along profile C2, the surface axially toplaps against overlying sequence III (upper surface on profiles B and C - Figs 2.10, 2.11a). Note axial sub-basins 10-30 ms in relief.

Contoured surfaces extend over distances of over 20 km on profiles B, C and F and up to 40 km along profile C2 (Fig. 2.8). In cross-section, the surfaces are gently concave (<1°) with relief generally less than 30 ms (27 m) across their widths (up to 5 km), although gradients up to 2° and relief up to 120 ms (110 m) are recognised on southern profile F (20 km, Fig. 2.15). On axial profiles, the surfaces include broad, shallow concavities (gradients less than 0.5°) up to 50 ms (45 m) in relief and 15 km long (southern profile F), which coincide with underlying axial basins. Concavities of 10-30 m relief are shown by contours on Fig. 2.26. Smaller irregularities also occur (e.g. profile C2 - Fig. 2.11a) and are inferred to be conformable to relief on the underlying surface of sequence I. A downlapping surface is recognised in one location, declining to the south (at 0.7°) over about 2 km (profile C2, 55 km).

The sequence is mostly less than 100 m thick but dominates the fill in places (e.g. northern profile B). Thicknesses greater than 100 m occur locally in axial basins: up to 170 ms (150 m) on southern profile F (20 km, Fig. 2.15), up to 140 ms (125 m) at 23 km on profile C1 and 53.5 km on profile C4 and up to 120 ms (110 m) at 49 km on profile C2 (all Fig. 2.11). The majority of the sequence is discontinuous within the basins but an upper part extends over axial sills in thicknesses as low as 20 m (e.g. profile C3, 45 km and C4, 60 km - both Fig. 2.11).

2.4.4 Boundary II/III

The boundary between sequences II and III is recognised as a prominent reflector on BGS profiles, distinguished by the discordant downlap of overlying reflectors (e.g. Fig. 2.22). This allowed interpolation between widely-spaced BGS profiles in the north-central UK sector (profiles B, C2-C4 - Figs. 2.10, 2.11) and locally in the southeast (profile D - Fig. 2.12). Discordance with underlying surfaces of sequence II is noted on BGS profiles (Cameron et al., 1987; Balson and Jeffery, 1991) and is inferred locally from contoured surfaces (56 km, profile C2 - Fig. 2.11).

In the north-central UK sector, the concave boundary onlaps the upper valley walls in cross-section (profiles C2-C4 - Fig. 2.11) and defines basins on axial profiles. On profile C2, a single basin is recognised with depths up to 170 ms (150 m bsl) at 71.5 km, corresponding to axial relief of at least 80 ms (70 m) between 50-80 km. The northern face of the basin ascends steeply from 71.5-74.5 km (axial gradient 1.3° to the south). On profile C3, two basins are separated by a sill at 58 km (120 ms), the southern with depths up to 140 ms (at 52 km), the northern at least 170 ms deep. On profile C4, a single basin occurs with depths up to 125 ms (76 km - Fig. 2.11).

The boundary rises to the south and is discordant against the valley-top unconformity (e.g. 53 km and cross-section 57.5 km, profile C2 - Fig. 2.11). It is recognised along 12 km of profile D, also rising to the south from 130 ms (125 m bsl) to 100 ms (90 m bsl) towards the valley-top unconformity (Fig. 2.12).

2.4.5 Sequence III (Complex Surfaces)

Sequence III is defined from BGS profiles and corresponds to facies unit 3 (Fig. 2.22) which comprises inclined reflectors of variable configurations (Cameron et al., 1987; Balson and Jeffery, 1991). It was not possible to contour reflecting surfaces within the sequence, due to their limited extent relative to BGS line spacing. The sequence is characterised by reflector configurations which include discordance with the lower and upper boundaries, as well as with each other.

The sequence discordantly overlies the axial basins of its lower boundary in the north-central UK sector (profiles B, C2-C4) and locally in the southeast (profile D). On profiles C2-C4 (Fig. 2.11), it thickens within the basins to a maximum of 110 ms (100 m) on profile C2 (71.5 km), to at least 100 ms (90 m) on profile C3 (80 km) and to a maximum of 60 ms (55 m) on profile C4 (76 km). It thins to the south and appears to pinch out against the overlying valley-top unconformity (e.g. 53 km along profile C2 and on the east side of the cross-section at 57.5 km - Fig. 2.11).

On cross-sections, the sequence thins but overtops the valley shoulders in the north (e.g. Fig. 2.22; see also Sections 2 and 3, BGS/RGD, 1986). Between about 60-80 km along profiles C2-C4, it extends between the three adjacent valleys in thicknesses less than 30 ms (25 m) beneath the valley-top unconformity (Fig. 2.11).

2.4.6 Interpreted Deposition and Erosion

The sequences and their boundaries are interpreted in term of deposition and erosion by assuming the reflecting surfaces to approximate strata, to resolution limits (Fig. 2.23). The lower fill (sequence I) records progradation to the north, beneath an upper boundary of irregular axial relief inferred to be of depositional origin. The upper fill (sequences II and III) records two different styles of deposition, separated by an erosional boundary.

2.4.6 a) lower fill

Sequence I records northward progradation within the tunnel-valley basins along concave clinoform surfaces (up to 3.5° in gradient, 3-20 km long, 1-6 km wide, up to 400 m high). Adjacent surfaces tend to parallelism and are stacked upon each other without apparent interruption on seismic profiles (e.g. Figs. 2.4, 2.14) suggesting continuous accumulation. 'Adjacent' surfaces, at vertical separable resolution as low as 15 m and gradients of 0.5-3°, may be separated by horizontal distances of 300-1700 m. Upward discordance between such surfaces (e.g. Figs. 2.4E, 2.15) is likely to record convergence to within resolution limits (see Fig. 2.6a).

Progradation resulted in the partial fill of the basins, generally over half of their relief. The discordant upper boundary defines a series of axial basins up to 120 m in relief. The boundary and basins are inferred to be of depositional origin, the result of variations in sequence I thickness during its northward accumulation. A depositional origin is supported by the correlation of axial basins between adjacent tunnel-valleys along lines transverse to the valley axes and so transverse to the direction of underlying progradation (see Fig. 2.27).

The upward termination of the clinoform surfaces against the sequence boundary is thus inferred to record depositional convergence below seismic resolution limits, rather than erosional truncation (*cf.* Fig. 2.6a). This could help to explain why the boundary is poorly defined on exploration profiles - at separable resolution as low as 15 m and axial gradients as low as 0.5° 'terminations' could be distributed over horizontal distances of 1700 m, or more if the convergence is asymptotic. In contrast, on BGS profiles the boundary is generally distinct due to its small-scale (<10 ms) irregular morphology (e.g. Fig. 2.22).

The axial basins are filled by subhorizontal surfaces of sequence II. As the axial basins formed during the northward progradation of sequence I, it is logical to infer that the onlapping lower parts of sequence II were deposited in successive basins and so are also younger to the north. If so, this implies that the tunnel-valleys were substantially filled (sequences I + II) during the course of deposition. This will be considered further in Chapter 4.

2.4.6 b) upper fill

Sequence II records subhorizontal accumulation up to 150 m thick within the upper parts of the tunnel-valleys. The inferred continuity of prominent reflectors at similar depths in adjacent valleys has been suggested to indicate deposition in "an interconnected network of open depressions" (Cameron et al., 1987, p. 52). However, only the upper part of the sequence is continuous throughout adjacent valleys, the majority being axially discontinuous within basins at the surface of sequence I. The draped character, concordant with small-scale basal irregularities, suggests low energy sedimentation (Balson and Jeffery, 1991).

The discordant upper boundary of sequence II is interpreted to record incision of underlying surfaces within a series of axial basins up to 70 m in relief. Incision is inferred both from reflector discordance along BGS profiles (Cameron et al., 1987; Balson and Jeffery, 1991) and against axial concavities of locally steep gradient (profile C2, Fig. 2.11). Sequence III records complex deposition within the basins and above the shoulders between adjacent tunnel-valleys in the northern UK sector. Reflectors are in places subhorizontal but also include configurations indicative of progradation and of cut-and-fill structures, both suggesting high energy environments (Balson and Jeffery, 1991).

Sequences II and III are both erosionally truncated to the south against the subhorizontal valley-top unconformity in the UK sector.

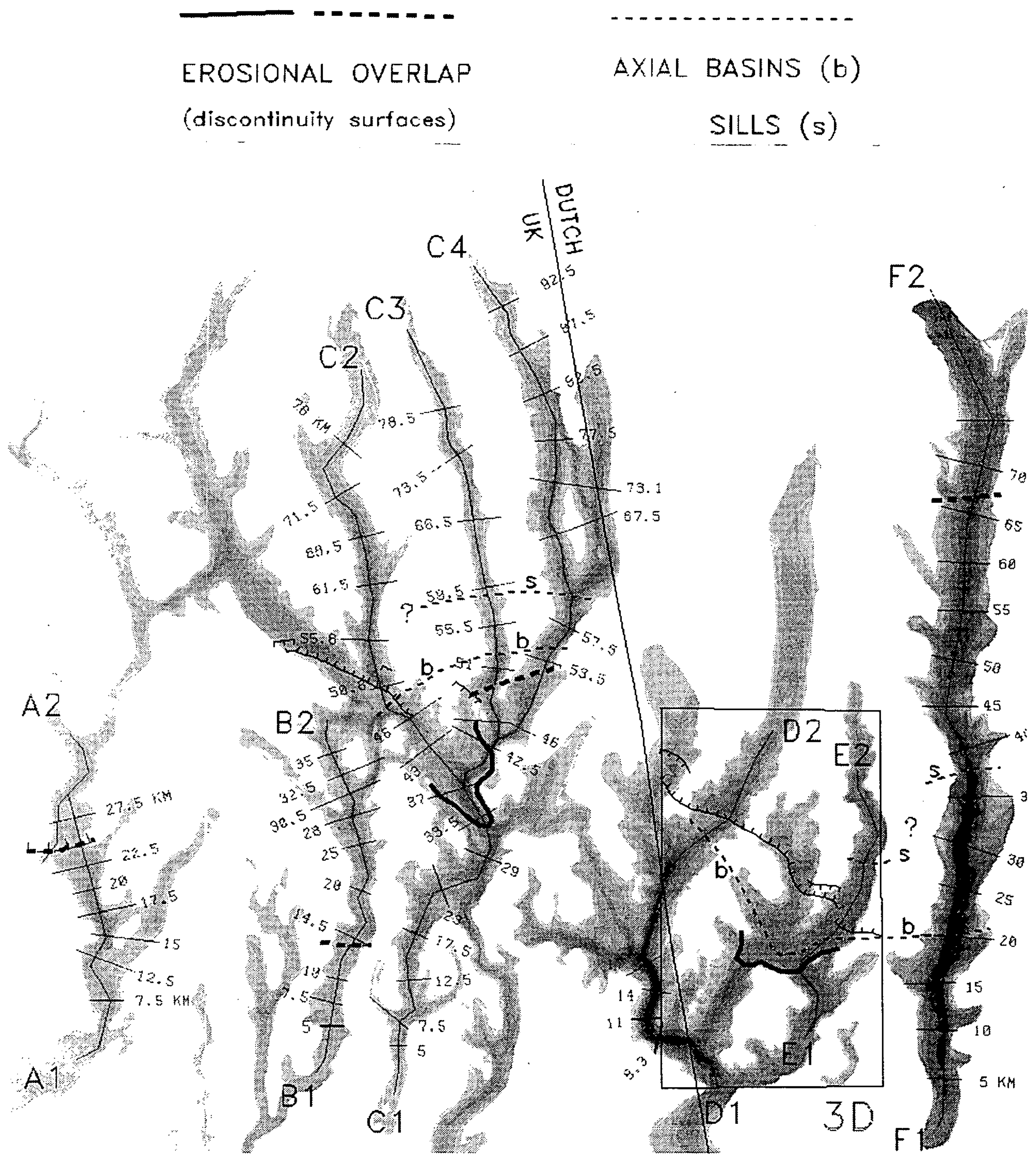


Figure 2.27 - Locations of northwards erosional overlap across surfaces of discontinuity (observed, inferred) and lines of correlation of axial basins and sills at the surface of sequence I. Profiles A to F correspond to Figs. 2.9-2.15.

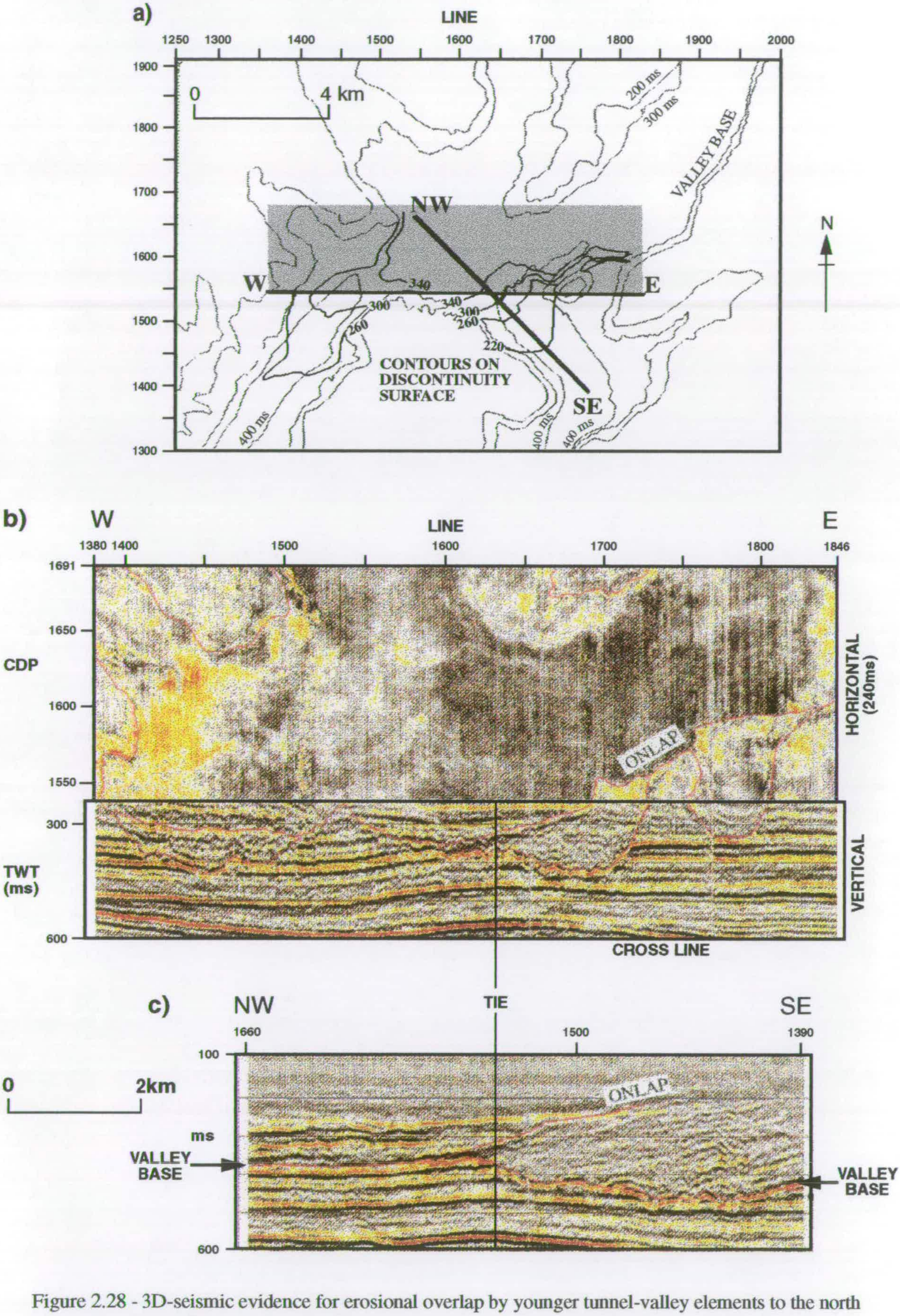


Figure 2.28 - 3D-seismic evidence for erosional overlap by younger tunnel-valley elements to the north across a surface of discontinuity, resulting in an apparent anastomosing plan form: a) time-slice contours drawn on the valley base (100 ms interval, lighter) and on the discontinuity surface (20 ms interval, bold); b) time-slice + vertical combination showing the surface in profile (note onlap, underlying discordance) and its horizontal extension (at 240 ms) as a contrast in fill character; c) vertical profile along the indicated NW-SE line, showing the onlap surface and its northward extension as the valley-base (note contrast in fill character south to north).

2.5 Erosional Overlap (Basal/Fill Discontinuities)

The valley base and the lower fill (sequence I) are interrelated via surfaces of discontinuity that record erosional overlap to the north. The discontinuities correspond to complex clinoform surfaces within sequence I, discordant with underlying surfaces and axially onlapped by younger (less steep) surfaces to the north. They decline to join the valley base to the north and coincide with plan form complexity (anastomosing patterns, angular offsets). The spatial associations of basal and fill characteristics across the study area indicate erosion of older cut-and-fill elements to the south and overlap by successively younger elements to the north (Fig. 2.27). I use the term erosional overlap to describe a form of cross-cutting in which erosion (and deposition) are progressively displaced to the north along similar lines.

Two surfaces of discontinuity are unambiguously recognised and shown to have resulted in 'apparent' anastomosing patterns. A number of other surfaces are inferred to be present from correspondence between fill discordances and angular offsets in valley orientation. I discuss the evidence for discontinuity observed in association with apparent anastomosing (2.5.1) and inferred in association with angular offsets (2.5.2).

2.5.1 Observed (Apparent Anastomosing)

The southwards convergence of the tunnel-valley basins is interrupted in places by anastomosing patterns, i.e. divergence around isolated elevations or 'islands' (section 2.3.2). Two main locations coincide with discordant surfaces within sequence I that record erosional overlap to the north (Fig. 2.27), within the 3D-seismic volume (profile E) and in the central UK sector (profiles C2-C4).

In the area of 3D-data, anastomosing occurs around two 'islands' (one large and a second to the northeast much smaller) which lie to the south of two main convergent branches (Figs. 2.16, 2.27). Vertical and horizontal seismic sections indicate a prominent fill surface coincident with the northern faces of the two islands (Fig. 2.28). The surface is discordant with underlying fill reflectors to the south and onlapped by fill surfaces to the north. Contours show that the surface declines northwards in two adjacent valleys from 220-340 ms and merges with the valley base to the north (NW-SE section, Fig. 2.28; see also profile E, 6-12 km - Fig. 2.14). The spatial relationships indicate erosional truncation of valleys to the south and overlap by a younger basal element to the north resulting in an 'apparent' anastomosing pattern (Fig. 2.28).

In the central UK sector, anastomosing occurs around several 'islands' along the southwest side of profile C2 and locally northeast between C2 and C3 (30-50 km, Fig. 2.27). A prominent discontinuity surface within sequence I indicates northward truncation by the valley of profile C2. The surface is recognised as a broad concavity eroded into underlying fill surfaces from 30-43 km along the southern portion of profiles C3/4 (Fig. 2.11b). Along profile C2, the same surface descends to the northwest and merges with the valley base (30-38.5 km, Fig. 2.11a). On cross-sections of profile C2 at 33.5 km and 37 km (Fig. 2.11a), the concavity is up to 3 km wide, comparable to the width of valleys in the area. The evidence is of incision of the (shallower) valley of profile C2 into (deeper) tunnel-valleys in its southern part. The anastomosing patterns along its southwestern and northeastern flanks are not directly traceable to the discontinuity surface, but are consistent with truncation of older valleys (Fig. 2.27).

Anastomosing also occurs along the valley of profile C4 from 60-75 km (Fig. 2.27). A fill discontinuity was not traceable as such, but basal and fill surfaces on three cross-sections (73.1 km, 77.5 km and 82.5 km - Fig. 2.11b) are interpretable in terms of westward truncation of a smaller valley by the valley of profile C4. The small valley may also be truncated to the east (Fig. 2.11b). The reflector relationships were drawn from data of poor quality and so are speculative, but are consistent with the plan form discordance (Fig. 2.27).

Plan form complexity can thus be related to erosional overlap by younger basal elements to the north across surfaces of discontinuity. Discontinuity surfaces are inferred elsewhere from discordant relations within sequence I which coincide with changes in plan form along individual valleys.

2.5.2 Inferred (Angular Offsets)

The discontinuity surfaces described above correspond to complex axial clinoform surfaces within sequence I, discordant with both under- and overlying surfaces (e.g. Figs. 2.11b, 2.14, 2.27). The lower discordance records erosional truncation, while the upper records axial onlap by surfaces of reduced gradient to the north. Discordant relations are observed elsewhere within sequence I which may also record erosional overlap across discontinuity surfaces. They are coincident with changes in basal morphology, notably angular offsets either of valley segments or of convergent branches.

Along profile A, an *en echelon* doubling of the valley axis is recognised north of about 25 km, the northwestern axis oriented at an angle to that in the south (Figs. 2.8, 2.9). Clinoform gradients reduce from 1-2° in the south to less than 1° north of about 20 km (Fig. 2.9). On a cross-section at 20 km, surface 6 is discordant against overlying surface 7 which defines a concavity with its axis to the west coincident with the northwestern of the two axes on the cross-section at 27.5 km (Figs. 2.8, 2.9). The fill relationships are not clear, but may record discordant overlap by the valley segment to the north (Fig. 2.27).

Along profile B, a northward thinning of sequence I (and thickening of sequence II) across an onlapped surface at 14 km (Fig. 2.10) corresponds with a local doubling of the axis and a deepening and an (angular) change in orientation to the north (Fig. 2.8). The axial fill relationship is comparable to that observed across the discontinuity along profile E (6-11 km, Fig. 2.14) and may similarly record a younger basal element to the north (Fig. 2.27).

Along profiles C3 and C4 (Fig. 2.11b), clinoform surface gradients reduce to the north of about 50 km from up to 3° to less than 1° in both valleys. An onlap relationship is not recognised from the (widely spaced) contoured fill surfaces. However, the valleys of both profiles shift orientation north of about 50 km to converge at an angle with the main axis to the south (Fig. 2.8). Both valleys are subparallel with that of profile C2, which is eroded into the same steeply inclined clinoform surfaces in the south across a prominent surface of discontinuity which separates gently inclined clinoform surfaces to the north, also of less than 1° gradient (Fig. 2.11a). The valleys of profiles C2 to C4 are thus all of similar orientation, with similar clinoform gradients. Relief at the surface of sequence I is also correlative between the three (Fig. 2.27). I infer that, as with profile C2, the valleys of profiles C3 and C4 erosionally overlap the steeper clinoform surfaces to the south across a surface of discontinuity.

Finally, along profile F (Fig. 2.15) gradients reduce north of 70 km across a prominent fill surface, which extends into a tributary valley to the northwest that appears to truncate the valley wall (cross-section 70 km, Fig. 2.15; also contoured surface 22, Fig. 2.24). The main valley axis shifts orientation to the northwest north of this surface, subparallel with the tributary (Fig. 2.8). The relationships are consistent with erosional overlap to the north (Fig. 2.27).

2.6 Interpreted Morpho-Stratigraphy

Seismic morphology and stratigraphy together provide evidence of basal erosion to the south, fill progradation to the north and erosional overlap to the north. These observations are reconciled by drainage south to an ice margin receding north, which resulted in contemporaneous erosion and deposition (backfill). This is comparable to previous interpretations of tunnel-valleys formed during deglacial recession, although stratigraphical evidence of backfill has not previously been recognised. Drainage and backfill are considered in Chapter 4 in terms of subglacial processes, here I am simply concerned with their general implications for erosion and deposition.

I discuss the evidence for ice margin recession (2.6.1) and compare it to previous work (2.6.2).

2.6.1 Ice Margin Recession

Northwards displacement of erosion (directed to the south) during the northwards progradation of the fill is compatible with recession of the Elsterian ice sheet margin northwards across the area. Basal and fill characteristics suggest changes in ice margin orientation and local readvances.

I discuss a) drainage erosion and b) backfill.

2.6.1 a) drainage erosion

The valley base records the excavation of broad, relatively shallow and steep-sided basins within the Cenozoic sediments. The convergent plan form is indicative of erosion by drainage directed to the south, by analogy with other arborescent systems. However, surfaces of discontinuity record erosional overlap by younger basal elements to the north (Figs. 2.27, 2.28). Thus drainage to the south was accompanied by the progressive northward migration of a limiting boundary of erosion. This corresponds to the migration of the ice margin across the study area.

Erosional overlap resulted in both the anastomosing patterns observed in places and the angular offsets of successive valley segments (see Fig. 2.27). The former record erosion of older cut and fill elements and imply readvances of the ice margin across previously occupied areas. Angular offsets record changes in the direction of erosion and suggest changes in the orientation of the ice margin.

Drainage to a linear boundary of changing orientation also helps to account for the subparallel character of the tunnel-valleys and their changing orientations across the study area (e.g. Fig. 2.27).

An origin of the rectilinear basal morphology in erosional processes determined by the ice margin is also consistent with evidence of its superimposition on structural trends in the substrate, which have only locally influenced plan form (Fig. 2.19). However, the eastward increase of valley size and spacing in parallel with substrate thickness (Fig. 2.21) suggests a regional influence on drainage processes, considered in Chapter 4.

2.6.1 b) backfill

Fill progradation along clinoform surfaces of sequence I resulted in the partial fill of the tunnel-valleys beneath a surface of depositional relief (e.g. Fig. 2.23). The prograding surfaces and the basins at their surface (and by extension the onlapping surfaces that fill the basins) record the northward migration of sites of deposition in adjacent tunnel-valley basins. I infer that they record the backfill of the basins in response to sediment supplied to the ice margin by drainage erosion.

Progradation was interrupted by erosional overlap, recorded by discordant clinoform surfaces onlapped by younger prograding surfaces of reduced gradient (e.g. Fig. 2.28). The discordant surfaces separate older and younger cut and fill episodes to the south and north, respectively. They demonstrate that erosion and progradation were contemporaneous during the recession of the ice margin and that the prograding surfaces were themselves eroded locally, by readvances of the margin.

Backfill during the northward displacement of erosion helps to account for the spatial correspondence of fill and basal characteristics. Comparable clinoform surface gradients in some adjacent tunnel-valleys of similar orientation (e.g. Fig. 2.11) reflect their contemporaneous deposition in features being eroded along parallel lines. Correlation of axial basins and sills at the surface of sequence I along lines transverse to tunnel-valley orientations (Fig. 2.27) is suggestive of deposition along successive ice margin lines during recession.

2.6.2 Comparison with Previous Work

Erosion by drainage to receding ice margins is an interpretation comparable to those of other tunnel-valley networks, mainly landforms of the last deglaciation. Backfill is a process that has been inferred but for which stratigraphical evidence has not been previously presented.

I discuss a) drainage erosion and b) backfill.

2.6.2 a) drainage erosion

Arborescent morphologies and plan form complexity due to erosional overlap during ice margin recession both have been recognised for tunnel-valley networks in Europe and North America.

Tunnel-valleys of the southern Baltic lowlands are characterised by convergence to former ice margin lines, both the last glacial maximum and younger margins (e.g. Fig. 1.4; also Andersen, 1931; Galon, 1965; Kozarski, 1966/67; Ehlers and Wingfield, 1991). Complex plan forms have resulted from the overlap of younger features due to changes in ice margin orientation or position, such as across the up to 350 km breadth of the last deglaciation of Poland (Galon, 1965, 1983). The formation of tunnel-valley networks during recession has also been demonstrated across 150 km of the last deglaciation in Minnesota, where overlap of successive segments 10-20 km long resulted in an 'apparent anastomosing pattern' (Mooers, 1989a). The segments contributing to this pattern are for the most part singular (unbranched) valleys, although arborescent patterns are recognised to the east along the last glaciation maximum line in Wisconsin (Attig et al., 1989).

Plan form complexity has also been inferred to result from erosional overlap, in the absence of information on ice margin lines. Wingfield (1990) argued that an *en echelon* series of buried tunnel-valleys in the central North Sea resulted from erosion to a receding ice margin position and suggested that the complex plan forms of other tunnel-valleys in the North Sea and beneath Hamburg were indicative of cross-cutting relationships. Loncarevic et al. (1992) also considered that a complex system of bathymetric and buried tunnel-valleys on the inner Nova Scotian shelf resulted from multiple erosion events under different ice configurations.

In contrast, the complex or anastomosing character of tunnel-valley networks has been regarded as *a priori* evidence of their synchronous subglacial formation. An interconnected network of buried tunnel-valley networks on the outer Scotian Shelf was attributed to erosion by floodwaters (Boyd et al., 1988). Complex surficial networks in Canada have been suggested to be explicable 'only' by occupation by floodwaters, although the same networks contain arborescent systems of esker ridges (e.g. Shaw and Gorrell, 1991; Brennand and Shaw, 1994). These studies make no reference to the location of the ice margin.

In Germany, net-like interconnections (see Fig. 1.5) have provided an argument for the non-fluvial origin of buried Elsterian tunnel-valleys, but not for the means of their formation (Kuster and Meyer, 1979; Ehlers, 1981; Ehlers et al., 1984). Erosion and deposition are recognised to take place during deglaciation and ice margin locations have been inferred from fill characteristics (Ehlers and Linke, 1989). However, it has not been possible to relate the complex plan form to ice margin lines as suggested by Wingfield (1990).

The rectilinear character of tunnel-valleys in the Baltic lowlands has attracted speculation as to possible structural controls on erosional processes. Buried tunnel-valley distributions have been suggested to be controlled by underlying halokinetic structural trends, although the correspondence of the two is generally poor (see Ehlers et al., 1984). Kronborg et al. (1977) proposed that the morphology of several adjacent surficial tunnel-valleys in central Jutland, which comprise correlative straight segments offset at angles, resulted from neotectonic displacements along (hypothetical) faults. In contrast, Andersen (1931) had proposed similar angularly segmented tunnel-valleys in southeastern Zeeland to be the result of erosion (and deposition) of adjacent features along parallel but changing lines during deglaciation. This is comparable to the southern North Sea, where I infer the erosional morphology to have been controlled by the receding ice margin and superimposed on structural trends.

2.6.2 b) backfill

Contemporaneous erosion and deposition was implicit in the original association of tunnel-valleys with outwash fans (Ussing, 1903). A depositional origin of the irregular axial relief of tunnel-valleys by outwash from the receding ice margin was proposed by Werth (1912) and in application to eskers within tunnel-valleys by Andersen (1931). These proposals effectively recognised the glaciofluvial backfilling of the tunnel-valleys during recession (section 1.2.1c). Mooers (1989a) also demonstrated the contemporaneous formation of tunnel-valley and contained esker segments during recession and considered that the two features might have formed together.

Woodland (1970) extended the concepts of Ussing and Werth to buried tunnel-valleys in East Anglia and argued that the sandy lower fill recorded deposition from subglacial streams during deglacial recession, i.e. backfill. A comparable interpretation has been applied to the sandy glaciofluvial fills of Elsterian tunnel-valleys in Germany (Grube, 1983; Ehlers, 1981; Ehlers and Linke, 1989) and in western Canada (Christiansen, 1987; McClung and Mollard, 1987). These ideas have been extended to the 'penecontemporaneous' deposition of lower fill sediments during incision of tunnel-valleys in the North Sea (Dingle, 1970; Caston, 1977; BGS/RGD, 1986; Cameron et al., 1987; Balson and Jeffery, 1991). Wingfield (1990) proposed that tunnel-valleys in the North Sea were contemporaneously cut and filled along ice margins during successive flooding events, but his schematic interpretation of backfill (his Figure 5) indicated surfaces dipping in the same direction as water drainage, i.e. away from the former ice margin.

2.7 SUMMARY

Grids of seismic reflection data yield contoured information on the valley base and on reflecting surfaces within the fill, morpho-stratigraphically interpretable in terms of erosion and deposition.

The valley base records the excavation of a series of broad (0.5-6 km), relatively shallow (maximum 500 m relief) and steep-sided (5-40°) elongate basins within the Cenozoic sediments. A subparallel arborescent plan form records drainage erosion directed to the south, while erosional overlap by younger basal elements to the north has resulted in apparent anastomosing patterns and angular offsets of segments up to 20 km long. The rectilinear pattern is superimposed on stratal and structural trends in the incised strata, which have only locally influenced cross-sections (benches) and plan form (across faults). However, a regional influence on erosional processes is suggested by an eastward increase in valley size and spacing in correspondence with substrate thickness trends.

The valley fill is divisible into three seismic sequences on the basis of discordant reflector relations, each bounded by basinal relief. Sequence I overlies the valley base with thicknesses up to 400 m and records northwards progradation along clinoform surfaces 0.5-3.5° in gradient and 3-20 km in extent. Gradients reduce northwards across discordant surfaces separating younger cut and fill elements to the north. The discordant upper boundary of sequence I defines a series of axial basins up to 120 m in relief of depositional origin, correlative between some adjacent basins. Subhorizontal surfaces of sequence II up to 150 m thick record deposition within the axial basins and throughout the upper parts of adjacent tunnel-valleys. Their discordant upper boundary defines a series of incised axial basins, filled by complex surfaces of sequence III which thicken to the north and record deposition between some adjacent tunnel-valleys in the northern UK sector.

Evidence for basal erosion to the south, fill progradation to the north and erosional overlap by younger elements to the north are compatible with drainage to receding ice margins. The results represent the first evidence for fill progradation to the north during tunnel-valley incision, prior to subhorizontal deposition in axial basins. In the next chapter, I present evidence for the composition of the fill and its interpretation in terms of glaciofluvial to lacustrine environments. I return to the question of time in the formation of the tunnel-valleys in Chapter 4.

CHAPTER 3
DOWNHOLE CHARACTERISATION

3.1 INTRODUCTION

The purposes of this chapter are:

- to present evidence of the sedimentological character of the valley fill and relate it to the seismic stratigraphy of the last chapter and the stratigraphical framework of the southern North Sea
- to interpret the results as a vertical succession of glaciofluvial sands (prograding backfill to onlapping outwash), lacustrine muds and marine deposits

Published information on the fill of the Elsterian tunnel-valleys in the southern North Sea is mainly limited to shallow boreholes (<100 m) from the Dutch sector, which proved lacustrine muds beneath marine sands at the top of some valley fills (Oele, 1971b; Laban et al., 1984; Laban, 1995). Several boreholes from the UK sector, including one from the study area (BGS-79/8), have been suggested to contain up to 30 m of glacigenic mud from the edges of buried valleys (Balson and Jeffery, 1991). The majority of the up to 400 m thick fill is thus of unknown composition. It has been proposed that three seismic facies units correspond to lower glaciofluvial sands or diamictons, overlying lacustrine muds and upper marine sediments (Balson and Cameron, 1985; Cameron et al., 1987, 1989b; Long et al., 1988; Balson and Jeffery, 1991), although it has also been suggested that glaciofluvial sands do not occur (Jeffery, 1990, 1991; Cameron et al., 1992).

I present downhole sedimentological and micropalaeontological information from 13 sites in the UK sector of the study area (Fig. 3.1, Table 3.1). Twelve are the shallow portions of exploration wells for which the data consist of continuous geophysical logs (nuclear, electric, acoustic) and samples of debris returned to surface during drilling. Microfossil analyses were available in three cases and I undertook them in five others (Table 3.1). The other site is borehole BGS-79/8 from which short cores provide *in situ* samples of at least 55 m of valley fill muds and sands. Unpublished analyses of foraminiferal, palynological and dinoflagellate content were available from core subsamples. In total I describe over 2 km of section (Table 3.1).

I distinguish a lower sand-dominated lithofacies assemblage that represents the majority of the fill from an upper mud-dominated 'cap'. A marine foraminiferal zone occurs within the upper muds. Sands and muds both include interbedding and fining- or coarsening-up intervals and appear waterlain in core samples. Foraminifera and pollen record reworking from sediments as young as early Pleistocene, which also represent a source for secondary lithological components (lignite, shell fragments, glauconite) abundant in intervals of some wells. The results indicate glaciofluvial sands overlain by lacustrine muds beneath Holsteinian marine sediments, comparable to other Elsterian tunnel-valleys from East Anglia to Germany.

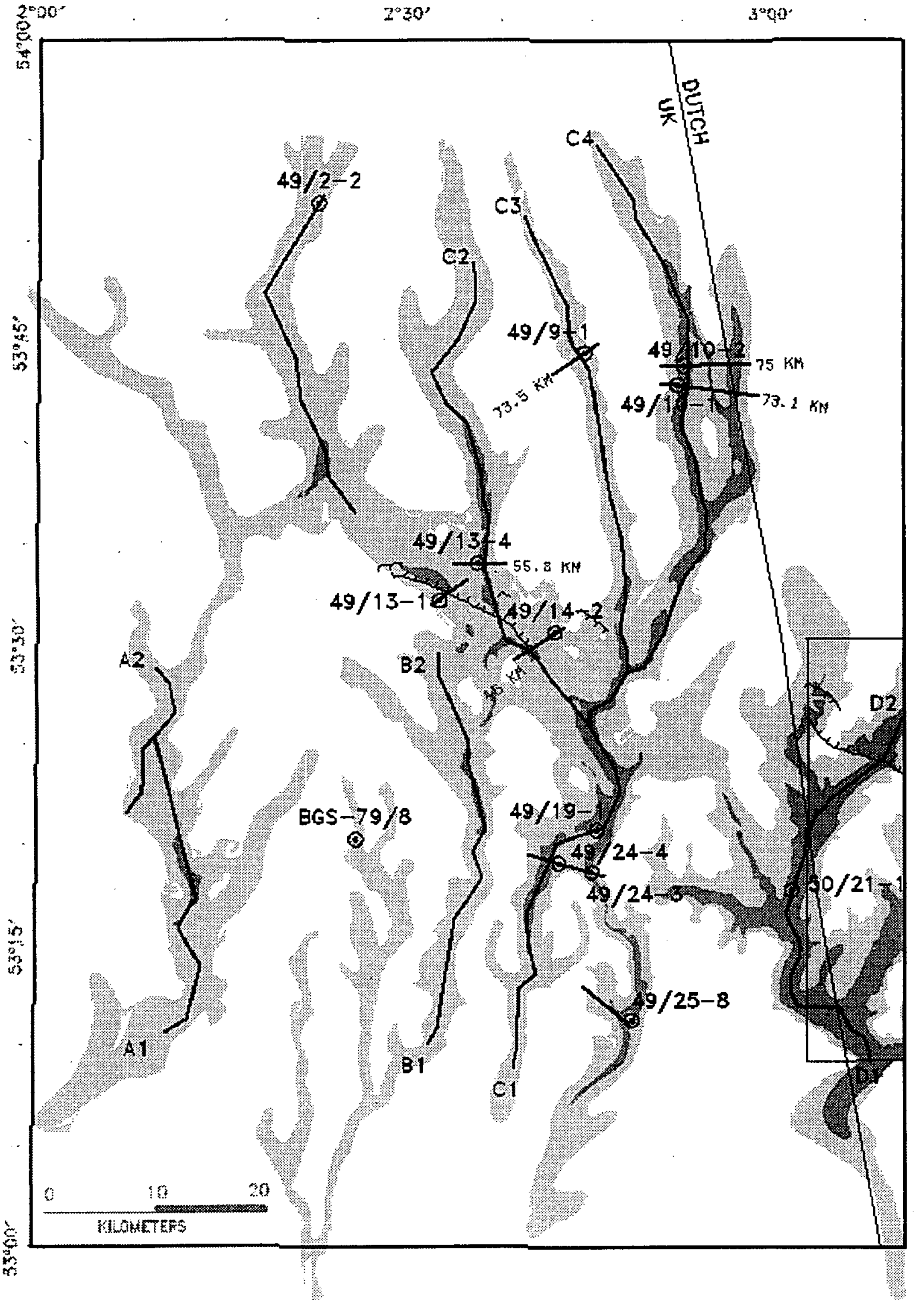


Figure 3.1 - Locations of 13 boreholes, superimposed on the valley base below 100 ms (darker below 300 ms; depths >500 ms occur along profile D, see Fig. 2.8 for depth contours). Axial profiles A to D (Figs. 2.10-12) are shown for reference; portions are shown, along with profiles along other bold lines, in individual well figures (see Table 3.1).

Table 3.1 - Holes bored through tunnel-valley fills in the UK sector, and geophysical log and sample data types available.

WELLS/ BOREHOLE		DATE	LAT/ LONG	*1 *K.B. elev.* (ft)	SEA-BED (m)	*2 TOP DATA (m bsl)	VALLEY BASE (m bsl)	SECTION LENGTH (m)	LOG / SAMPLE DATA *3 GEOPHYSICAL GR	SO	EL	*4 CUTTINGS int.	mp	Figure	Table
49/2-2	MOBIL	12.69	53°51'57" 2°22'41"	*88*	36	43	183	140	√			-		3.2	
49/9-1	SHELL	3.77	53°44'30" 2°44'16"	*110*	38	110	242	132	√			-		3.3	
49/10-1	SHELL	7.71	53°42'57" 2°52'00"	*81*	37	98	269	171	√	√	√	10'	√	3.4 to	3.5
49/10-2	SHELL	4.89	53°43'54" 2°52'30"	*110*	37	137	290	153	√			30'		3.6	
49/13-1	GULF	6.65	53°32'15" 2°32'30"	*31*	29	147	205	58	√		√	31'	√	3.7	3.6
49/13-4	SHELL	7.77	53°34'07" 2°35'36"	*114*	28	78	269	191	√			20'	x	3.8	3.7
49/14-2	TEXACO	6.89	53°30'39" 2°41'55"	*124*	29	76	130	54	√			30'		3.9	
49/19-1	SHELL	9.65	53°23'48" 2°45'21"	*81*	31	78	310	232	√		√	30'	x	3.10	3.8
49/24-3	SHELL	1.69	53°18'40" 2°44'57"	*96*	31	123	312	189	√	√		20-40'	√	3.11	3.9
49/24-4	SHELL	2.7	53°19'09" 2°42'12"	*97*	33	89	233	144	√	√	√	30-50'	x	3.12	
49/25-8	SHELL	9.88	53°11'21" 2°48'15"	*116*	30	117	275	158	√	√	√	30-60'	x	3.13	
50/21-1	BRITISH GAS	1.83	53°17'48" 3°1'15"	*102*	31	50	506	456	√	√	√	20'	x	3.14	
79/08	BGS	8.79	53°20.33' 2°25.73'	-	21.5	55	110	55	driven cores (Table 3.4)			15'	√	3.15	3.10
							TOTAL:	2133							

*1: K.B.= Kelly Bushing, the drill-floor reference level for downhole measurements.

*2: Upper limit of log or cuttings data (bottom of surface casing; see Figures).

*3: See Table 3.2 for geophysical log types and abbreviations

*4: int = cuttings sample interval during drilling (nominal)
mp = micropalaeontological analyses; available from unpublished reports, or:
x = own subsamples (examined by B. Austin, written communication)

Correlation to the seismic stratigraphy shows that sequences (or facies) do not correspond to lithofacies assemblages. Clinoform surfaces (I) lie within the lower glaciofluvial sands, onlapping surfaces (II) bracket the transition from sands to lacustrine muds and complex surfaces (III) correspond to the upper marine muds. I correlate the latter interval with the Egmond Ground Formation (Holsteinian interglacial). The underlying Swarte Bank Formation is thus divisible into three members: clinoform lower sands, onlapping upper sands, and onlapping muds, which I interpret as a depositional succession of glaciofluvial backfill, proglacial outwash and lacustrine mud.

I begin by discussing the collection of log and sample data (section 3.2). I then characterise the valley fill in terms of lithology and microfossil content (3.3) and relate it to the seismic and regional stratigraphical frameworks (3.4). Finally, I review downhole studies of other Elsterian tunnel-valleys and propose depositional environments for the sediment sequence (3.5).

3.2 Downhole Methods

Here I discuss the principles of drilling and logging and their application to the downhole analysis of unconsolidated sediments in exploration wells. Shallow well data (<500 m) were generally collected incidentally in the course of reaching deeper targets. Analysis of such data relies on an understanding of top-hole drilling procedures and their implications for the borehole environment.

Information on top-hole procedures was difficult to obtain, a reflection of the lack of exploration interest. Bibliographies of well-logging literature compiled by Prensky (1987, 1992) yielded few relevant publications. The following is drawn from general accounts of drilling in Beckmann (1976), Short (1983) and Shell (1983), of cuttings analysis in Chilingarian and Varabutr (1983), Jorden and Campbell (1984, Chapter 4) and Exlog (1985) and of geophysical logging in Desbrandes (1985) and Rider (1991). I also discussed drilling and cuttings evaluation with A. Witterick, a former mud-logger with southern North Sea experience, over the Internet discussion group *sci.geo.petroleum*.

I discuss shallow drilling (3.2.1), cuttings evaluation (3.2.2) and geophysical logs (3.2.3). Then I consider glacigenic well logging, including the analysis of data from the study area (3.2.4). Finally, I describe the collection of cores from BGS-79/8 (3.2.5).

3.2.1 Shallow drilling

Drilling is effected by two complementary motions, the abrasive action of a bit on the end of a string of pipe and the hydraulic action of fluid circulating down the interior of the pipe, out the bit and back up the annulus between the pipe and the borehole wall. Penetration depends on both the force applied at the bit and the ability of the fluid to remove debris to the surface. Drilling fluid is referred to as mud, although the shallow section offshore is normally drilled using seawater 'thickened' via the incorporation of mud during drilling or through the introduction of additives (typically bentonite gel). Muds serve to counterbalance formation pore pressures, leading to invasion of permeable formations. Ideally, a mud or filter cake forms at the borehole wall preventing fluid loss and protecting the wall from erosion. However, in shallow holes low-density muds and wide diameters promote turbulent flow, which both inhibits filter cake formation and promotes erosion.

Smaller diameter bits are used as drilling progresses so that borehole diameter telescopes downwards. A surface hole up to 36" wide may be drilled up to 100 m below seabed bed and stabilised with 30" casing/cement, or a 30" metal pipe may be hammered into the seabed and drilled through with a 26" bit. The 'conductor' hole is generally lost to interpretation. A 26" bit is standard below the 30" casing, stabilised with 20" casing and cement. Deeper bit sizes vary but are always of relatively large diameter in the shallow section (≤12" - Table 3.2). Borehole stability is maintained by the periodic cementation of metal pipe or *casing*. Casing is suspended in the hole, cement is pumped down the interior and back up the annulus. Requisite cement volumes are calculated from annulus size, which in shallow sections are nominally up to 3" in radial width (see Table 3.2). In unconsolidated sections, cement volumes may be more than doubled to accommodate borehole size (Short, 1983), which may greatly exceed bit size.

The stability of boreholes in unconsolidated sands was discussed by Stephens et al. (1992) for different situations of effective pressure. As sands have no cohesive strength borehole stability depends on effective pressure, but even where this is positive strength remains zero at the borehole surface. A filter-cake will act to stabilise the borehole but may be easily eroded by turbulent flow. Sands are therefore highly susceptible to erosion by the circulating mud. Unconsolidated sediments are also susceptible to mechanical failure, by plastic flow in response to differential stresses along the borehole wall (or simply the action of the drillstring). Jorden and Campbell (1984) cite a study of stresses in sand around a borehole by Risnes et al. (1982) who modelled the development of a plastic state in an originally elastic, homogeneous medium and predicted a zone of low shear strength in poorly consolidated materials up to 0.9 m in radius.

Erosion or failure tends to lead to the formation of cavities. *Caving* may involve increases in borehole diameter of over 100%. Misk et al. (1977) presented examples of the progressive degradation of boreholes over the course of several days. Caving is a serious problem in unconsolidated sediments, especially sands, as it affects the interpretation of both cuttings and geophysical log data.

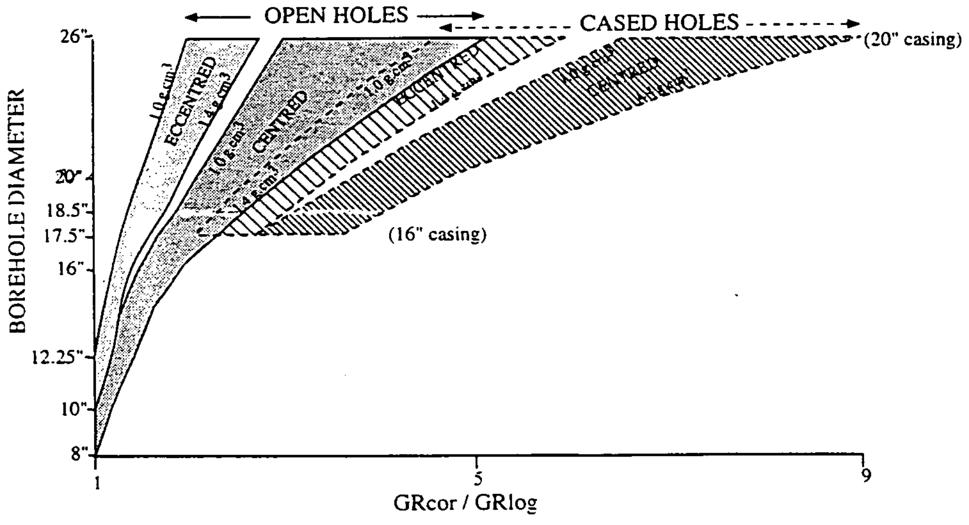
3.2.2 Cuttings evaluation

Drilling debris or cuttings provide direct evidence for downhole sediment properties. The term is a misnomer in unconsolidated materials as drilling acts to disaggregate rather than yield discrete lithological fragments. Cuttings debris is removed prior to mud recirculation by passage across a shaker screen, through settling tanks and if necessary through hydrocyclone de-sanders and de-silters. Samples are taken at regular intervals (5-50') at depths estimated from the bit depth and the lag-time for circulating fluid to reach surface from the bottom of the hole. Samples from the shaker may be washed to remove drilling mud. Cuttings descriptions are recorded on a 'mud log' along with drilling information.

Table 3.2 - Gamma ray device correction factors for shallow well diameters (cased and uncased) encountered in the study area (derived from charts Por.-7&8 in Schlumberger, 1969; 26" by extrapolation).

			CENTRED			ECCENTRED		
BIT SIZE	CASING DIAMETER (width*)	CEMENT THICKNESS (radial)	detector distance to wall	mud density (g/cm3) 1.4 GRcor/GRlog**	mud density (g/cm3) 1.0 GRcor/GRlog**	detector to wall (min/max)	mud density (g/cm3) 1.4 GRcor/GRlog**	mud density (g/cm3) 1.0 GRcor/GRlog**
CASED HOLES:								
36"	30" (1-2")	3"	18"	OUT OF RANGE		≈5.5 / 30.5	OUT OF RANGE	
26"	20" (0.876")	3"	13"	(OUT OF RANGE) ≈9	≈6	5.25 / 20.75	≈6.5	≈4.6
17.5"	13.375" (0.760")	2.1"	8.75"	3.5	2.8	5.5 / 12	3.1	2.6
OPEN HOLES:								
26"	-	-	13"	(OUT OF RANGE) ≈5	≈3	1.8 / 24.2	≈2.8	≈2.0
18.5"	-	-	9.25"	2.6	1.9	1.8 / 16.7	1.8	1.4
17.5"	-	-	8.75"	2.3	1.7	1.8 / 15.7	1.6	1.3
16"	-	-	8"	1.9	1.5	1.8 / 14.2	1.4	1.2
12.25"	-	-	6.13"	1.5	1.1	1.8 / 10.4	1.2	1.0

* API Standard sizes (from Jordan and Campbell,1984, Chart 4.2); 36" from log notes (varies)
** Relative to response of API calibrated 3.625" sonde eccentred in 10" open hole (or centred in 8" open hole) at mud density of 1.4 g/cm3



Various materials may be introduced into the mud during drilling, some of which are easily mistaken for native components. In the shallow section, lost circulation material (LCM) may include (in addition to bentonite gel) ligno-sulphates or lignite, mica flakes, walnut shells, cellophane, paper or whatever is at hand. Lignite and mica can be confused for native cuttings components.

Evaluation of cuttings is complicated by uncertainties of transport, a function of the type of mud and type of flow (e.g. Desbrandes, 1985, p. 26-28). A *transport ratio* is defined as the ratio of mud velocity to cuttings velocity. Sifferman et al. (1974) showed that for seawater or light muds ratios for 'average' cuttings (3x3x6 mm) in conditions of laminar flow will not exceed 0.5 even at high annular velocities. Transport ratios for sand and silt-sized materials should be higher, especially in conditions of turbulent flow, although this will contribute to erosion and caving of the borehole walls. Transport ratios imply differential transport of different grain sizes and so the *spread* of cuttings originating from a given depth. This is significant for unconsolidated sediments due to the possible range of particle sizes. Clays may be incorporated into the drilling fluids, depending on their cohesiveness, and so tend to be under-represented. In addition, debris is contributed from shallower parts of the borehole by caving. Cavities may also acts as sites of temporary deposition and later recirculation of sediment.

The result of spread and caving is that the depth of origin of cuttings is difficult to determine, even given a close collection interval. Evaluation of lithology from cuttings is therefore complementary, or subordinate, to the interpretation of geophysical logs.

3.2.3 Geophysical logs

Geophysical or wireline logging provides continuous data on downhole properties, amenable to qualitative or quantitative analysis. I am concerned with the qualitative evaluation of lithology (sand versus mud). The following is drawn mainly from Rider (1991), Desbrandes (1985) and Schlumberger (1972, 1974).

Logs available for this study comprise nuclear (gamma ray), acoustic (sonic) and various electrical devices (Table 3.3). The gamma ray log is common to all sections (cased and open) of all wells, whereas sonic and electrical logs are available only for some open-hole sections. Different logs were run in combination on a composite tool or sonde, except some recent gamma ray measurements made during drilling (Table 3.3). Sondes may be held either centred or eccentred (against the wall) in the borehole, which is important in larger-diameter and cased holes (see Table 3.2).

Logging devices are associated with different depths of investigation, or sample volume, which are naturally traded-off against vertical resolution (Table 3.3). Variations in borehole gauge significantly affect log results, through changing the depth of investigation. Caliper data on borehole size were not available in the shallow section, so that borehole conditions had to be evaluated from data quality. This is considered further in discussion of each of the three basic types of data.

Table 3.3 - Geophysical logging devices available, with approximate vertical resolution (emitter-receiver spacing) and radial sample dimensions.

TYPE	DEVICE	Abb.	Vertical Spacing	Radius of Investigation	Comment
NUCLEAR	Gamma Ray (simple)	GR	≈30 cm (12") (per measurement)	≈30 cm	sensitive to borehole size/sonde position
ACOUSTIC	Sonic (BHC)	SO	≈60 cm (24")	12-100 cm	very sensitive to borehole conditions
ELECTRIC		EL			
Resistivity (spontaneous)					
unfocused:	Short Normal	SN	16"	1-3 m	≈uninvaded zone
focused:	Laterolog	LL7	32"		
	Spherical	SFL	12"	<1 m	≈invaded zone
Conductivity (induced)					
	6FF40	IL	40"	1-5 m	uninvaded zone
	Dual: medium	DILm	40"		
	deep	DILd	40"		

TOOL COMBINATIONS:

BSGR	SO+GR
IES	6FF40+SN+GR
ISF	SFL+ILd+SO+GR
'GRANDSLAM'	up to 9, including CALIPER+DIL+SO+GR
PLUS:	individual GR / LL7 runs GR measured while drilling (MWD)

3.2.3 a) gamma ray

Gamma ray logs provide information on lithology in open and cased holes and proxy information on the borehole environment (caving).

Gamma ray devices measure the spontaneous radioactivity of the borehole environment, reported in API units, which are relative to the difference between two radioactive standards in a reference well in Texas. The depth of investigation depends on absorption in the borehole environment due to Compton scattering, a function of density. Contributions to the overall signal decrease asymptotically away from the detector and at typical densities it is estimated that 75% of the signal comes from within 30 cm of the detector. Measurements are made over time-constants (1-6 seconds) which depend on logging speed such that the device travels no more than 31 cm per constant. Gamma ray emissions remain subject to a statistical variability in time such that shorter measurements and lower counts are less precise.

The shallow depth of investigation means that gamma ray logging is sensitive both to variations in borehole environment and sonde position within the borehole. This is illustrated in Table 3.2 where correction factors are derived (from Schlumberger, 1969) for centred and eccentred sondes in open and cased holes of different sizes. Correction factors are lower for eccentred sondes as detector separation from the borehole wall remains constant (sonde radius) in one direction, but increase markedly with borehole diameter and in cased holes. Much of the data in this study comes from 26" holes for which centred sondes are likely to yield invariant values. This discussion can be extended to variations in borehole size due to cavities. Low, invariant gamma ray log character thus provides a proxy indication of poor logging conditions.

The lithological significance of the gamma ray log lies in the distribution of the three elemental sources (^{40}K and families of Th and U) the first two of which show a first-order association with clay. A simple quantitative relationship can be applied to a given log section by extrapolation between minimum (100% sand) and maximum (100% clay) values. Empirical testing indicates that the relationship is non-linear and to a greater degree in unconsolidated sediments (Dresser-Atlas, 1982). Also, factors other than clay content may influence readings in sandy materials - common radioactive components are mica, glauconite and feldspar (all ^{40}K), heavy minerals (Th) and lithic fragments (clay).

Log shapes or patterns nonetheless may provide persuasive evidence of progressive changes in texture, notably coarsening- or fining-up sequences. Rider (1991, p. 79) notes that empirically such log patterns may be taken as evidence of grain-size changes, but a lack of log variability cannot be interpreted. Gamma ray log variations thus provide a qualitative indication of sand versus clay content, ideally subject to corroboration by other logs and cuttings.

3.2.3 b) sonic log

The sonic device measures the transit time of compressional waves (tens of kHz) moving from emitters to receivers at either end of the tool, the separation of which approximates the vertical resolution (Table 3.3). The device is designed to be deployed centred in the hole so the sound pulse radiates symmetrically. The depth of investigation is theoretically 12-100 cm depending on wavelength and formation velocity. However, in practice sound moves along the walls so that sonic logs are very sensitive to variations in borehole gauge - caving leads to extreme variations in apparent transit times (Rider, 1991).

Sonic logs thus provide proxy indications of borehole conditions. Velocity is not diagnostic of lithology but as with the gamma ray device log patterns may provide sensitive indications of textural change. In addition, well correlation to seismic reflection data may be obtained by travel-time integrations calibrated by downhole check-shots.

3.2.3 c) electrical

Electrical devices measure the natural resistivity or induced conductivity (=1000/ resistivity) of formations. A range of devices is designed to measure resistivity at increasing depths from the borehole wall, in successive zones of drilling fluid invasion. All comprise arrays of emitting and measuring electrodes (or transmitting and receiving coils for induction) the separations of which are related to resolution and depth of investigation (Table 3.3). The distribution of signal also depends on focusing: the 40" induction devices include medium and deep pairs, designed to obtain different information across the same depth range. Depth of investigation is not strictly defined but is considered maximal for the 40" induction devices at 1-5 m (Schlumberger, 1972). Only some of the deeper devices are used here, in principle measuring the undisturbed formation.

Earth resistivity is expressed as the product of pore fluid resistivity and a dimensionless formation factor, that is: $Ro = Rw \times F$. The formation factor F is a measure of the passive (non-conductive) influence of the matrix on electrical current in the pore fluid. It is found to be constant in given sandy intervals, whereas in conductive clays it is variable. Absolute resistivity values depend on the comparative resistivities of drilling and pore fluids. Qualitative indications of lithology derive from the basic equation: assuming a constant Rw, variations in Ro can be related to changes in F. This may be due to changes in texture or to (related) changes in porosity. Porosity changes may be inferred from paired deep/medium device readings - separations indicate increased porosity (non-porous materials being equally resistive). As with other logs, vertical variations in resistivity patterns may provide sensitive indications of textural changes.

Deep resistivity data have the advantage of being relatively insensitive to variations in borehole gauge. However, deep incursions of drilling fluid or large cavities, both of which are likely in large boreholes in unconsolidated sands, may generate apparent vertical variability.

3.2.4 Glacigenic sediment logging

Here I a) review previous experience of logging glacigenic sediments, b) describe the acquisition and analysis of well data from the UK sector and c) present an example of valley fill logging.

3.2.4 a) previous experience

Well logging as a general method is of demonstrated utility in glacigenic sequences, including tunnel-valleys. It has not previously been applied to North Sea exploration wells.

Downhole logging of glacigenic sediments has had wide application both in western North America (e.g. Dyck et al., 1972; Baldwin and Miller, 1979) and in northern continental Europe (Ehlers and Iwanoff, 1983). Scientific study has been consequent on groundwater exploitation, which involves the rotary drilling of relatively narrow (≤16"), uncased, freshwater-filled boreholes and the collection of electrical and gamma ray logs (among others). It has been found that continuous log data greatly enhance the crude indications of driller's (debris) logs and allow gross discrimination of sandy versus muddy intervals. Log patterns further allow a discrimination of intermediate lithologies and textural (facies) changes and well to well correlation. These results are despite lack of caliper data on borehole diameter.

In northern Germany, combined resistivity (16" Normal Device) and gamma ray logging have allowed the identification and correlation of outwash sands, lacustrine silts, tills and peats (Ehlers and Iwanoff, 1983). These 'best' results depend on a knowledge of local stratigraphy as mixed and muddy sediments, including silts and diamictons, are all associated with similar gamma ray (high) and resistivity (low) values. In buried tunnel-valleys, sand/mud contrasts are found to be favourable for interpretation and correlation (Ehlers and Linke, 1989; Piotrowski, 1994). Comparison of geophysical log results with continuous core data from one valley fill shows a sensitive correspondence between log character and texture (sand/mud) (Ehlers and Linke, 1989, Figure 6).

In contrast, the use of shallow well data for interpretation of unconsolidated Quaternary sediments in the North Sea has few precedents. Caston (1977, 1979) compiled available downhole information, noting that exploration wells provided useful indications of the character of lower Quaternary sediments where thicker but lithological data from shallower depths were of poor quality. Cook et al. (1992) discussed the re-use of top hole (<1000 m) log and cuttings data for site evaluation, noting that data is often suspect and rarely available in the upper 100 m below seabed (conductor hole). Studies of upper Quaternary sediments have mainly relied on drilled and cored boreholes without supporting geophysical log data (e.g. Balson and Jeffery, 1991).

3.2.4 b) available data

Wells were selected through a comparison of locations with the seismically defined valley base (Fig. 3.1) followed by an appraisal of the type and quality of shallow data available (Table 3.1). Wells from the UK sector were preferred due to ready access to the results, notably archived cuttings held in Edinburgh. Also, well logs from the Dutch sector were of generally poorer shallow character (e.g. Pegler, 1995, p. 128-139). The UK wells have been collected over three decades and the type and quality of data varies (Table 3.1). I found that older wells were likely to provide more useful shallow data as subsurface conditions were largely unknown at the outset of exploration, whereas recent wells may bypass the Plio-Pleistocene section entirely.

Well data are placed on file with Her Majesty's Government and in principle are available to the public after five years. Archived cuttings samples and microfiche copies of log results are held in Edinburgh at the D.T.I. core-store (Gilmerton). In practice, it was not possible to obtain copies of logs from D.T.I. and most were obtained directly from exploration companies. Archived cuttings were examined at the core-store and in some cases sampled for microfossil analysis (Table 3.1).

I examined cuttings under a binocular microscope, making visual estimates of composition and the proportions of secondary components (e.g. lignite, shell fragments). Archived samples were either washed or unwashed. Unwashed samples were difficult to evaluate as most samples appeared similar, whereas washed samples afforded ready estimation of the proportions of sand-sized and larger components and of any residual mud fraction. Overall lithology was estimated in conjunction with geophysical logs.

The abundance of certain components, including microfossils, are plotted against depth for some wells to provide an indication of relative vertical variability within a given borehole. Comparison of cuttings depths with prominent lithological contrasts on logs allowed determination of the effective drilling lag (e.g. Desbrandes, 1985). In this study, the valley base often provided such a boundary as did certain sand-mud contrasts within the valley fill. I found cuttings to be 'too deep' by a typical 30-50' (or one to four samples at collection intervals varying from 10' to 60'). An upward correction of up to 40' was therefore applied to cuttings samples where plotted against depth on diagrams. While this produces an apparent correspondence with log results, I recognise that it does not reduce the vertical 'resolution' problems associated with spread and caving.

I obtained copies of the composite log for all wells. This provides a plot against depth of all geophysical data, along with summary cuttings descriptions and interpreted lithology. Composite log interpretations were in some cases consistent with my own and in other cases at odds. This variability appeared to reflect both the difficulties with and lack of interest in the shallow section.

Data were unavailable above depths between 50-150 m bsl (Table 3.1), or 10-120 m below seabed, due to casing emplacement. At greater depths, variability in borehole environment (size, casing) influenced log response, as discussed. Sonic and gamma ray log characteristics were used to infer borehole environment from data quality, as shown in the following example.

3.2.4 c) example in valley fill

An example of the different log data in measured borehole conditions is provided by well 49/25-8, the sole case of shallow caliper data in the study area (Table 3.1, Fig. 3.13). A centred 'grandslam' log combination yielded caliper (2-arm), sonic, conductivity (dual induction) and gamma ray log data (among others), in a nominal 12.25" open hole section below the 30" casing. Seismic data indicate the valley base is located at about 300 ms or 270 m bsl, or 1020' K.B. (±10 ms/30').

The caliper log shows the borehole to be approximately on gauge (bit size) below 1170' K.B. but to abruptly expand above this, via a series of cavities, to beyond the 23" caliper limit above 1060' K.B. The 2-arm caliper has a natural tendency to orient itself along the longer dimension of an elliptical borehole (e.g. Rider, 1991) but significant washing-out is nonetheless indicated above 1170' K.B. The geophysical logs show deflections in close correspondence to individual cavities recorded by the caliper, although a less abrupt increase in borehole diameter is suggested by the sonic and gamma ray devices. Sonic values suggest additional cavities above the top of caliper data at 1060' K.B., then flatten to fluid velocities (1.5 km/s) above 1000' K.B. The gamma ray progressively reduces in value to invariance above 1020' K.B. suggesting coarser sediments prior to loss of signal due to borehole diameter. The latter depth coincides with a separation of the two deep-induction conductivity logs - higher medium-focused conductivity could be due to the large diameter (drilling fluid) as well as to greater porosity or invasion in coarser sediments.

Cuttings data indicate that the transition from on-gauge conditions below 1170' to large diameters above 1000' K.B. also corresponds to a contrast between lower muddy and upper sandy sediments. The changing borehole environment recorded by the geophysical data is thus interpreted to reflect a change in lithology. That is, sandy sediments susceptible to caving are expressed in large borehole diameters (sonic, gamma ray) and resistivity values above about 1000' K.B. Similar correspondence between borehole conditions and lithology (i.e. caved sandy fills) was inferred in other wells.

In general, I have applied redundancy of information to the interpretation of lithology as illustrated above - comparison of cuttings with multiple log where available and correlation to seismic. Multiple datasets are not available in all wells or sections and in two wells gamma ray logs were used alone, without cuttings, as log character suggested interpretable information (Table 3.1).

3.2.5 BGS 79/8 driven cores

BGS-79/8 (Table 3.1) was drilled using rotary techniques following British Geological Survey procedures outlined by Fannin (1989). The hole was drilled without casing using an API standard 5" diameter drillstring and cuttings were returned to seabed. Cores were collected by piling at 15' intervals using 4" inner diameter thin-walled sample tubes. No geophysical logs were run. The 15' (4.6 m) sample spacing is comparable to nominal cuttings intervals in exploration wells (Table 3.1).

The cores average 25 cm in length and recovery over the 189.33 m bore was 9.19 m or 5.3% of total penetration (Table 3.4). Detailed visual descriptions made on collection in 1979 were obtained from BGS. I made an additional examination of archived core halves during 1995 - muddy cores were lithified, whereas sandy cores had largely disaggregated. I made X-radiographs of muddy core halves 6-12 (Table 3.4) - the results were largely homogeneous, save for scattered granular clasts or nodules and abundant cracking and it was uncertain to what extent the character was original homogeneity or reflected 16 years of drying. I therefore relied on the original BGS visual descriptions to characterise lithology.

Subsamples of the cores (Table 3.4) had been taken at the time of collection for micropalaeontological analyses by BGS and RGD. I obtained unpublished reports of the analyses of calcareous microfossils (Gregory, 1980), dinoflagellate cysts (Harland, 1980), and pollen (de Jong, 1981) and present them together for the first time along with the descriptions of core lithology.

Table 3.4: Borehole BGS-79/8 driven core samples, and subsamples (core halves) used for micropalaeontological analyses (see Figure 3.15).

Core #	Depth (m seabed)	Length (cm)	Subsample #	Depth (m seabed)	Length (cm)
(1)	0 (seabed)	(grab)	2820	seabed	<5
2	7.98-8.22	24	2821	7.98-8.04	6
3	12.38-12.70	32	2822	12.38-12.45	7
4	17.27-17.37	10	2823	17.27-17.37	10
5	21.58-21.94	36	2824	21.74-21.94	20
6	26.62-26.82	20	2825	26.62-26.82	20
7	30.79-31.09	30	2826	30.89-31.09	20
8	35.44-35.66	22	2827	35.4635.66	20
9	39.91-40.23	32	2828	40.00-40.23	23
10	44.51-44.80	29	2829	44.50-44.74	24
11	49.11-49.37	26	2830	49.11-49.37	26
12	53.70-53.94	24	2831	53.70-53.94	24
13	57.11-57.51	40	2832	57.11-57.51	40
14	62.84-63.09	25	2833	62.84-63.09	25
15	67.53-67.76	23	2834	67.53-67.76	23
16	71.81-72.23	42	2835	71.81-72.04	23
	"	"	2836	72.04-72.23	19
17	76.44-76.80	36	2837	76.57-76.80	23
18	81.04-81.37	33	2838	81.14-81.37	23
19	85.71-85.95	24	2839	85.76-85.95	19
20	90.25-90.51	26	2840	90.25-90.51	26
21	94.81-95.09	28	2841	94.85-95.09	24
22	99.30-99.66	36	2842	99.42-99.66	24
23	103.89-104.23	34	2843	103.99-104.23	24
24	108.57-108.80	23	2844	108.57-108.80	23
25	113.10-113.37	27	2845	113.13-113.37	24
26	117.76-117.94	18	2846	117.76-117.94	18
27	121.28-121.51	23	2847	121.28-121.51	23
28	125.95-126.08	13	2848	125.95-126.08	13
29	130.46-130.65	19	2849	130.46-130.65	19
30	135.11-135.22	11	2850	135.11-135.22	11
31	139.61-139.79	18	2851	139.61-139.79	18
32	144.10-144.36	26	2852	144.10-144.36	26
33	148.70-148.93	23	2853	148.70-148.93	23
34	153.26-153.53	27	2854	153.26-153.53	27
35	159.00-159.10	10	2855	159.00-159.10	10
36	163.60-163.67	7	2856	163.60-163.67	7
37	168.19-168.33	14	2857	168.19-168.33	14
	AVERAGE	25		AVERAGE	20
	TOTAL	891		TOTAL	750
	RECOVERY:	**5.3%**			

3.3 LITHOLOGY AND MICROPALAEONTOLOGY

The valley fill is characterised from 13 boreholes in the east-central UK sector (Fig. 3.1). The valley base, located by reference to seismic data (section 3.4), was encountered at depths from 100-500 m bsl (Table 3.1). The range in part reflects location relative to the valley axis, but is mainly due to variations in axial depth across the area - the shallowest and deepest examples (BGS-79/8 and 50/21-1, respectively) are both located near valley axes (Fig. 3.1). The upper limit of data (Table 3.1) was below the seismically defined top of the valley fill (40-60 m bsl), with the exception of BGS-79/8.

Exploration wells are summarised in 12 figures (see Table 3.1), which show the environments below seabed (bit and casing sizes), the geophysical log data and interpreted lithology (including cuttings descriptions). Downhole proportions of cuttings components (microfossils, lignite) are plotted where available (e.g. Fig. 3.4). Microfossil analyses are detailed in a series of tables (see Table 3.1). Seismic and regional stratigraphy are correlated against depth on the right-hand side of the diagrams (discussed in section 3.4). A summary is presented as Fig. 3.16.

BGS-79/8 driven cores are summarised in Fig. 3.15. The location of the valley base is uncertain from both seismic and core data to ±10 ms, or ±9 m, or ±2 core intervals of 15' (4.6 m). The base lies in the range of cores 18-21 and the fill extends from core 6 to cores 18-21 for a total of 55-68 m of section. The fill was previously interpreted to extend from cores 6-13, corresponding to 30 m of upper muds (Balson and Jeffery, 1991, Figure 163).

I use the results to distinguish lower sand-dominated lithofacies assemblages, which make up most of the valley fill, from an upper mud-dominated 'cap' and to identify a marine microfossil zone within the upper muds. In addition, I recognise evidence for reworking of Cenozoic sediments, from microfossils (foraminifera and palynomorphs) and component clasts (lignite, shells, glauconite). I discuss lithofacies assemblages (3.3.1), microfossil zones (3.3.2) and evidence of reworking (3.3.3).

3.3.1 Lithofacies assemblages

Well logs and cuttings data were interpreted in terms of three textural groupings, or lithofacies: mainly sand, sand-mud, and mainly mud. Mud refers to silt and clay, which cannot be distinguished in cuttings and are of similar log character (e.g. Ehlers and Iwanoff, 1983). Log patterns were used to infer progressive textural changes (fining- and coarsening-up intervals). Lithofacies interpretations are shown on individual well-diagrams along with generalised cuttings descriptions (Figs. 3.2 to 3.14, see Table 3.1). A graphic summary of cores from BGS-79/8 is presented in Fig. 3.15.

I recognise two main assemblages of lithofacies, a lower sand-dominated interval which overlies the valley base in all wells with thicknesses up to 350 m, and an upper mud-dominated interval which 'caps' the fill with thicknesses up to 90 m beneath the valley-top unconformity (Fig. 3.16). Cores from BGS-79/8 indicate a waterlain character for both. The boundary between the two assemblages is sharply defined from gamma ray log data in northern wells, but less so in wells to the south (Fig. 3.16) which may be due to poor borehole conditions, discussed below.

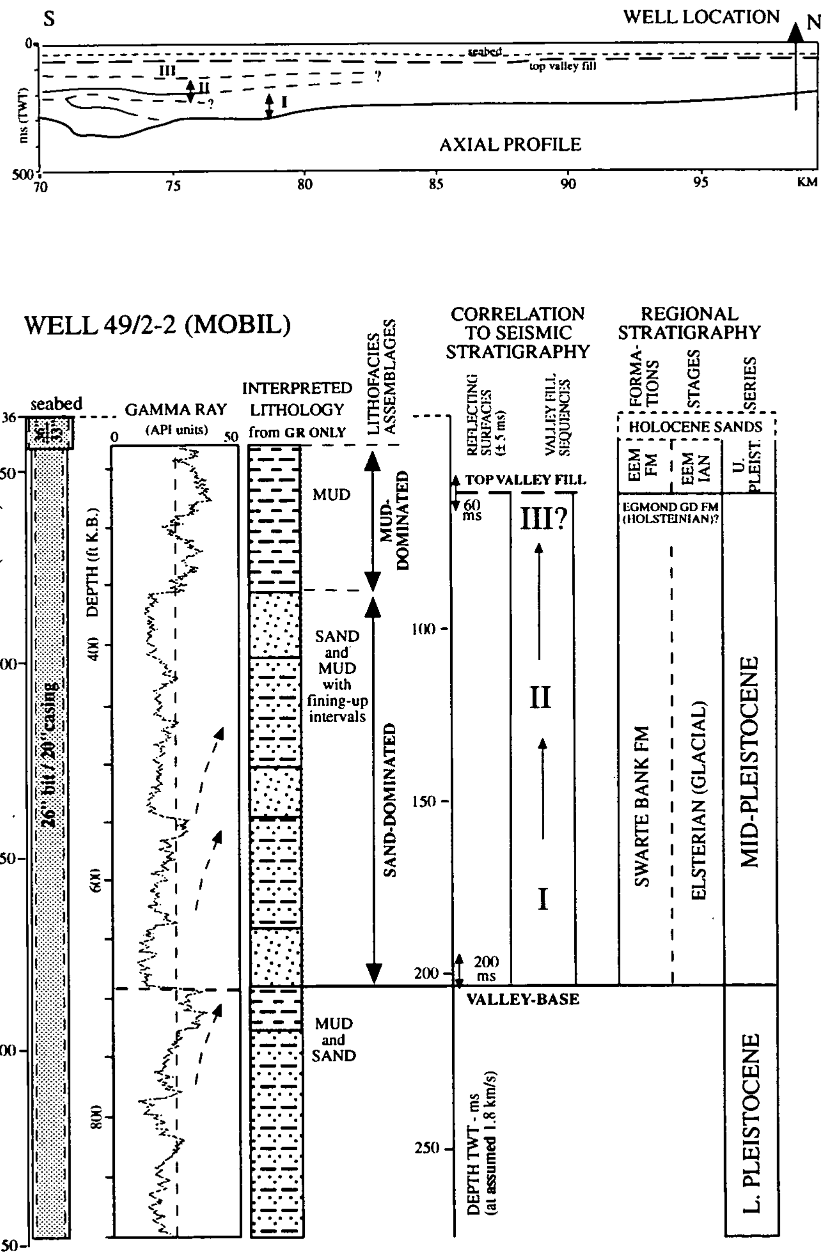


Figure 3.2 - Summary diagram well 49/2-2. The valley fill seismic stratigraphy could not be extended to the well location - some or all of sequences I-III may be present. No cuttings were retained, but the gamma ray log indicates an upwards transition from sand- to mud-dominated lithofacies assemblages comparable to that of other wells (compare Figs. 3.3, 3.4). The axial profile is based on contoured surfaces not presented in Chapter 2 (see Fig. 3.1 for location).

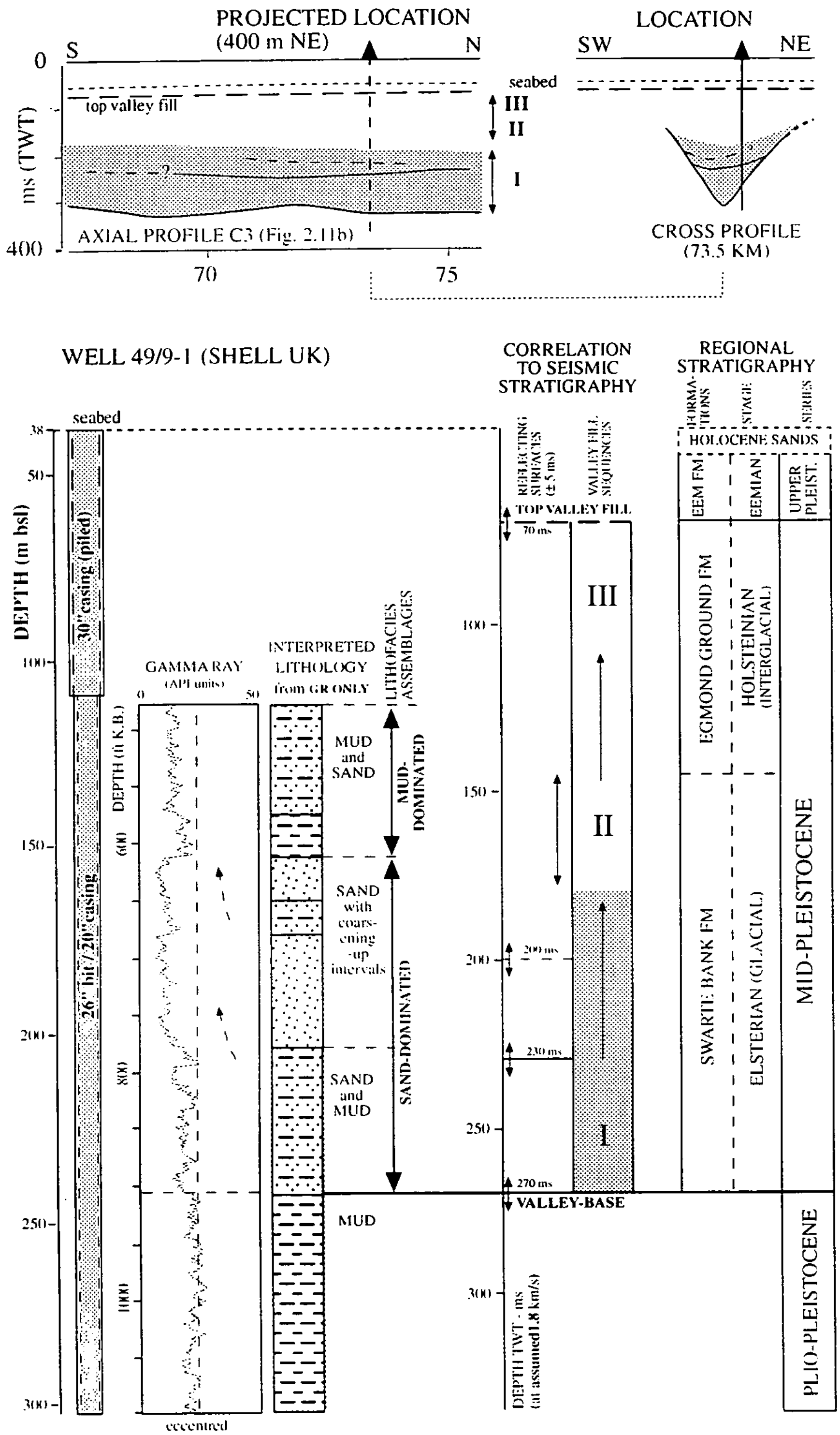


Figure 3.3 - Summary diagram well 49/9-1. No cuttings were retained, but the gamma ray log indicates an upward transition from sand- to mud-dominated lithofacies assemblages (compare Figs. 3.2, 3.4). The boundaries between the seismic sequences are not certain at the well location, but least half the section corresponds to sequence I, within the lower sand-dominated sediments. The boundary between sequences I and II may lie near the base of the upper mud 'cap' (see Fig. 3.4). Sequence III thus lies within the upper mud-dominated sediments. See Figure 2.11a for full axial profile C3 (Fig. 3.1).

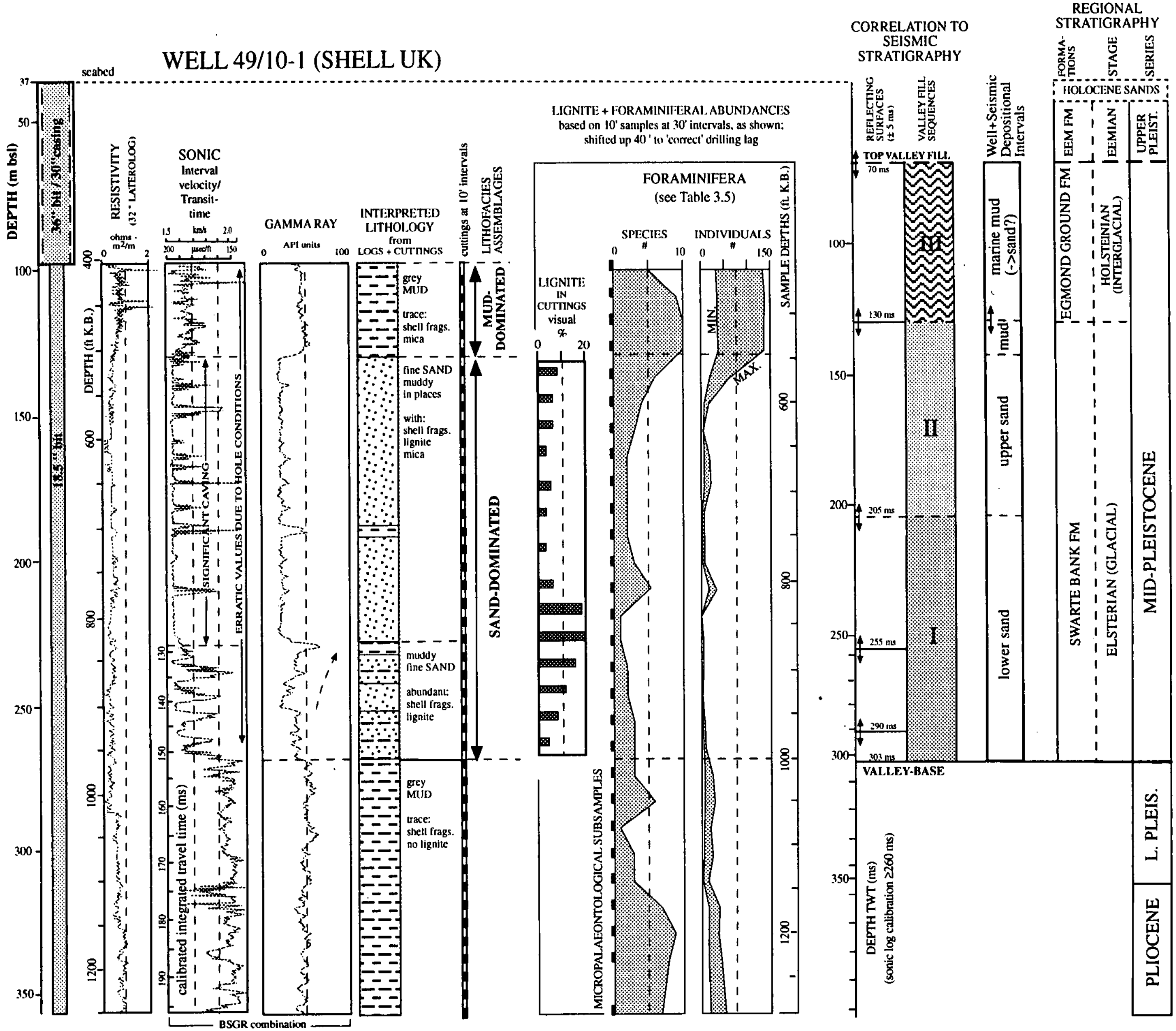
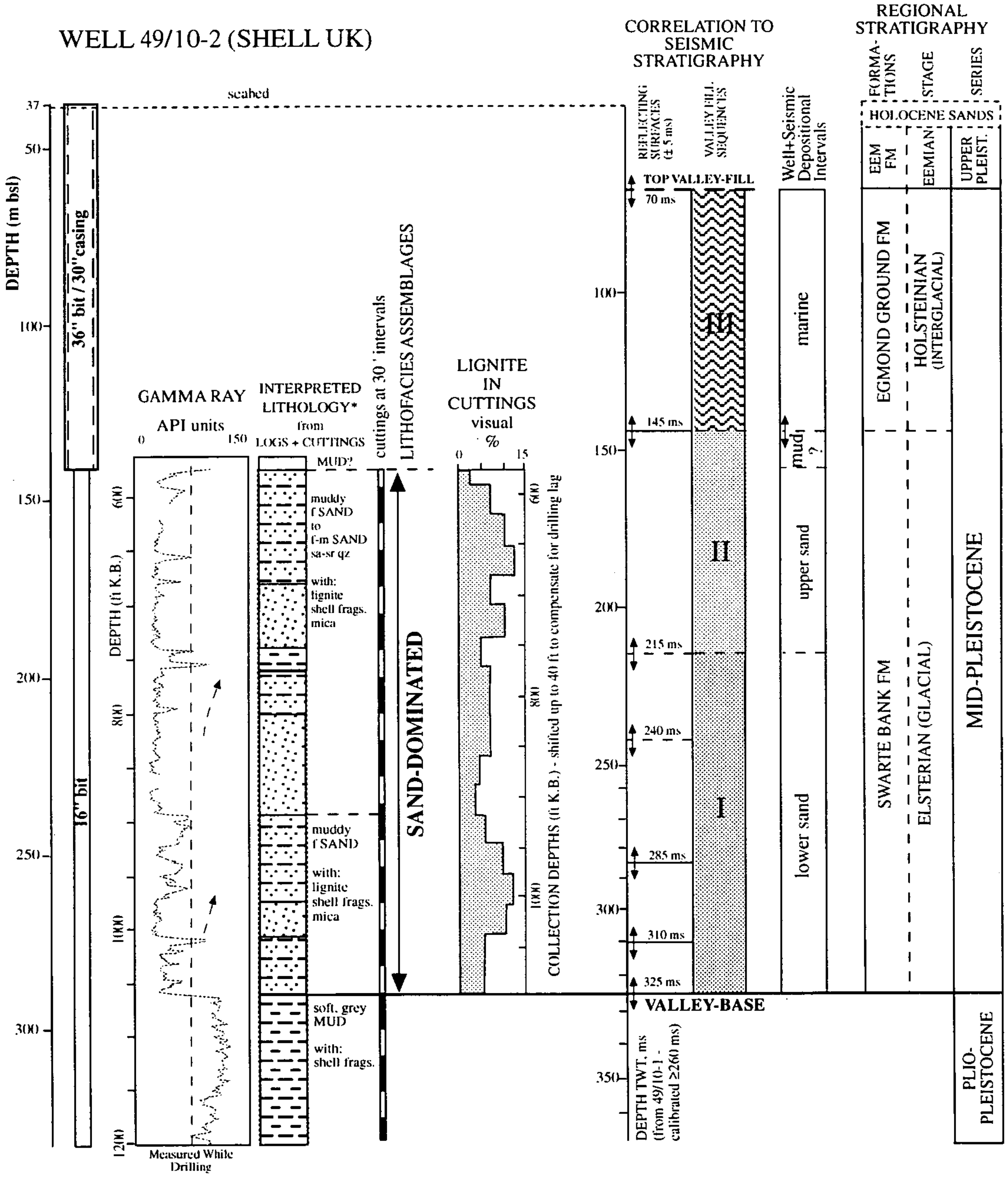
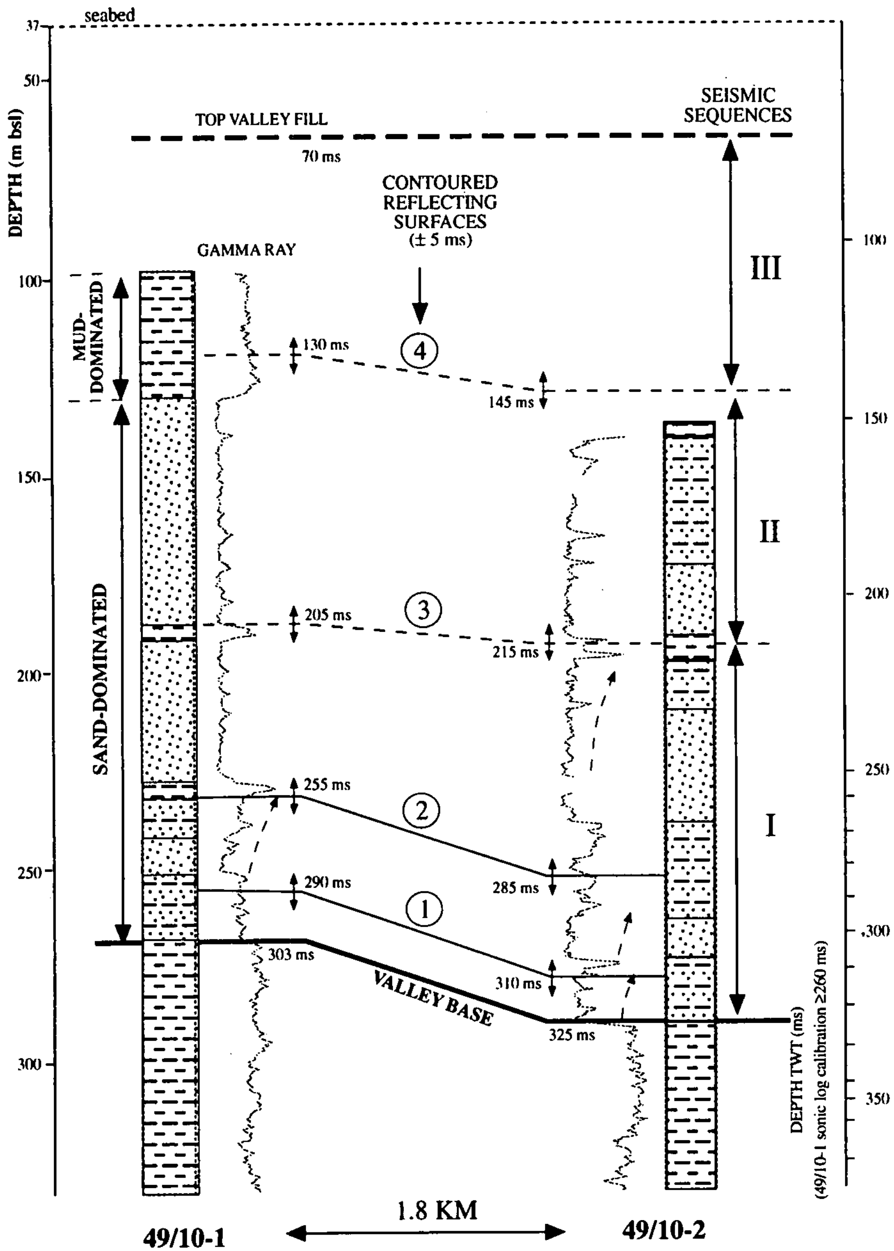


Figure 3.4 - Summary diagram well 49/10-1. Cuttings subsamples have been shifted upward 40' to obtain apparent correspondence with sand/mud contrasts at the valley base, and at the base of upper muds. The lower limit of the marine foraminiferal zone appears to correspond to the base of the muds, but this may be due to contamination from above. See Fig. 3.6 for location along axial seismic stratigraphy, and correlation to adjacent 49/10-2 (Fig. 3.5).



*LITHOLOGY ABBREVIATIONS (also Figs. 3.7-3.14):
f=fine, m=medium, cse=coarse, a=angular, sa=subangular, r=rounded, sr=subrounded, qz=quartz

Figure 3.5 - Summary diagram well 49/10-2. High gamma ray values (measured while drilling) at the very top of the logged interval may indicate mud, supported by correlation to nearby well 49/10-1 (Fig. 3.6). Two peaks in lignite abundance in cuttings also appear to be correlative between the wells (Fig. 3.4). For well to well correlation and location along axial seismic stratigraphy, see Fig. 3.6.



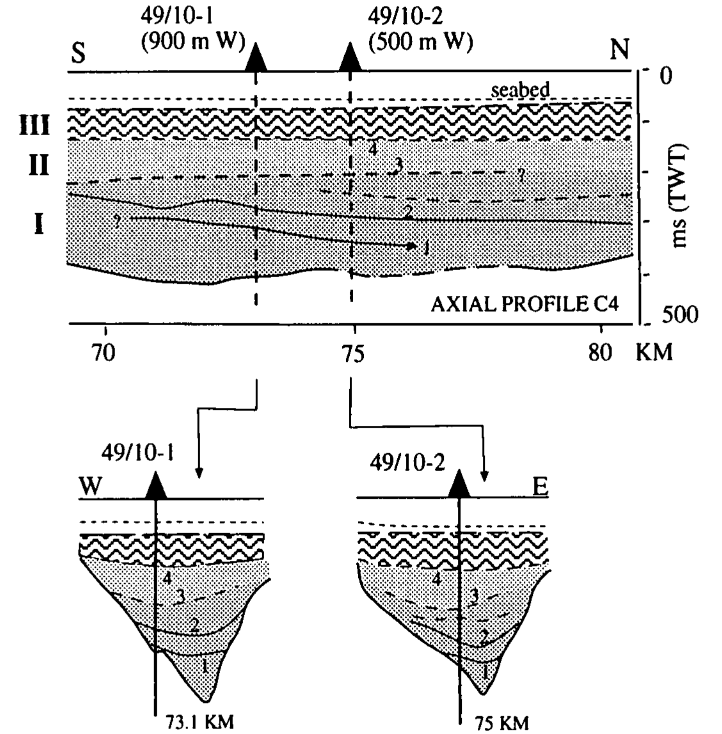


Figure 3.6 - Correlation of adjacent wells 49/10-1 and 10-2 (Figs. 3.4, 3.5). Both are located west of the valley axis. Four contoured reflection surfaces traceable between the wells correspond with log-defined lithofacies changes. The lower two are clinoform surfaces within sequence I, the upper two approximate sequence boundaries I/II and II/III. See Fig. 2.11b for full axial profile C4.

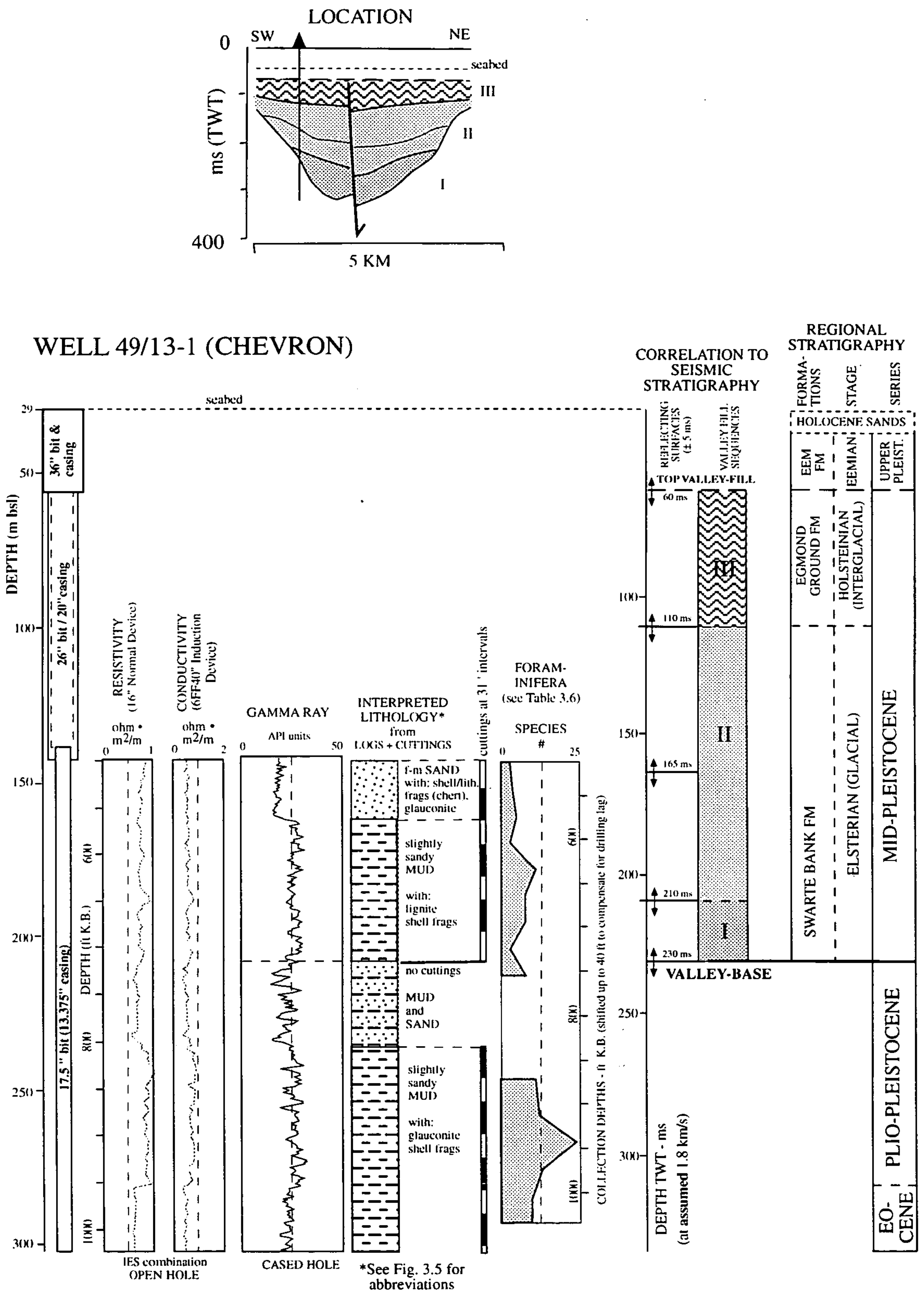


Figure 3.7 - Summary diagram well 49/13-1. The well is located on the flank of a valley up to 350 ms deep (see Fig. 3.1 for cross-section location). The valley base is defined from seismic data. It appears to be overlain by mud-dominated sediments, according to the gamma ray log and cuttings, although electrical log data (collected earlier) does not accord. The lower interval may be a poorly resolved complex sequence, such as observed in wells 49/10-1 and 10-2 (Figs. 3.4-3.6).

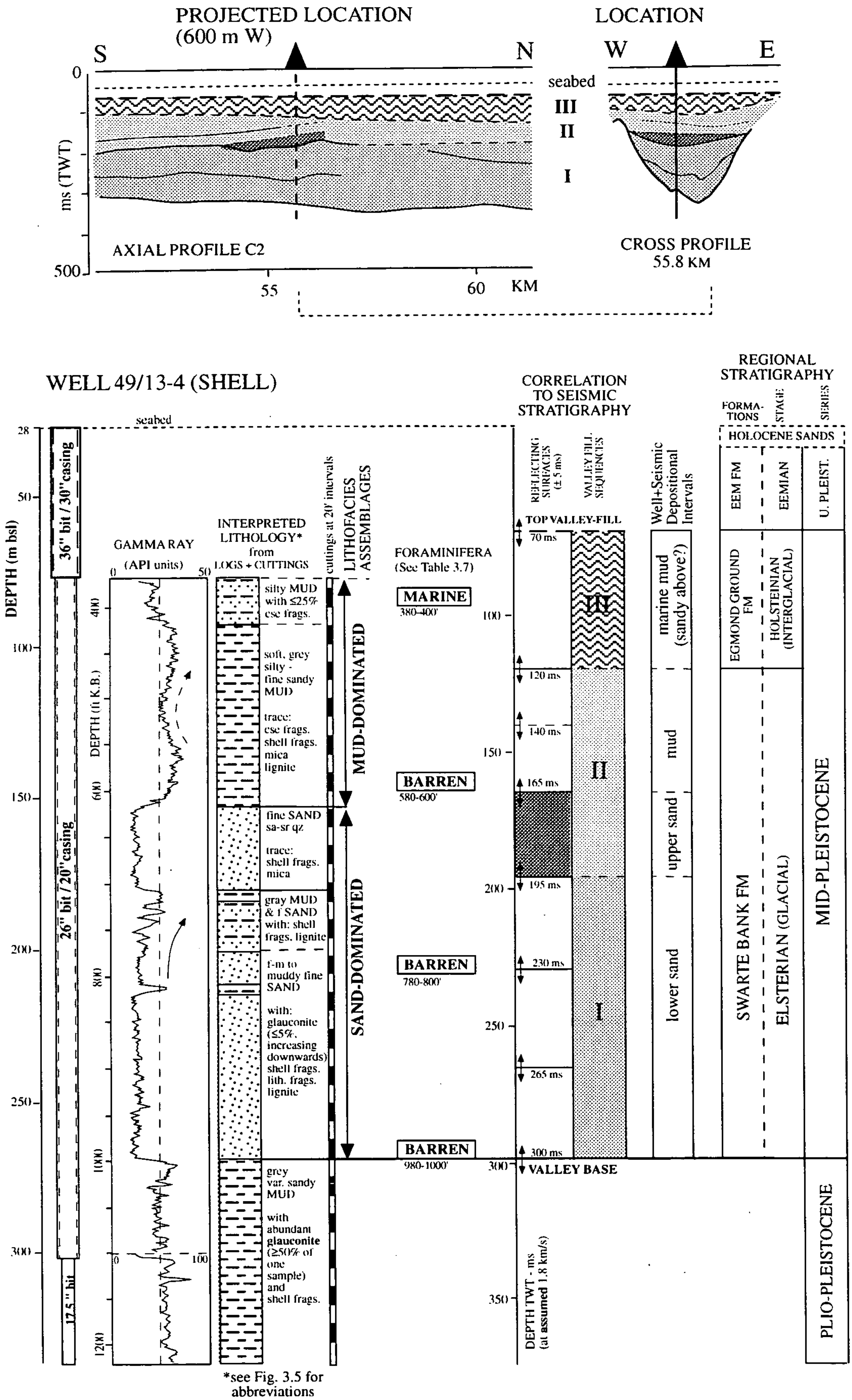


Figure 3.8 - Summary diagram well 49/13-4. The well encountered almost a complete fill sucession of sequences I-III. The lower sand of sequence II corresponds to a gently downlapping lobe (stippled).

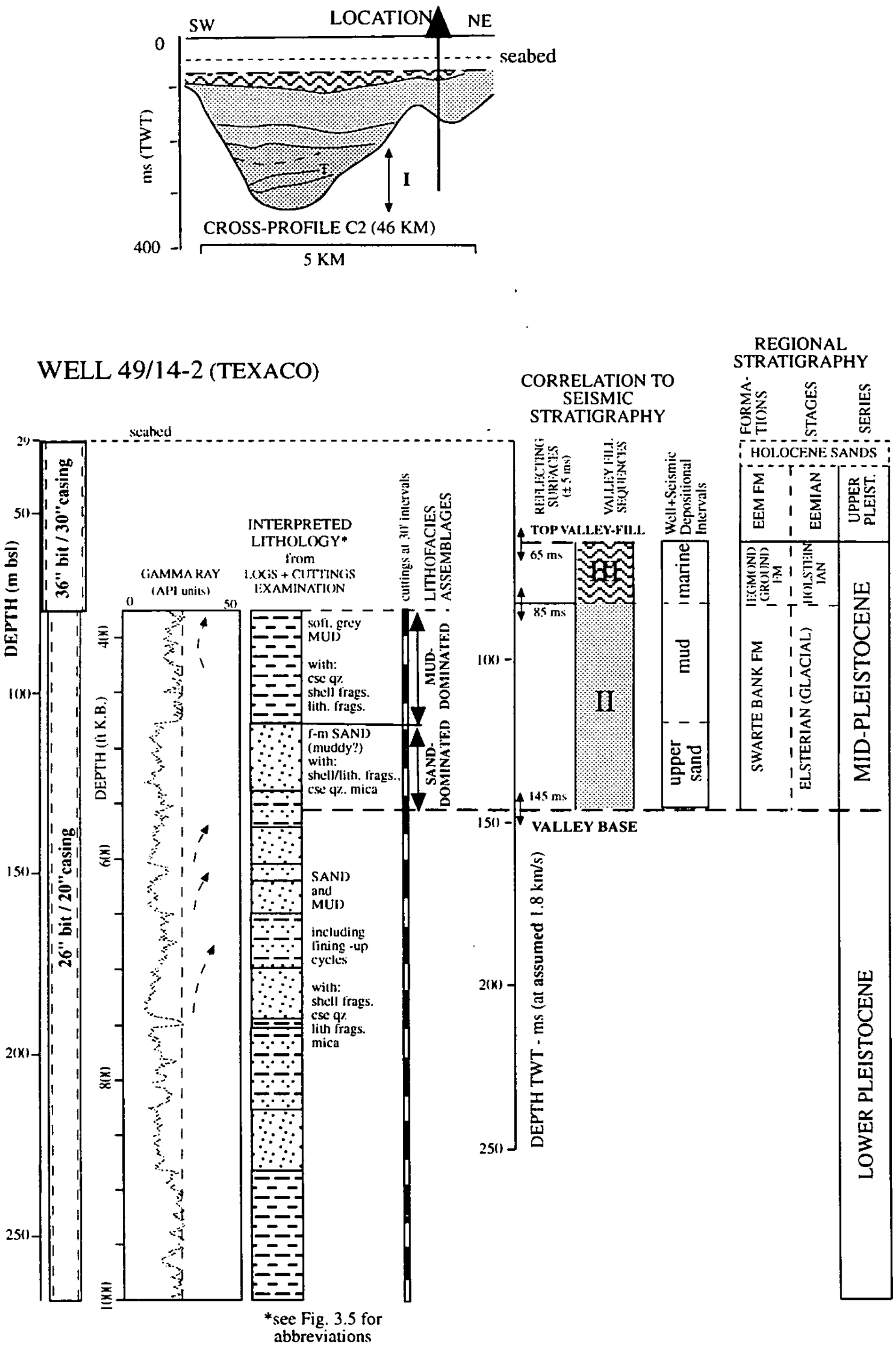


Figure 3.9 - Summary diagram well 49/14-2. The well is located on a valley flank and has penetrated only the upper fill (sequence I). Overlying sequence III corresponds to marine mud in nearby well 49/13-4 (Fig. 3.8; see Fig. 3.1 for locations).

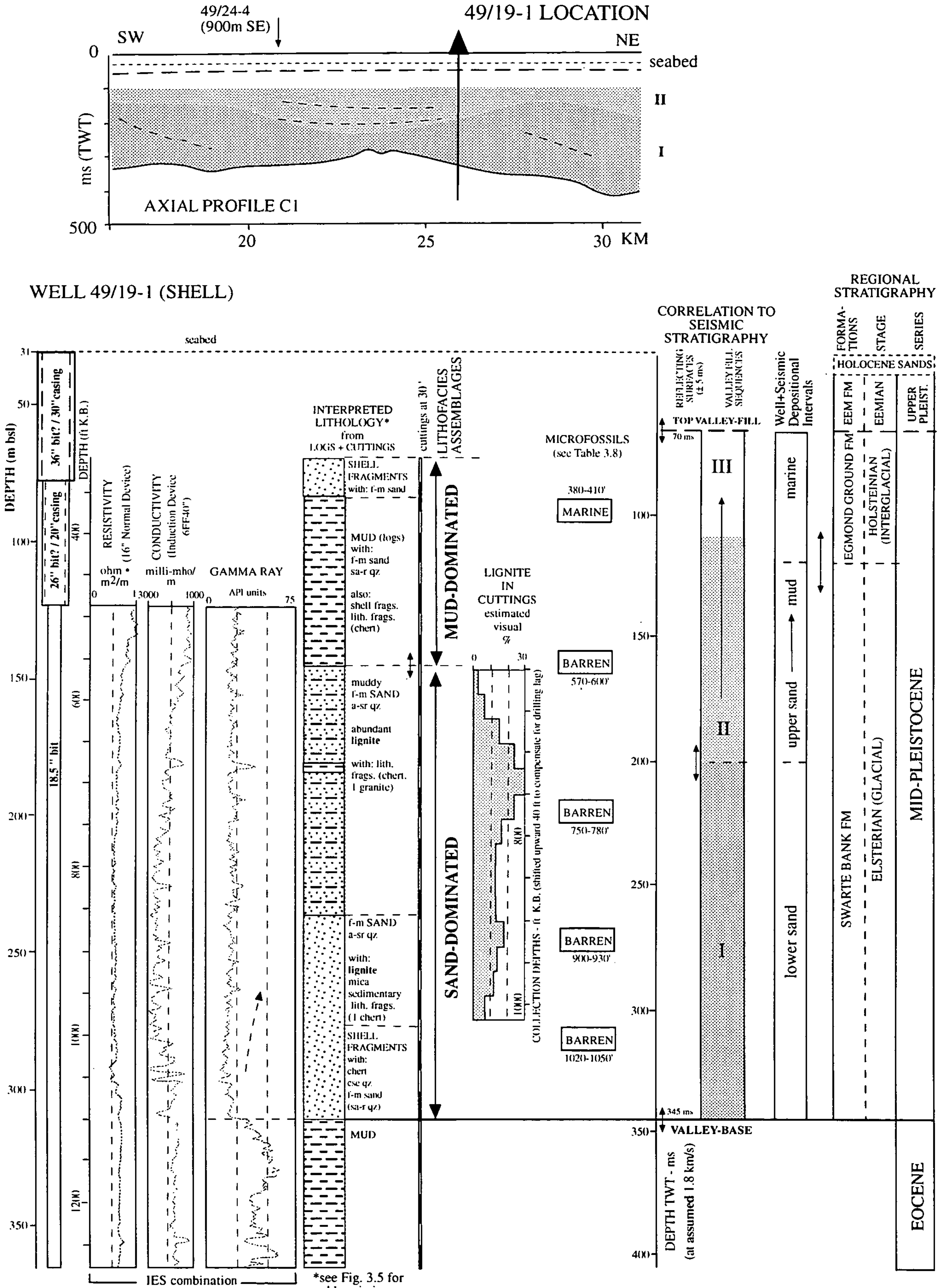


Figure 3.10 - Summary diagram well 49/19-1. Electrical logs indicate muddy sediments at the top of the section, the poor gamma ray response may be due to borehole conditions. Cuttings indicate abundant lignite in the central part of the valley fill and shell fragments at the top and bottom of the section. The boundary between sequences I and II is inferred. Sequence III could not be traced on seismic data but its presence is suggested by marine microfossils in the upper sample (see Table 3.8).

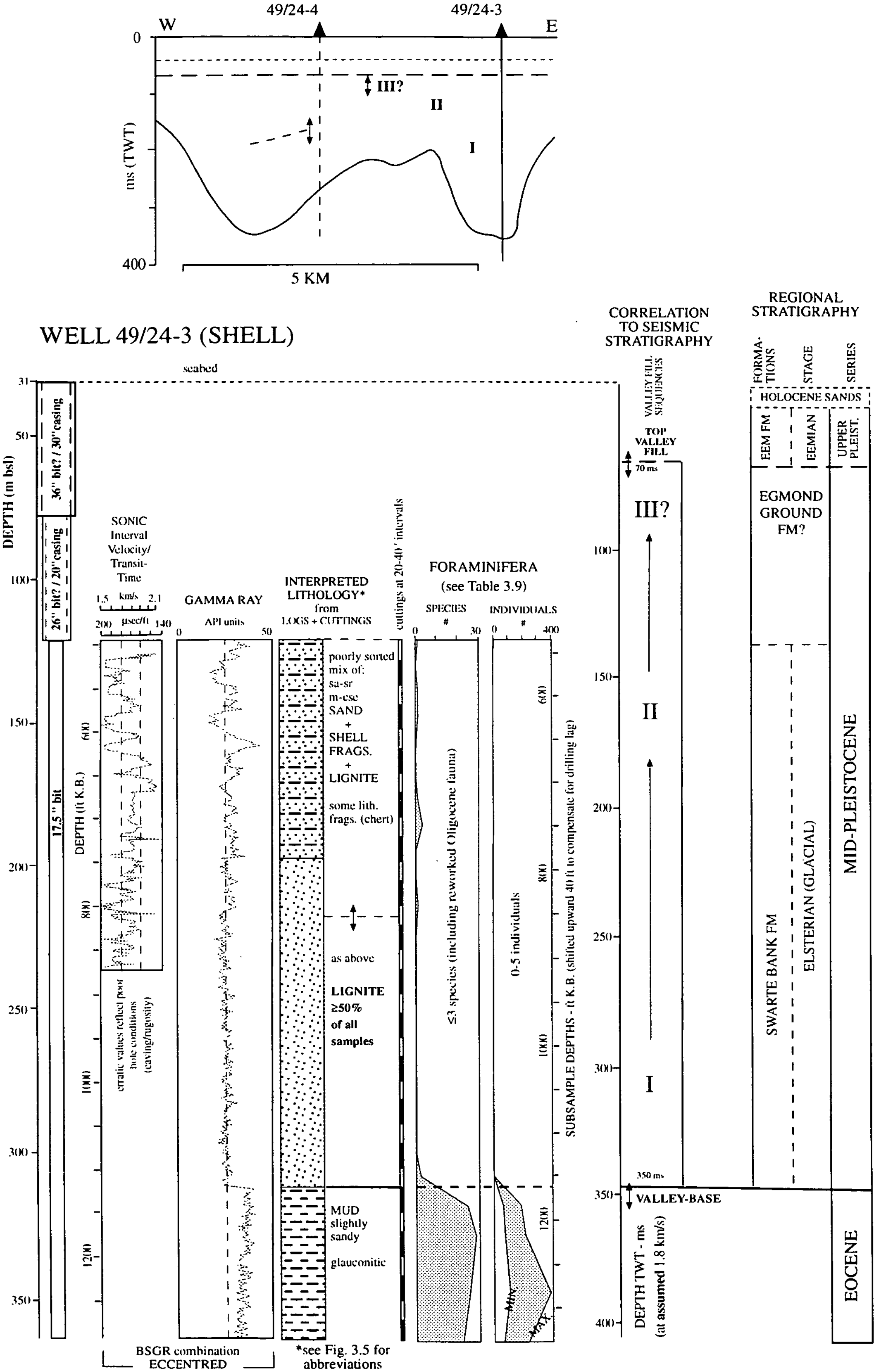


Figure 3.11 - Summary diagram well 49/24-3. The sonic log indicates variations in borehole diameter due to caving at the top of the section, high gamma ray values outwith cavities indicate muddier sediments. Seismic data did not allow contouring of fill reflecting surfaces, but comparison with nearby wells 49/19-1 and 24-4 suggests that sequence I dominates the valley fill beneath sequence II. See Fig. 3.10 for well locations.

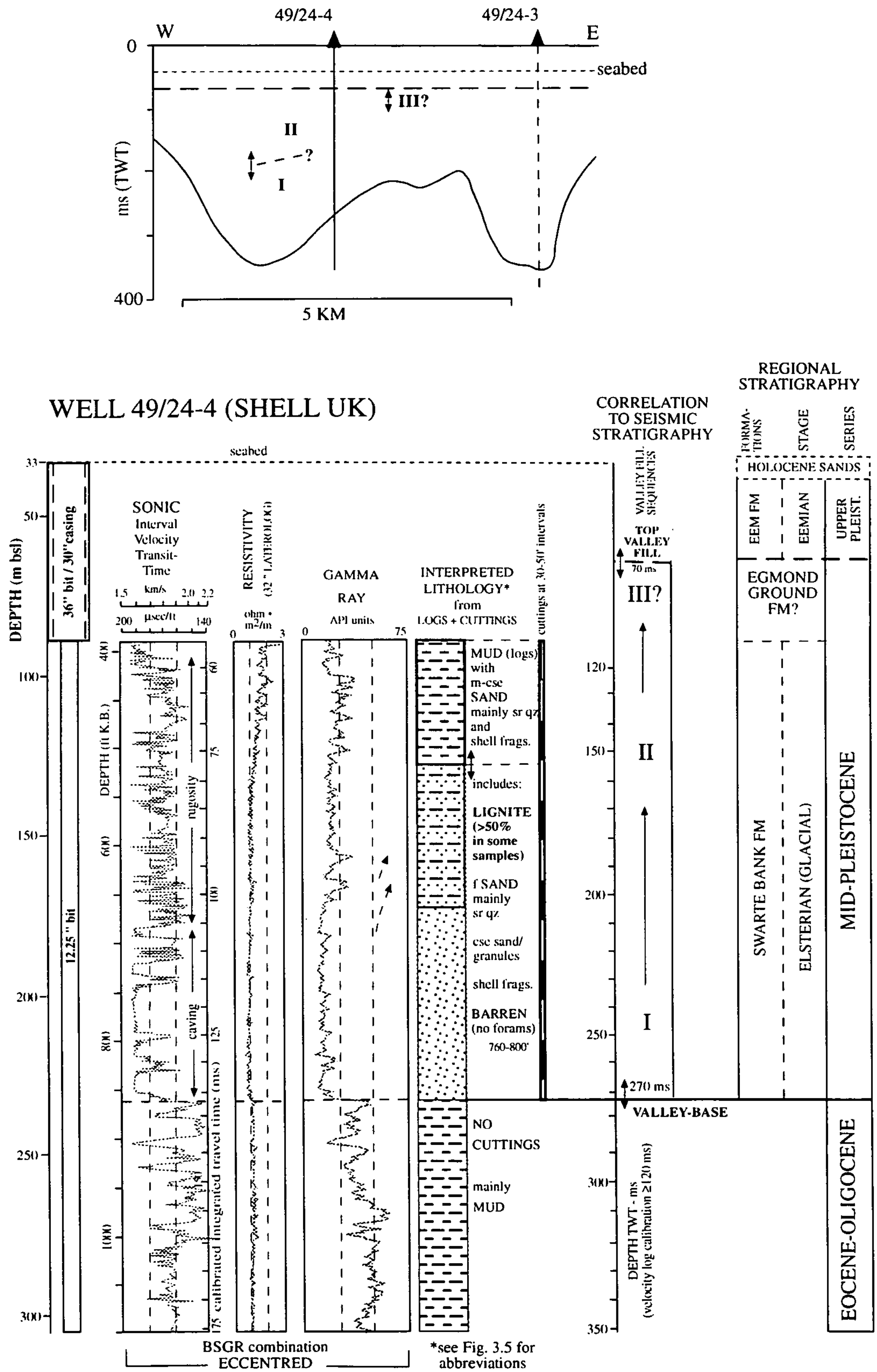


Figure 3.12 - Summary diagram well 49/24-4. The overall fining-up section, with abundant lignite in the lower sandy sediments, is comparable to nearby wells 49/19-1 (Fig. 3.10) and 49/24-3 (Fig. 3.11). The seismic sequences in the cross-section are projected from the axial profile (see Fig. 3.10), and are not certain at the well location.

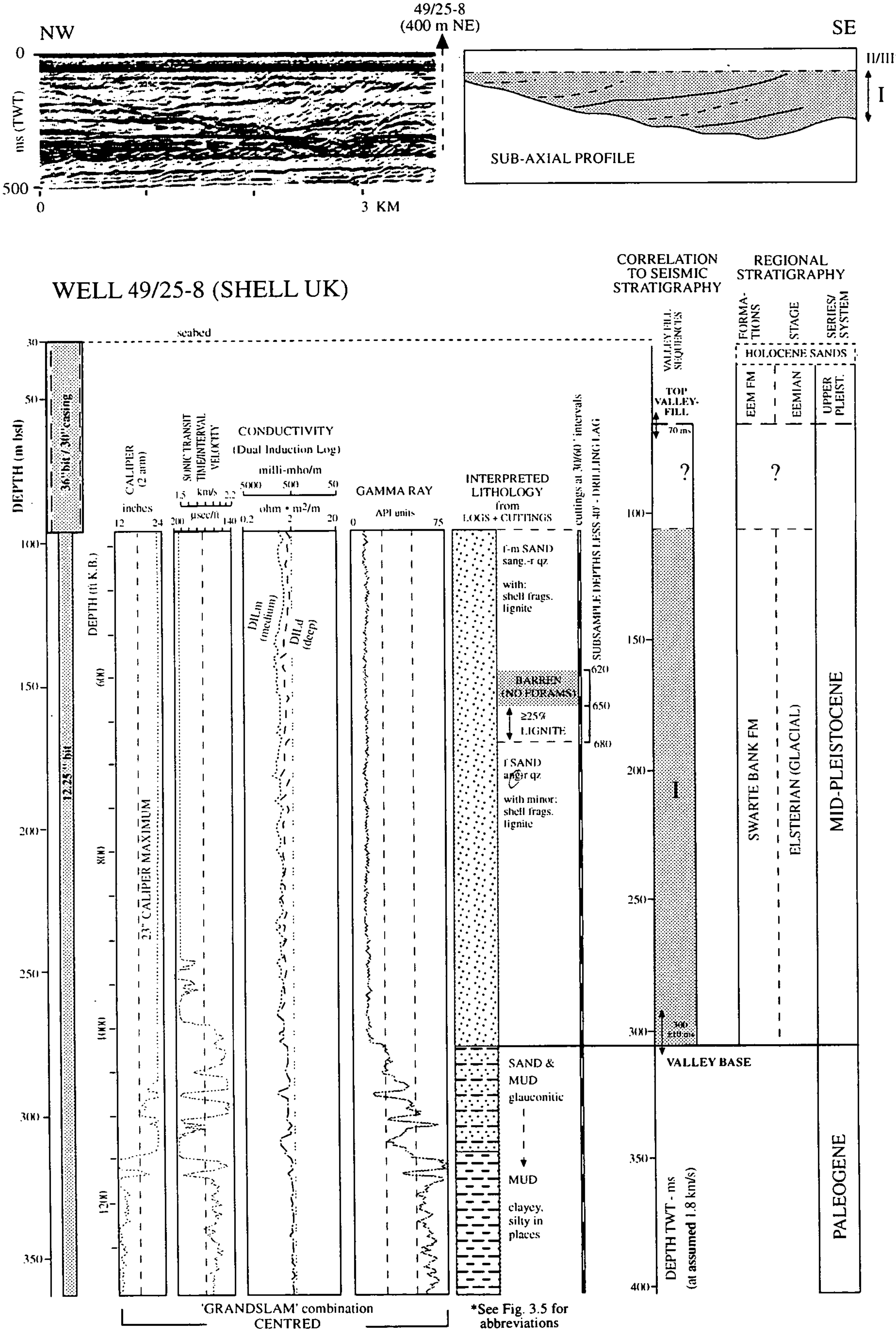


Figure 3.13 - Summary diagram well 49/25-8. Fill reflecting surfaces were not contoured along this valley, but the sub-axial seismic profile shows that the well sequence mainly corresponds to axially downlapping clinoform surfaces of sequence I. See Fig. 3.1 for location.

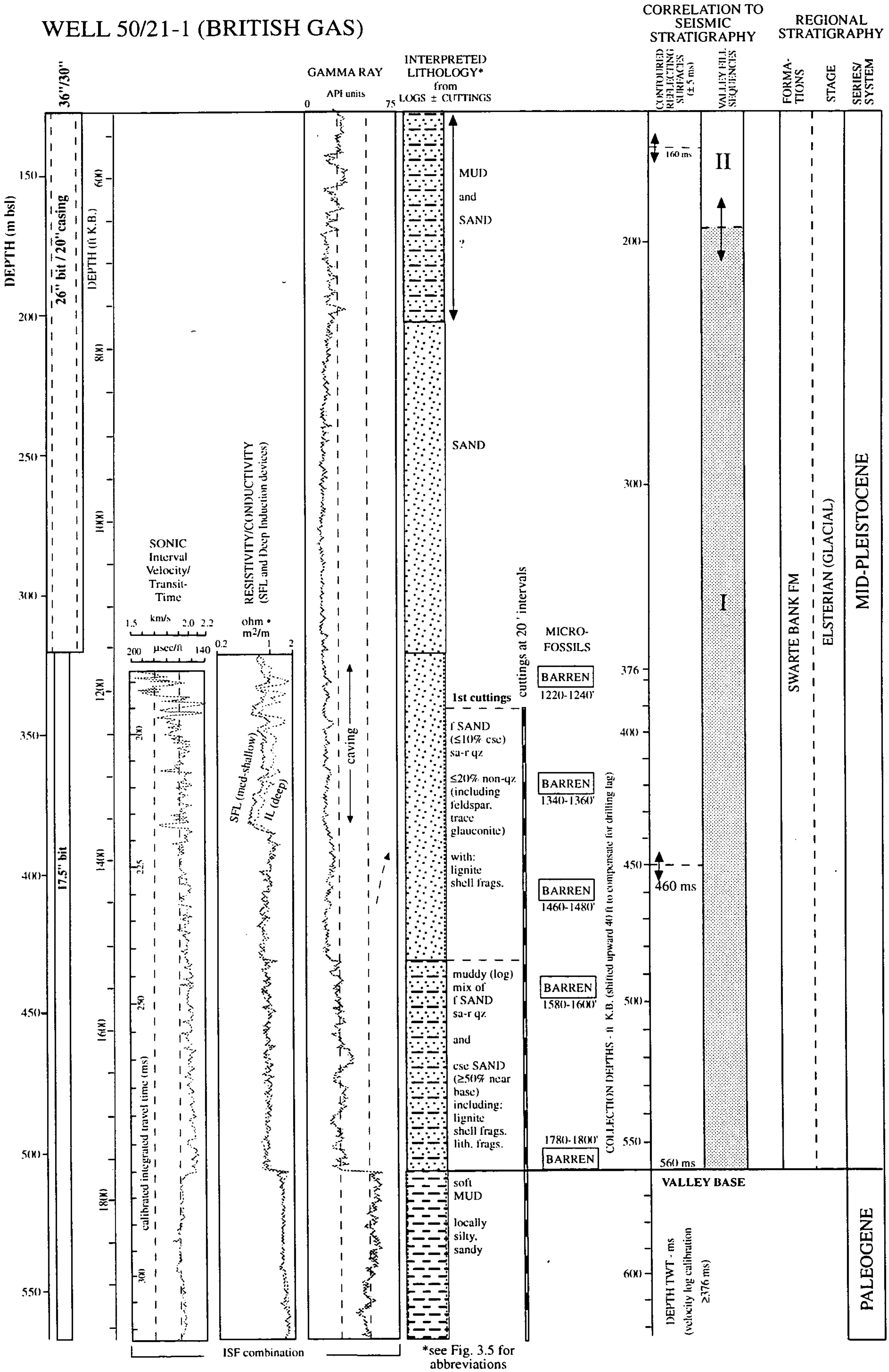


Figure 3.14 - Summary diagram well 50/21-1, located in the deepest tunnel-valley in the study area, 100 m west of the axis (see Fig. 2.12 for location along seismic stratigraphy). The well is only shown from below the base of the surface casing. Log and cuttings data are available from the lower 179 m, within sequence I.

BOREHOLE BGS-79/8 (driven cores, 25 cm average, 5.3% recovery)

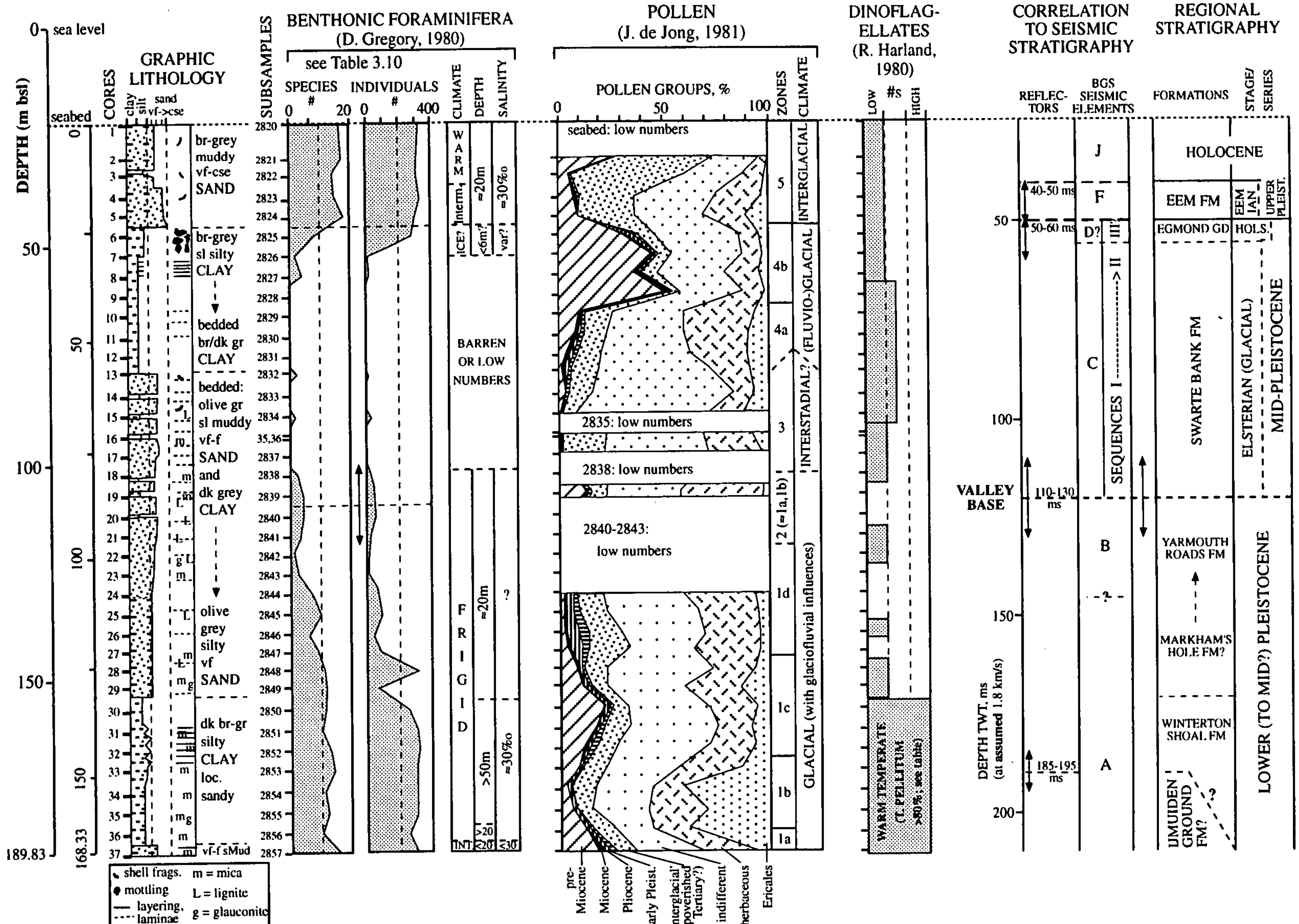


Figure 3.15 - Summary diagram borehole BGS-79/8. The valley base is poorly defined on seismic and sample data, but lies in the range of cores 18-21 (±10 ms or ±2 core intervals). BGS/RGD seismic elements from Cameron et al, 1992, from interpretation of BGS seismic profile 81/38. See Fig. 3.1 for borehole location, Table 3.4 for core and subsample depths, and Table 3.10 for foraminiferal analysis.

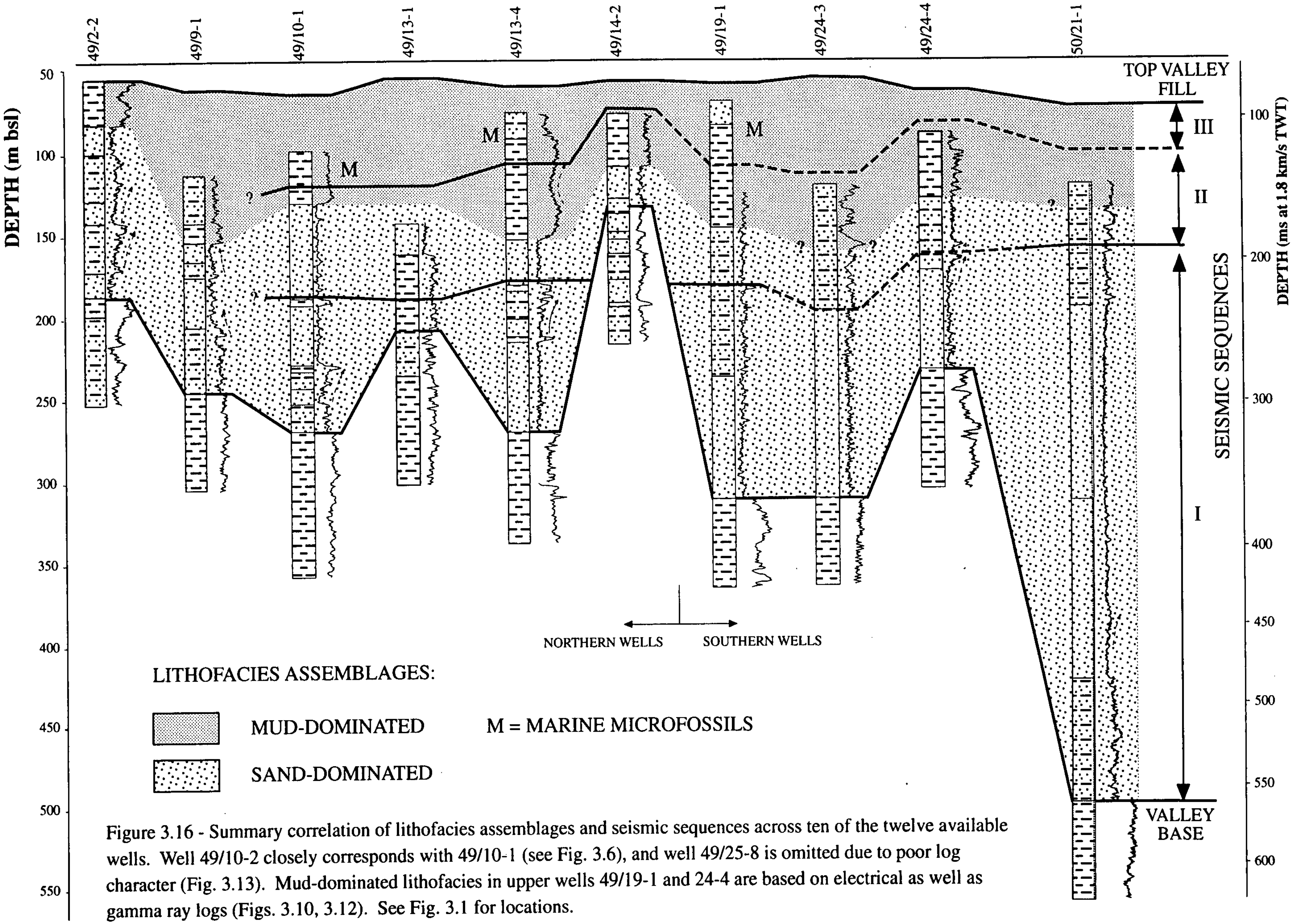


Figure 3.16 - Summary correlation of lithofacies assemblages and seismic sequences across ten of the twelve available wells. Well 49/10-2 closely corresponds with 49/10-1 (see Fig. 3.6), and well 49/25-8 is omitted due to poor log character (Fig. 3.13). Mud-dominated lithofacies in upper wells 49/19-1 and 24-4 are based on electrical as well as gamma ray logs (Figs. 3.10, 3.12). See Fig. 3.1 for locations.

The two lithofacies assemblages are described in turn, followed by a discussion of lignite and shell concentrations observed in cuttings and cores.

3.3.1 a) lower sand-dominated

Assemblages dominated by sand and sand-mud lithofacies, with local interbeds of mud, are recognised in all wells above the valley base unconformity (Fig. 3.16). The vertical assemblage varies in character from well to well and is difficult to generalise. Log patterns include both homogeneous intervals and texturally variable intervals with sharp or transitional boundaries. The latter include fining-up intervals and in two cases coarsening-up intervals, both more common in the lower parts of the assemblage (Fig. 3.16).

Interpretations are based mainly on gamma ray log patterns and shifts. Complementary electrical log data from six wells (Table 3.1) are of comparable character, although of lower resolution due to their greater radius of investigation (Table 3.3). Sonic log data from six wells indicate generally poor borehole conditions, with irregular diameters due to caving. Poorer conditions generally correspond to gamma ray and cuttings indications of sandier intervals (e.g. Fig. 3.4), as shown above for well 49/25-8 (section 3.2.4c).

Gamma ray log patterns are correlative between adjacent wells 49/10-1 and 10-2 (1.8 km apart, Fig. 3.6) and descend gently to the north in parallel with seismic reflecting surface trends (see section 3.4.2a). Sonic log data from 49/10-1 indicate intervals of significant caving, but the gamma ray log from 49/10-2 was measured while drilling and so should be less influenced by variations in diameter. The correspondence of the two shows that gamma ray logs may provide reliable indications of lithological variations within the valley fill, even in the presence of caving.

Well 49/13-1, located on a valley flank, is the only well to include an interval of apparently mud-dominated sediments above the valley base, overlain by sands (Fig. 3.7). Electrical log data collected in an open hole and gamma ray data collected later in a cased hole do not closely correspond. The gamma ray interpretation is consistent with cuttings data. Lower intervals of mud-sand facies and mud interbeds occur in other wells (e.g. 49/10-1, 10-2; Figs. 3.4-3.6) and the lower part of 49/13-1 may be a similar interval, possibly poorly resolved.

Cuttings samples were available for all but two wells (Table 3.1). Cuttings were dominated by very fine to coarse quartz sand (mainly fine to medium), angular to rounded (mainly subangular to subrounded). Ubiquitous minor (<5%) cuttings components included one or more of shell fragments, lignite, mica, glauconite and lithological fragments (granules, mostly sedimentary, notably chert - a single granite granule was noted in well 49/19-1, Fig. 3.10). Clasts larger than gravel-size were only observed as shell fragments. Lignite and shell fragments occur in significant secondary or even primary abundances in intervals of six wells (see below).

In BGS-79/8, the valley base lies in the range of cores 18-21, which bracket an apparent upward transition from fine-medium sand, below core 20, to a similar sand with mud interbeds (Fig. 3.15). Given the core recovery (5.3%), mud interbeds could also be present but unsampled below core 20.

The lower valley fill consists mainly of sand in 6-9 cores (13 to 18-21) spanning a 24-38 m section. The cores are dominated by olive-grey sand (mostly fine to medium, locally very fine to coarse, clean to slightly muddy), interbedded with light to dark grey clay. The sand is in places shelly, lignitic and micaceous (abundances ≤5%), components which also occur in sands below the valley base. Faint to distinct layering is observed within the sand, additional to its interbedding with the mud. The layered character, and moderate sorting, both indicate a waterlain origin for the cored intervals.

3.3.1 b) upper mud-dominated

Evidence for an upper mud-dominated lithofacies assemblage ('mud' and 'sand+mud' facies) is recognised in most wells and in BGS 79/8. Two wells have their upper limit of data below the level of the mud 'cap' (49/10-2, 13-1; Figs. 3.5, 3.7).

The upward change from sand - to mud-dominated lithofacies corresponds to a sharp contrast on gamma ray logs in wells to the north, whereas in wells to the south upper muddy sediments are less clearly distinguished (Fig. 3.16). In two of these wells (49/19-1 and 24-4), resistivity rather than gamma ray data are indicative of muddy sediments (Figs. 3.10, 3.12). This may reflect poor top-hole conditions and the greater radial sample dimensions of resistivity devices (Table 3.3).

Sand incursions within mud-dominated intervals are indicated by gamma ray log patterns in two wells (e.g. 49/2-2, 50/21-1; Figs. 3.2, 3.14) and by log and cuttings data in well 49/13-4 (Fig. 3.8). In well 49/19-1, cuttings data indicate muds are overlain by sand and shell fragments (Fig. 3.10). Cuttings samples generally consisted of mud with or without secondary sand. Some washed samples consisted only of sand and the presence of mud was inferred from the log character. Trace components included shell fragments, mica and lignite.

In BGS-79/8, eight cores (6-13, Fig. 3.15) record an upper 30 m interval of brownish grey to dark grey, weakly laminated to massive mud. Silty laminae are noted in two upper cores (7-8), beneath a mottled silty clay (cores 6-7) suggestive of bioturbation of the top of the section. This is supported by foraminiferal evidence of marine conditions in core 6 (section 3.3.2).

3.3.1 c) lignite/shell abundances

Lignite and shell fragments are common in at least trace amounts in cuttings from both lower sand- and upper mud-dominated facies assemblages, along with mica and glauconite (see descriptions on figures - Table 3.1). In intervals of six wells, however, lignite and/or shell fragments are abundant as secondary (≤50%) or locally primary cuttings components. Visual estimates of lignite abundances are plotted against depth for wells 49/10-1, 10-2 and 19-1 (Figs. 3.4, 3.5, 3.10) and included in the cuttings descriptions of wells 49/24-3, 24-4 and 25-8 (Figs. 3.11-3.13). All but one of the (shelly) intervals occurs within the lower sand-dominated lithofacies assemblage.

In adjacent wells 49/10-1 and 10-2 (1.8 km apart, Fig. 3.1), lignite proportions up to 20% occur over the entire (150 m) intervals of sand-dominated assemblages (Figs. 3.4, 3.5). Lower and upper peaks in lignite abundance in both wells coincide with (lesser) peak abundances of shell fragments (not plotted on figures). In both wells, lower and upper peaks correspond with gamma ray log patterns correlative between the two wells (Fig. 3.6) - the lower peak coincides with log patterns interpreted as fining-up intervals of sand to mud (increasing values), whereas the upper peak lies within sandier sediments (lower values). Log patterns are therefore not a response to lignite abundances, which in any case are expected to provide a low gamma ray response (e.g. Rider, 1991).

Lignite and shell fragments are abundant in wells 49/19-1, 24-3 and 24-4, all located within five kilometres of each other (Fig. 3.1). In 49/19-1, lignite is estimated at up to 30% of cuttings over a 120 m sand-dominated interval, with two peaks in abundance spanning an upward transition from sand to mud-sand facies (Fig. 3.10). In addition, two intervals dominated by shell fragments occur, below and above the lignitic interval, the upper a sandy interval at the top of the mud-dominated facies (recognised from cuttings alone). In well 49/24-3, lignite dominates (≥50%) all samples in a lower sandy interval 100 m thick and occurs in lesser abundances with shell fragments in an overlying 100 m of sand-mud facies (Fig. 3.11). In well 49/24-4, lignite is abundant (<50%) and dominates some samples (>50%) in a lower interval of sand up to 100 m thick, overlain by sand-mud facies (Fig. 3.12).

In well 49/25-8, a single cuttings sample (30' interval) contained ≥25% lignite (Fig. 3.13).

Lignite and shell fragment abundances in cuttings may reflect concentration of low-density debris during transport and separation from circulating drilling mud, exacerbated by caving in sandy sediments. Gamma ray log patterns independent of plotted lignite variations (Figs. 3.4, 3.5, 3.10) also suggest actual abundances lower than 30-50%. On the other hand, the correlation of lignite peaks between adjacent wells shows that vertical variations in abundance may be real. The six wells thus indicate relatively greater abundances of lignite, and of shell fragments, within some valley fill intervals up to 200 m thick than in either overlying muds or underlying (incised) sediments. Lignite was not noted in cuttings from sediments below the valley base, but minor amounts of shell fragments occur in underlying muds from wells 49/10-1, 10-2, 13-1, 13-4 and 14-2 (Figs. 3.4-3.9).

The relative abundance of lignite and shell fragments in valley fills is to some extent corroborated by cores from BGS-79/8 (Fig. 3.15). Lignite is not observed in the upper muds, but pre-Pleistocene pollen abundances are interpreted to include disseminated lignite (section 3.3.3). The six to nine cores from lower sand-dominated sediments (#s 13-18/21), spanning 24-38 m of section, contain visible amounts of both lignite and shell fragments (I estimate up to 5%). Lignite is similarly abundant immediately below the valley base in cores from the Yarmouth Roads Fm (#s 18/21-28, 31/45 m of section -Fig. 3.15), but is not observed in underlying sandy formations.

The Yarmouth Roads Formation is the uppermost unit of the Pleistocene fluvio-deltaic sediments into which the tunnel-valleys of the area are incised (see Figs. 1.7, 1.8, 1.9b). It is up to 180 m thick in the study area and so is not represented in most of the exploration wells presented here. It has been sampled in a dozen BGS boreholes and vibrocores in the southern North Sea and is described as very fine to fine sand, with clay laminations, scattered pebbles (chert) and abundant 'wood clasts and plant debris' (lignite?), as well as local intercalations of shelly sand (Cameron et al., 1984; 1989a;

1992). The presence of shelly sand is also noted in the underlying (Pleistocene) Winterton Shoal and Smith's Knoll formations (see Fig. 1.8 for formations).

The Pleistocene sediments of the southern North Sea were described by Pegler (1995) based on cuttings from up to 50 exploration wells, mainly in the Dutch sector (northern K, L and M-blocks). He noted that lignite and shell fragments, as well as glauconite and mica, are common minor cuttings components in the upper sandier units. His descriptions suggest an overall upward increase in lignite and shell abundances from the Brielle Ground Formation (upper Pliocene) to the Yarmouth Roads Formation. A single concentration of shell fragments (≤40%) was described in one well (L8-1) within the Westkapelle Ground Formation (see Fig. 1.8), suggested to be a coquina. None of the other wells included notable concentrations of lignite or shell fragments within pre-glacial Pleistocene sediments. However, local abundances of lignite and shell fragments were noted in glacial sediments overlying the Yarmouth Roads Formation, including the Elsterian valley fill sediments of the Swarte Bank Formation (Pegler, 1995, p. 164).

Lignite and shell fragments are thus characteristic minor components of the pre-glacial Pleistocene sediments of the southern North Sea and lignite is a notable minor component of the upper Yarmouth Roads Formation. Both are also characteristic minor components of sand-dominated valley fills in the study area, in at least the low abundances observed in BGS 79/8 cores (up to 5%). Relative abundances in cuttings from six wells, notably of lignite, suggest concentrations far above those in the incised Pleistocene sediments.

Concentration of reworked lignite in glacigenic sediments is a recognised phenomenon in Germany, particularly in glaciofluvial sands where some layers have been mistaken for interglacial horizons (Johannsen, 1960; Ehlers and Grube, 1983). In some locations, thicknesses and concentrations are sufficiently high that the deposits have been economically exploited (Ehlers, personal communication). Lignite in Quaternary sediments is ultimately derived from Tertiary 'brown-coal' deposits which are widespread beneath northern Germany and the Netherlands (e.g. Zagwijn and Doppert, 1978).

3.3.2 Microfossil zones

Micropalaeontological data are presented from nine boreholes (Table 3.1). Exploration company reports of cuttings analyses for calcareous microfossils were available for three wells and I took cuttings subsamples from a further five for examination by B. Austin (personal communication, 1994-95). Unpublished BGS/RGD reports of analyses of core subsamples for foraminifera, pollen and dinoflagellates were obtained for BGS-79/8 and are presented here together for the first time.

Foraminiferal species identifications and numbers are summarised in tabular format for five wells (49/10-1, 13-1, 13-4, 19-1 and 24-3; Tables 3.5-3.9) and BGS 79/8 (Table 3.10). The remaining three wells yielded barren samples (B. Austin) the locations of which are shown on figures (49/24-4, 25-8, 50/21-1: Figs. 3.12-315). The number of species/individuals are plotted against depth for wells

49/10-1, 13-1 and 24-3 (Figs. 3.4, 3.7, 3.11). Foraminiferal numbers are also plotted for BGS-79/8 along with pollen zones and dinoflagellate abundances (Fig. 3.15).

Microfossil data show that non-fossiliferous sediments represent most of the valley fill, but are overlain by a marine zone within the upper part of the mud-dominated lithofacies assemblage. The two zones are described separately, followed by an evaluation of pollen and dinoflagellate content in BGS-79/8 cores.

3.3.2 a) lower non-fossiliferous

A non-fossiliferous character of both lower sand and overlying muds is recorded in cores from BGS-79/8 (Table. 3.10, Fig. 3.15), which yielded 0-5 benthic foraminiferal species (0-49 individuals) from 15 subsamples above the maximum probable valley base (cores 7-21, subsamples 2826-2841). The results do not help to identify the valley base in the borehole (see section 3.3.1a) as samples from sandy sediments in the possible range (cores 18-21) all contain similar assemblages (Table 3.10).

Low numbers of species (0-10) and individuals (0-20) also characterise most of the valley fill in well cuttings. This is shown by samples at regular 30-40' intervals in wells 49/10-1, 13-1, 24-3 (Tables 3.5, 3.6, 3.9) and 100-200' intervals in 49/13-4, 19-1 and 50/21-1 (Tables 3.7, 3.8 and Fig. 3.10). It is also suggested to be the case by individual samples from wells 49/24-4 and 25-8 (Figs. 3.12 and 3.13). Barren samples from the lower part of mud-dominated assemblages are noted in wells 49/13-4, 19-1 and 24-3 (Figs. 3.8, 3.10, 3.11).

Identifications of those benthonic and planktonic foraminiferal species present are available for wells 49/10-1, 13-1, 13-4, 24-3 and BGS-79/8 (Tables 3.5-3.7, 3.9, 3.10). Of interest are index species indicative of pre-Anglian/Elsterian or pre-Quaternary series within the North Sea Basin. The following derives mainly from contributions in Jenkins and Murray (1981). Pre-Anglian species include *Elphidiella hannai* (late Pliocene to early Pleistocene benthonic, common to glacial-stages: Funnel, 1981, King et al., 1981), which is observed in individual samples from wells 49/13-1 and 13-4 (Tables 3.6, 3.7) and in up to seven samples from BGS-79/8 (Table 3.10). An individual of *Cibicides subhaidingerii* (early Pleistocene benthonic, common to temperate stages - Funnel, 1981), occurs in one sample from 49/13-4 (Table 3.7). *Elphidium pseudolessoni* (early Pleistocene benthonic - Funnel, 1981) is present in up to three samples from BGS-79/8 (Table 3.10). Tertiary species are recorded in two wells. In 49/24-3 (Table 3.9) an individual of *Asterigerina guruchi* (early Oligocene benthonic - King et al., 1981) and in 49/13-1 (Table 3.6) planktonic foraminiferal species *Globoquadrina pozonensis* and *G. venezuelana* (genus range Eocene to Miocene - Loeblich and Tappan, 1964). A single occurrence of *Quinqueloculina carinata* (Triassic-late Eocene benthonic - Murray et al., 1981) in well 49/10-1 may be caved from an overlying marine assemblage (Table 3.5; see below). Cretaceous benthonic species (Hart et al., 1981) are identified in 49/13-1 (*Stensioina cf. pommerana*) and 49/13-4 (*Heterohelix* sp. and *Hedbergella* sp.). Finally, reworked Pliocene fossils

Table 3.5 - Calcareous microfossils identified in well 49/10-1 cuttings subsamples (Shell UK, unpublished biostratigraphical analysis). See Figure 3.4.

*DEPTH (ft K.B.) --------->	450	480	510	540	570	600	630	660	690	720	750	780	810	840	870	900	930	960	990	1020	1050	1080	1110	1140	1170	1200	1230	1290
SERIES	MID-PLEISTOCENE																			L. PLEISTOCENE					PLIOCENE			
STAGE (environment)	HOLSTEINIAN (marine)				ELSTERIAN (tunnel-valley fill)																							
OSTRACODA (unspecified)	X	o	o	o	o	o	•	•		•		\	\												•	•	•	
PLANKTONIC FORAMINIFERA:																												
Globigerinella sp.			•																									
Globorotalia sp.																					•					•		
BENTHONIC FORAMINIFERA:																												
Elphidium selseyense	X	X	X	X	o	\	\	o	o	\	•	\	o	•		\	•		\	o	o	o	o	\	o	o	o	o
Ammonia beccarii	\	\	\	\		•							•													\		
Bulimina marginata	\	\	\	•																								
Nonion sp.	\		\	\	\	\							\								•							
Quinqueloculina carinata (TR-u.Eoc)	o	\		\	\																							
Globocassidulina subglobosa		\	•	\																								
Elphidium sp.		\	\	\	\				•			•				•							•			\	\	•
Pyrgo alata		\						•																				
Buccella frigida		\	\	\	\		•				\													\			•	
Nonion compressum		\	\	\	o													\								•	\	\
Guttulina problema (TR-u.Eoc)			\																									
Bathybiphon spp.				•			•												\	\						•		
Astacolus spp.						•																						
Bulimina sp. (elongata?)								•																				
Nonion spp.												•									\		\		\			
Elphidium excavatum													\						•									
Elphidium spp.													\		\			\							\	•	•	
Eponides spp.																	•											
Cibicides spp.																		•						\				•
Cibicides refulgens																				•	\				•	•	•	
Elphidium sp. (or Cribrononion pseudolessonii)																								\				
Ammodiscus spp.																					•							
Elphidiella hannai																									•			
Nonion barleeanum inflatum																									\			\
Cibicides lobatulus var. grossa (u. Plio)																									\	\	\	\
Cassidulina leavigata pliocarinata																										•	\	o
OTHER FOSSILS: fish remains																									•			
FORAMINIFERA: # species	5	9	10	10	6	4	3	3	2	2	2	3	5	1	1	2	2	3	3	3	6	1	3	4	7	9	8	7
#individuals: minimum	33	37	37	37	20	6	4	8	7	3	3	4	13	1	2	3	2	5	5	9	13	6	9	8	16	17	17	20
maximum	135	140	137	137	60	12	7	22	21	6	6	7	36		5	6		11	11	26	33	20	26	21	42	40	43	57

* Bottom of nominal 10ft drilling interval; actual depths shallower due to drilling lag, and probable mixing of material from shallower depths.

KEY: • 1 | \ 2 to 5 | o 6 to 20 | X 21 to 100 — sp. = individual species (unspecified) spp. = more than one species

Table 3.6 - Calcareous microfossils identified in well 49/13-1 cuttings subsamples (Chevron, unpublished biostratigraphical report). See Fig. 3.7.

*DEPTH (ft K.B.) --------->	513	544	575	606	637	665	697	728	759	883	914	945	976	1007	1038
SERIES	MID-PLEISTOCENE									PLIO-PLEISTOCENE				EOCENE	
STAGE	ELSTERIAN														
(environment)	(tunnel-valley fill)														
OSTRACODA (unspecified)	+	+		+	+	+		+			+	+	+	+	
PLANKTONIC FORAMINIFERA															
Globoquadrina pozonensis (u.Eo-Mio)			+												
Globoquadrina venezuelana (u.Eo-Mio)					+	+	+	+	+		+	+			
Orbulina universa					+	+	+		+	+		+	+		
Globigerina falconensis					+	+			+			+	+		
Globigerinoides triloba triloba					+		+								
Globigerina bulloides												+	+	+	
Globigerinoides triloba sacculifera														+	
Globigerina linaperta linaperta														+	+
BENTHONIC FORAMINIFERA															
Elphidiella cf. arctica	+	+													
Rotalia beccarii	+	+	+	+		+	+		+		+				
Eponides frigidus	+														
Lenticulina calcar		+			+										
Nonion. sp.1		+													
Nonion depressulus			+		+	+	+			+	+	+	+		
Cassidulina laevigata			+						+	+	+	+	+		
Nonion granosum			+									+	+		
Robulus vulgaris				+	+		+	+		+					
Quinqueloculina undosa				+	+	+					+	+		+	
Triloculina sp.1					+					+		+			
Elphidiella hannai (u. Plio-l. Pleist.)					+		+	+		+	+				
Cibicides scaldisiensis					+	+			+	+	+	+	+	+	
Cibicides lobatulus						+				+	+	+	+	+	
Eponides repandus							+								
Eponides karsteni									+	+					+
Stensioina cf. pommerana (u. Cret.)									+						
Nonion elongatum										+		+			
Elphidium cf. pseudolessonii (lower Pleist.)										+					
Pyrgo cf. fornasinii											+				
Triloculina oblonga											+				
Cassidulina laevigata carinata											+	+			
Lagena lateralis											+				
Glandulina laevigata												+			
Polymorphina charlottensis												+			
Triloculina cf. circularis												+			
Dimorphina nodosaria												+			
Nonion commune												+			
Quinqueloculina seminulum												+	+		
Lagena costata												+	+	+	
Textularia sagittula												+			
Lenticulina peregrina												+			
Pullenia bulloides													+	+	
Globulina gibba													+		
Nonionella graciosa													+		
Anomalina grosserugosa														+	+
Quinqueloculina vulgaris														+	
Verneuilina triquetra															+
Lenticulina cultrata															+
Dentalina spinulosa															+
Cibicides refulgens															+
Lenticulina decorata															+
Gaudryina hiltermanni															+
Guttulina austriaca															+
BENTHONIC: # species	3	4	4	3	7	5	5	2	5	10	11	18	9	7	9
PLANKTONIC: # species	0	0	1	0	4	3	3	1	3	1	1	6	4	3	1
TOTAL FORAMINIFERA: # species	3	4	5	3	11	8	8	3	8	11	12	24	13	10	10

* Bottom of nominal 31 ft drilling interval (gap 759-852'); actual depths shallower due to drilling lag, and probable mixing of material from shallower depths.

KEY: + present

Table 3.7 - Benthonic foraminifera identified in well 49/13-4 cuttings subsamples (examined by B. Austin, written communication). See Figure 3.8.

*DEPTH (ft K.B.)-------->	400	600	800	1000
SERIES	MID-PLEISTOCENE			
STAGE	HOLSTEINIAN	ELSTERIAN		
(environment)	(marine)	(tunnel-valley fill)		
Elphidium excavatum (f. selseyense & clavata)	7			
Ammonia beccarii	6			
Elphidium asklundi	1			
Elphidium williamsoni	1			
Cibicides lobatulus	1			
Haynesina germanica	2			
Glandulina sp.	1			
Bolivina sp.	1			
pre-Quaternary sp. (inc. Heterohelix sp. Hedbergella sp., both Cretaceous)	13	2 to 4		
Stainforthia fusiformus			1	
Elphidiella hannai (u.Plio-l.Pleist.)			1	
Cibicides subhaidingerii (broken specimen)				1
# species	≥9	2 to 4	2	1
#individuals: minimum	33	2 to 4	2	1

* Bottom of nominal 20ft drilling interval; actual depths shallower due to drilling lag, and probable mixing of material from shallower depths.

Table 3.8 - Calcareous microfossils identified in well 49/19-1 cuttings subsamples (examined by B. Austin, written communication). See Figure 3.10.

*DEPTH (ft K.B.)-------->	410	600	780	930	1050
SERIES	MID-PLEISTOCENE				
STAGE	HOLSTEINIAN	ELSTERIAN			
(environment)	(marine)	(tunnel-valley fill)			
OSTRACODA (many species)	abundant				
BENTHONIC FORAMINIFERA					
Elphidium excavatum f. clavata	6				
Ammonia beccarii	4				
Nonion orbiculare	3				
Quinqueloculina seminulum	2				
pre-Quaternary	none				
BENTHONIC: # species	4	barren	barren	barren	barren
#individuals	15				

* Bottom of nominal 30ft drilling interval; actual depths shallower due to drilling lag, and probable mixing of material from shallower depths.

Table 3.9 - Foraminifera identified in well 49/24-3 cuttings subsamples (SHELL UK, unpublished biostratigraphical analysis). See Fig. 3.11.

*DEPTH (ft K.B.) ---------->	535	565	595	624	660	695	718	749	779	809	840	875	907	980	997	1027	1063	1087	1126	1150	1185	1216	1279	1341
SERIES	MID-PLEISTOCENE																				L. EOCENE			
STAGE	ELSTERIAN																							
(environment)	(tunnel-valley fill)																							
PLANKTONIC FORAMINIFERA																								
Globigerina primitiva																					\	\	o	o
Globigerina spp.																					\	o	o	\
Globigerinella pseudohastigerina sharkriverensis																					\	o	X	X
Globigerina parva																						•	\	
Globigerinella pseudohastigerina wilcoxensis																						o	o	o
Globigerinella pseudohastigerina micra																								\
BENTHONIC FORAMINIFERA																								
Nonion sp.	\																							
Alabamina wolterstorffi			•																			\		\
Cristellaria spp.				•				•													\	\	\	
Cibicides sp.							•														o	\		
Cibicides spp.								•															\	
Buccella frigida								•																
Asterigerina gurichi (l. Olig.)											•													
Nodosaria spinescens																				•	\	\	•	\
Textularia karreriella chilostoma																				•				
Alabamina wilcoxensis																					\			
arenaceous spp.																					o	\		
Bolivina cookei																					\		\	
Bulimina sp.																					•			
Cibicides sulzensis																					\	\	o	\
Cibicides sp.																					\	\	\	\
Clavulina parisiensis																					o	\		\
Cristellaria sp.																					o	o	\	.
C. lent. (robulus) cultrata																					\	\	\	•
Cristellaria sp.																					\	\		
Gyroidina soldanii var. girardana																					\	\		•
Gyroidina sp.																					\	\	\	\
Nodosaria longiscata																					o	o	\	\
Nodosaria sp.																					\	\	o	\
Nodosaria sp.																					•			
Nonion affine																					\			
Textularia smithvillensis																					\	•	\	\
Uvigerina batjesi																					\	\	\	
Vaginulina lent. (vaginulinopsis) decorata																					•		\	
Bulimina turrilina brevispira																						\		\
Bulimina sp.																						o	X	o
Ceratobulimina sp.																						\		
Dentalina terquemi																						\	\	\
Epistomina elegans																						\		
Nodosaria spinulosa																						•		
Angulogerina sp.																							•	
Cibicides sp.																							\	
Cibicides sp.																							\	\
Pullenia bulloides																							\	
Textularia spiroplectammina carinata																							\	\
Globulina gibba																								•
Quinqueloculina carinata (TR-u. Eoc.)																								•
Vulvulina spiroplectammina excolata																								
FORAMINIFERA: # species	1	0	1	1	0	0	1	3	0	0	1	0	0	0	0	0	0	0	0	2	25	28	26	23
# individuals: minimum	2		1	1			1	3			1									2	67	77	108	73
maximum	5																				188	218	387	239

*** Bottom of nominal 20-40 ft drilling intervals; actual depths shallower due to drilling lag, and probable mixing of material from shallower depths.**

KEY: • 1 | \ 2 to 5 | o 6 to 20 | X 21 to 100 | sp. = individual species | spp. = more than one species

Table 3.10 - Benthonic foraminifera identified in BGS-79/08 core subsamples (D. Gregory, 1980 examination - BGS Internal Report PDL 80/36). See Table 3.4 and Fig. 3.15.

SUBSAMPLE #	2820	2821	2822	2823	2824	2825	2826	2827	2828	2829	2830	2831	2832	2833	2834	2835	2836	2837	2838	2839	2840	2841	2842	2843	2844	2845	2846	2847	2848	2849	2850	2851	2852	2853	2854	2855	2856	2857
SERIES	HOLOCENE			U. PLEIST.		MID-PLEISTOCENE																	MID(?) - LOWER PLEISTOCENE															
STAGE				EEMIAN		HOL.	ELSTERIAN							(tunnel-valley fill)					?																		
GENUS & SPECIES																																							
Ammonia batavus	150	150	120	132	60																									2						1		
Elphidium selseyense	90	60	24	13	12																																		
Elphidium clavatum	17	25	120	132	180	60	4	2					1						1	1	15	1		4	24	10	22	40	255	5	132	21	123	99	24	69	21	27
Quinqueloculina seminulum	27	14	17	8	6																																		1
Quinqueloculina cliarensis	2	1	5	2																																			
Cibicides lobatulus	8	13		16	10														2					5	5	3	4	1	3						1		37
Elphidiella hannai (u.Plio-l.Pleist.)	6	4					1						2		25				24	35	20	16	10	1	19	62	9	10	21	42	2	1	2	2			2	12
Bulimina marginata	2	3	3	1	2																																
Elphidium williamsoni	3	10	1	8	1																																
Miliolinella subrotunda	6	9		5	6																							1				1			2		
Nonion depressulum	3	6	3	1	5																																
Elphidium exoticum	1	3	2																1																				
Elphidium albiumbilicatum	1																																				
Planorbulina sp.	1																																				
Polymorphinid	1					1																			2	3	2	1			2					1		
Pyrgo williamsoni	1		1																													1	1		2	4	
Cassidulina obtusa		1		1	6	4																														1		7
Elphidium bartletti		1																															1		7			
Uvigerina peregrina		3																																			
Lagena clavata		1	2																																		
Trifarina angulosa		1			1																						2									2
Virgulina schreibersiana			1																																		
Bolivina sp.			1		2																																
Buccella vicksburgensis			1																									1	1	1		5	3		6		3	
Buccella frigida				1	2	8		2											4	5									2				1		1	4	
Fissurina laevigata				2																																	
Fissurina lucida				4	3																			1								2				
Protoelphidium anglicum				4																																	
Elphidium macellum					1																												1				
Oolina squamosa					1																				1											
Patelina corrugata					1																																
Protoelphidium orbiculare						156															2				4	7		2	2		9	35	45	25	1		5	1
Elphidium asklundi						39																			2			1			25	7	6				
Elphidium magellanicum						1																															
Rosalina sp.						1																															
Cassidulina laevigata/obtusa								2																												180	
Cibicides scaldisiensis								2																													
Elphidium pseudolessoni																			3	7	1			17	1	5	13	11	6	3	1		1	1	1		4
Bulimina elongata																						1			1												16
Eponides repandus																									2	6	2	1	5	13	4	1							
Nonion barleeanum																													1	1	1	5	38	40	1	7	15	7
Quinqueloculina sp.																													1									
Elphidium subarcticum																														1	2			1				3
Trifarina fluens																														1									
Cassidulina teretis																															96	207	84	111	228	195	13	13
Cibicides lobatulus var. grossa (u.Plio.)																															12	2	4	6	2		4	7
Islandiella helenae																															3	4	7	6	25	15	3	6
Sigmoilina celata																															2							
Dentalina frobisherensis																																	1		2			
Dentalina baggi																																			1			
Lagena laevis																																				1		
Elphidium selseyense/clavatum																																						180
Lenticulina sp.																																						1
Oolina sp.																																						1
BENTHONICS: # species	16	17	14	15	17	8	2	4	0	0	0	0	2	0	1	0	0	0	3	5	5	4	1	3	8	10	6	9	11	11	12	10	13	14	11	13	11	17
#individuals:	319	305	301	330	299	270	5	8	0	0	0	0	3	0	25	0	0	0	26	45	49	19	10	7	78	94	42	78	308	68	265	306	317	302	298	297	254	325

are reported on the composite log of well 49/24-4 within the valley fill below 550' K.B., although analytic results could not be obtained from Shell UK.

The lower valley fill sediments thus contain foraminiferal evidence for non-marine conditions and reworking of older (Cretaceous to early Pleistocene) sediments.

3.3.2 b) upper marine

Samples containing foraminiferal assemblages indicative of temperate marine conditions are recognised within the upper part of mud-dominated sediments in three wells, 49/10-1, 13-4 and 19-1 (Tables 3.5, 3.7, 3.8; Figs. 3.4, 3.8, 3.10). A single core subsample from BGS-79/8 records transitional marine conditions at the top of the valley fill section (Table 3.10, Fig. 3.15).

In well 49/10-1, five samples (at 30' intervals) contain up to 10 species, dominated by *Elphidium selseyense* with consistent secondary numbers of *Ammonia beccarii* and *Bulimina marginata* (Table 3.5; Fig. 3.4). Ostracods are also abundant but no species identifications were made. Two Tertiary species occur, *Quinqueloculina carinata* and *Guttulina problema* (both Triassic to late Eocene - Murray et al., 1981). The marine fauna appear to represent the entire upper mud-dominated assemblage at the well location (Fig. 3.4) although this could be due to downward contamination from upper fossiliferous muds during drilling.

Marine fauna are recognised in single samples from wells 49/13-4 and 19-1, based on examination of small numbers of individuals (B. Austin, personal communication). In 49/13-4, 33 individuals in a sample from upper sand-mud facies are dominated by *E. selseyense* and *A. Beccarii* and include pre-Quaternary species (Table 3.7, Fig. 3.8). In 49/19-1, upper mud-dominated sediments contained 15 individuals dominated by *Elphidium clavatum* and *A. beccarii*, as well as an abundant and diverse ostracod fauna (Table 3.8, Fig. 3.10).

E. selseyense (Heron-Allen and Earland) and *E. clavatum* (Cushman) are variant forms of *E. excavatum* (Terquem). Feyling-Hanssen (1972) distinguished *E. excavatum* f. *clavata* and f. *selseyensis* and related them to arctic (cold) vs boreal (temperate) environments, respectively. *E. selseyense* is common in shallower parts of the North Sea today and dominates both Holocene and last-interglacial assemblages in central-northern North Sea cores (Sejrup and Knudsen, 1993). *Ammonia beccarii* (Linné) is typical of temperate (interglacial) stages in the North Sea basin, including the present, but is almost unknown from cold (glacial) stages (Funnel, 1981). In marine sediments from the upper parts of Elsterian tunnel-valleys in northern Germany, assemblages of *A. beccarii* and *E. excavatum* dominate many samples, arctic forms of the latter giving way upwards to boreal forms (Knudsen, 1993). The two species indicate interglacial conditions and suggest shallow water depths (<25 m).

In BGS-79/8 an unusual marine fauna occurs in subsample 2825 (8 species, 270 individuals; Table 3.10) from the top of the valley fill (Fig. 3.15). The assemblage consists almost exclusively of three species, *Protoelphidium orbiculare, E. clavatum and E. asklundi. P. orbiculare* is particularly characteristic of transitional cold to temperate conditions in the upper Pliocene to mid-Pleistocene of the North Sea Basin (King et al., 1981). Gregory (1980) cited a report of a similar modern fauna by Lagoe (1979), from the shallow (<6 m) fringe of the Arctic Ocean where fast ice freezes on or into the seabed during the winter. The sample may record a transition from glacial to interglacial marine conditions. The sample corresponds to mottled clay at the top of the valley section suggestive of bioturbation (section 3.3.1b).

The upper marine zone thus records interglacial deposition of muds and muddy sands, with possible evidence of the transition from preceding non-marine glacial environments in BGS-79/8. Pre-Quaternary species in the marine assemblages record reworking of older sediments.

3.3.2 c) BGS-79/8 pollen and dinoflagellates

Analytical results from core subsamples are summarised in Fig. 3.15. Dinoflagellate cysts are absent or present in low to intermediate numbers throughout the upper part of the borehole such that no interpretive comment is warranted (Harland, 1980). A complete pollen analysis was presented by de Jong (1981) who recognised a number of pollen groups and zones summarised in Fig. 3.15. The location of the upper borehole within a tunnel-valley was not recognised at the time of his analysis, nor were the foraminiferal results available.

Pollen spectra have been applied to the differentiation of glacial and interglacial stages within the Quaternary sediments of the North Sea Basin (e.g. Zagwijn, 1975). Pollen stratigraphy is based on the identification of warm versus cold floras but is complicated by mixing during transport, especially in marine settings, and by common reworking of palynomorphs characterised by remarkable longevity. Abundant pre-Quaternary palynomorphs in fluvial successions have been related to reworking by glacial erosion, but reworking may also occur in marine interglacial environments as observed in Holocene sediments of the southern North Sea (Zagwijn and Veenstra, 1966). Glacial erosion may also result in reworking of interglacial flora, which in any case could resemble an impoverished Tertiary flora (de Jong, 1981). Interpretation of pollen spectra therefore depends on the presence of contrasts and the stratigraphical context.

De Jong (1981) proposed the distinction of five pollen zones in BGS-79/8, interpreted in terms of glacial and interglacial climates (Fig. 3.15). Zones 2-4 are now seen to correspond to valley fill sediments, zone 1 to lower Pleistocene sediments. Several samples with numbers too low to interpret occur in the lower valley fill and 'zone' 2 is based on a single sample bracketed by such lacunae. Zones 2 and 4 each contain high numbers of herbaceous pollen and pre-glacial (early Pleistocene and pre-Quaternary) palynomorphs, assemblages suggestive of (fluvio-) glacial influence. Zone 3 shows low values for herbaceous and pre-glacial pollen, distinct values for interglacial tree pollen (e.g. Abies, Picea, Quercus) but no thermophilous taxa, for which de Jong posited an interstadial or late interglacial. The valley fill is overlain by sediments with unambiguous interglacial spectra, consistent with foraminiferal indications of marine conditions (Fig. 3.15).

Zones 2-3 (inclusive of barren samples) correspond to lower sandy valley fill, while zone 4 corresponds to the upper mud (Fig. 3.15). The contrast between zones 3 and 4 may thus reflect sedimentary processes. I suggest that the stratigraphical context of zone 3 - within sandy valley fill, bracketed by zones dominated by reworking - makes an interstadial interpretation untenable. No change in pollen content is noted in association with the transition to marine interglacial conditions recorded by foraminifera in sample 2825 (upper zone 4b).

I conclude that the valley fill sediments contain a mixed pollen assemblage broadly indicative of glacial influence, that is of reworking of previously deposited sediments.

3.3.3 Evidence of reworking

Sand- and mud-dominated lithofacies assemblages contain foraminiferal and palynological evidence of reworking of pre-glacial Pleistocene and Tertiary sediments. Reworking of such sediments also accounts for the presence of lignite, shell fragments and locally glauconite within valley fills.

Pre-Anglian foraminiferal species are recognised within the valley fill in cuttings from four wells and in cores from BGS 79/8 (section 3.3.2a). Examples include species whose range is limited to the lower Pleistocene in the North Sea Basin, as well as longer-ranging and/or older species (Triassic to Tertiary). Samples come from both the lower sand- and upper mud-dominated assemblages.

Pollen assemblages from valley fill sands and muds in BGS-79/8 cores include groups identified as early Pleistocene through to pre-Miocene, the latter up to 50% within upper muds in zone 4b (Fig. 3.15). Other groups may also be reworked, including the 'interglacial' assemblages. Reworked Tertiary pollen is common within Elsterian glaciolacustrine sediments of the northern Netherlands, which are correlative to the upper valley fill muds of this study (Peelo Formation; see section 3.5.1b). The pollen is argued to represent disseminated lignite (van Gijzel, 1963).

Lignite, as well as shell fragments, are ubiquitous minor components of cuttings and core samples. Relative abundances are noted in six wells over sand-dominated intervals up to 200 m thick (section 3.3.1c). Lignite and shell fragments also occur in low abundance in offshore core and well data from lower Pleistocene sediments and lignite appears to be most abundant within the lower to mid-Pleistocene Yarmouth Roads Formation (BGS-79/8, Fig. 3.15; Pegler, 1995).

In well 49/13-4, cuttings from a lower 60 m interval of sand-dominated valley fill sediments contain up to 5% glauconite, above glauconitic sands (up to 50%, Fig. 3.8) within the incised sediments. The depth of the glauconitic sands indicates that they correspond at least in part to the lower Pleistocene Ijmuiden Ground Formation (see Figs. 1.8, 1.9b), previously described as glauconitic (Cameron et al., 1984; Pegler, 1995). The sediments may record proximal reworking of glauconite within the lower fill, although glauconite occurs throughout the Plio-Pleistocene succession and in abundance at depth (Pegler, 1995).

Foraminifera and palynomorphs thus record erosion of sediments as young as early Pleistocene, the same age as those into which the tunnel-valleys are incised. The fluvio-deltaic sequence into which the tunnel-valleys are incised also represents a likely source for the lignite, shell fragments and glauconite within valley fills. The valley fill sediments are consistent with a local derivation.

3.4 CORRELATION TO SEISMIC AND REGIONAL STRATIGRAPHY

Correlation of downhole data to the seismic stratigraphy of Chapter 2 is shown on individual well-diagrams. Depth-to-base velocities were derived from identification of the valley base on both well and seismic data, in some cases supported by sonic log integrations to TWT within the fill (Table 3.11). Log-defined lithofacies changes within the fill correspond with contoured reflecting surfaces in some cases and are correlative between two adjacent wells. However, lithofacies assemblages do not correspond with seismic sequences, the boundaries of which lie within the lower sand- and upper mud-dominated assemblages. This is shown in summary on Fig. 3.16 and schematically in relation to the regional stratigraphy on Fig. 3.17.

Sequence III approximates the upper (marine) foraminiferal zone, and is correlated with the Egmond Ground Formation of regional mapping. Underlying valley fill sediments of the Swarte Bank Formation are seen to comprise three intervals: lower sands (clinoform surfaces, sequence I), middle sands (onlapping surfaces, lower sequence II) and upper muds (onlapping surfaces, upper sequence II).

I explain the basis of well-to-seismic correlation (3.4.1), and the correlation of well data to the seismic stratigraphy (3.4.2), and the regional stratigraphy (3.4.3).

3.4.1 Depth versus TWT

Correlation was effected using depth-to-base velocities calculated from identification of the basal unconformity on both seismic data and within the thirteen boreholes. Sonic log calibrations within the valley fill were also available from intervals of seven wells (Table 3.11). Both methods yielded velocity estimates comparable throughout the study area.

3.4.1 a) depth-to-base velocities

Average velocities to the valley base were calculated for each well (Table 3.11). Velocities from sea level average 1.8 ± 0.1 km/s. Velocities from seabed differ by ≤0.1 km/s, reflecting the shallow water depths (30-40 m) above most of the study area. A value of 1.8 km/s was assumed in correlating well to seismic data within the valley fill in the absence of sonic log calibrations. This value was also useful in identifying the valley base in some wells with poor log contrast (e.g. Figs. 3.7, 3.9).

The precision of the estimated velocities depends on the identification of the valley base on both well and seismic data (Table 3.11). Basal definition in boreholes is estimated to range from ±1-10 m, depending on data quality and formation contrast. Contoured reflection surface depths depended on several factors difficult to quantify, such as vertical resolution (up to 15 ms), measurement accuracy (±0.5 mm, or ±5 ms), wavelet calibration to sea level during processing and the data density available for contouring. In Table 3.11, precisions of either ±5 ms or ±10 ms are ascribed to seismic measurements, except three based on sonic log calibrations accurate to ±1-2 ms. These categories of

Table 3.11 - Valley fill velocities from downhole sonic log integrations, and depth-to-base velocities from correlation to seismic (see Figures).

					CORRELATION TO SEISMIC				
	SEA-	SONIC INTEGRATION			VALLEY BASE		VELOCITY		
BORE-HOLE	BED (m)	interval (m)	interval (ms TW)	velocity (km/s)	well (m bsl)	seismic (ms TWT)	to sealevel (km/s)	to seabed (km/s)	Figure
49/2-2	36				183 ± 2	200 ± 5	*1.83 ±.07*	2.01 ± .07	**3.2**
49/9-1	38				242 ± 3	270 ± 5	*1.79 ±.08*	1.92 ± .08	**3.3**
49/10-1	37	37	43	*1.72*	269 ± 1	303 ± 2	*1.78 ±.03*	1.87 ± .03	**3.4**
		(232-269)							**to**
49/10-2	37				290 ± 1	325 ± 5	*1.78 ±.06*	1.88 ± .06	**3.6**
49/13-1	29				205 ± 2	230 ± 5	*1.78 ±.07*	1.89 ± .07	**3.7**
49/13-4	28				269 ± 1	300 ± 5	*1.79 ±.06*	1.87 ± .06	**3.8**
49/14-2	29				130 ± 2	145 ± 5	*1.79 ±.07*	1.99 ± .07	**3.9**
49/19-1	31				310 ± 1	345 ± 5	*1.80 ±.06*	1.87 ± .06	**3.10**
49/24-3	31	113	132	*1.71*	312 ± 1	350 ± 5	*1.78 ±.06*	1.85 ± .06	**3.11**
		(125-238)							
49/24-4	33	136	154	*1.77*	233 ± 1	274 ± 1	*1.70 ±.02*	1.78 ± .02	**3.12**
		(97-233)							
49/25-8	30				275 ± 10	300 ± 10	*1.8 ±.2*	1.9 ± .2	**3.13**
50/21-1	31	179	184	*1.95*	506 ± 1	560 ± 1	*1.81 ±.02*	1.85 ± .02	**3.14**
		(327-506)							
79/08	21.5				100 ± 9	120 ± 10	*1.7 ±.2*	1.8 ± .2	**3.15**
			AVERAGE:	***1.79***		AVERAGE:	***1.8 ±1***	1.9 ±.1	

precision have also been applied to reflector depths plotted on well-diagrams, indicated by arrows. However, this extension of uncertainties resulting from identification of the valley base neglects the effects of vertical velocity variations within the fill.

3.4.1 b) sonic log velocities

Sonic velocity data, integrated to obtain TWT and calibrated to downhole check-shots, are available from open-hole intervals of four wells (49/10-1, 24-3, 24-4 and 50/21-1) and shown on individual well diagrams (Table 3.11).

Sonic log character indicates poor borehole conditions (caving) in all the wells except 50/21-1. The erratic transit times in the other three wells resulted in integrated velocities that are erroneously low. Log header information indicates that well 49/24-4 had a 9.5 μs/ft (32 m/s) correction applied above 1682' K.B. (483 m bsl) following integration, in order to give a reasonable velocity to seabed. Wells 49/10-1 and 24-3 were uncorrected, and have lower interval velocities (Table 3.11). Interval velocities for all seven wells average 1.79 km/s, comparable to the 1.8±1 km/s depth-to-base velocity from sea level, but less than the 1.9±1 km/s from seabed (Table 3.11).

The interval velocity average would be lower still but for well 50/21-1, which yielded good quality sonic data (Fig. 3.14) from the lower 179 m of a deep valley fill (327-506 m bsl). Velocity over this interval averages 1.95 km/s. An upward decrease in velocity is noted from up to 2.1 km/s at the valley base to 1.9 km/s near the top of the interval. Depth-to-base velocities from seabed and sea level remain within the average range (Table 3.11).

3.4.2 Seismic stratigraphy

Contoured seismic reflecting surfaces within the valley fill are plotted on figures (see Table 3.11), with arrows indicating estimated precision (±5 or ±10 ms, as above). Sequence boundaries are also shown and correspond in some cases to contoured surfaces and in other cases were inferred from reflector geometries (section 2.4). The sequence boundaries are shown on the summary diagram (Fig. 3.16).

Reflection surfaces coincide with log-defined lithofacies changes in a number of wells, mainly within the lower sand-dominated assemblage. Well to well correlation is possible in one case. However, sequence boundaries (I/II and II/III) do not correspond to the lithofacies assemblages, but lie within the lower and upper sand- and mud-dominated assemblages (Fig. 3.16).

I discuss the relation of reflectors to lithofacies changes, and sequences to lithofacies assemblages.

3.4.2 a) reflectors versus lithofacies changes

Correspondence between contoured reflecting surfaces and log-defined lithofacies changes is recognised in seven wells (49/9-1, 10-1, 10-2, 13-4, 19-1, 24-4 and 50/21-1; Table 3.11). Reflection surfaces correspond, in different instances, with sharp boundaries between different facies, the top/bottom of inferred fining/coarsening-up intervals, or gamma ray peaks indicative of mud interbeds.

Some wells did not penetrate any contoured surfaces. Seismic surfaces were traced and contoured only to the extent that data quality and density allowed and not on the basis of well data, so that correspondences may be more common than recognised. However, examples of contoured surfaces which do not correspond with lithofacies changes also occur in both sand- and mud-dominated lithofacies assemblages (e.g.wells 49/9-1, 10-1, 13-1, 13-4; Table 3.11).

In adjacent wells 49/10-1 and 10-2 (1.8 km apart, Fig. 3.6), lithofacies changes are associated with three reflectors that can be correlated from well to well. The lower two reflectors approximate the base/top of inferred fining-up intervals, while the upper reflector corresponds to a gamma ray peak in each well. The lower reflectors are clinoform surfaces within sequence I, the upper reflector a subhorizontal surface within lower sequence II.

In well 49/13-4 close correspondence was obtained between two contoured surfaces and a 25 m sand interval, bounded above and below by mud, for which reflector configurations define a gently downlapping (≤1°) sand lobe at the base of sequence II (Fig. 3.8). In the same well, an underlying reflector (230 ms) corresponds to a mud interbed (gamma ray peak) within sands. However, no lithofacies changes are observed in association with a lower reflector in sands (265 ms) and two upper reflectors within muds (120/140 ms).

These examples show that reflectors traced within the valley fill correspond in some cases to verifiable physical changes in the sediments, correlative over several kilometres.

3.4.3 b) sequence characterisation

The relation of seismic sequences I-III to the lithofacies assemblages and foraminiferal zones is shown on Fig 3.16 and schematically in Fig. 3.17. Sequence boundaries I/II and II/III lie within the lower sand- and upper mud-dominated lithofacies assemblages, respectively. Sequence I thus corresponds to sandy sediments, sequence II includes lower sands and upper muds and sequence III approximates the upper (marine) muds. Here I briefly characterise each of the three seismic sequences.

<u>Sequence I</u> (clinoform surfaces) lies everywhere within the lower sand-dominated sediments. The sands include concentrations of cuttings components (lignite, shell fragments) and some reworked microfossils. Log data indicate both sharp and progressive lithofacies changes, such as mud interbeds and large-scale (10-50 m) fining-up or in some cases coarsening-up intervals. Clinoform reflectors in part correspond to lithofacies boundaries, correlative between wells 49/10-1 and 10-2 (Fig. 3.6). The sequence dominates valley fills with seismically defined thicknesses of up to 400 m near well 50/21-1 and is observed in the well with thickness up to 370 m (Figs. 3.14, 3.16).

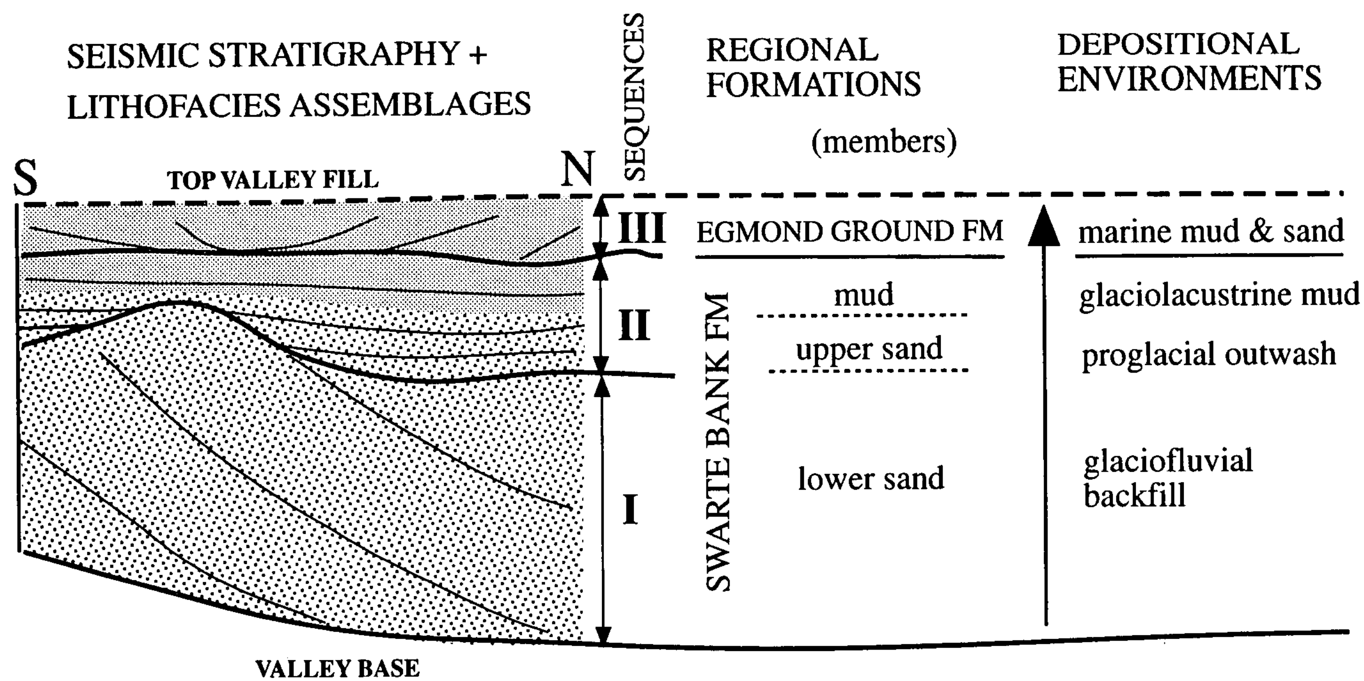


Figure 3.17 - Schematic illustration of the relationships between seismic sequences and lithofacies assemblages, and correlation to regional stratigraphy. The Swarte Bank Formation (valley fill) is recognised to have three members, based on the seismic stratigraphical division of lower and upper sand. The vertical succession records increasingly distal glacial depositional environments, prior to interglacial marine conditions of the Egmond Ground Formation.

Sequence II (onlapping surfaces) brackets the upper part of the sand-dominated sediments and the lower part of the mud-dominated sediments. The sequence fills axial basins at the surface of sequence I with seismically defined thicknesses of up to 125 m in the UK sector (see Fig. 2.11) and is observed in wells with thicknesses up to at least 80 m (Fig. 3.16). The lower sands (up to 50 m thick in wells) are undistinguished from unconformably underlying (clinoform) sands by log or cuttings character (e.g. 49/10-1, 19-1; Figs. 3.4, 3.10). In well 49/13-4, the lower sands correspond to two reflectors defining a gently downlapping lobe (Fig. 3.8). Elsewhere the log-defined boundary between the sand- and mud-dominated sediments within the sequence does not correspond to a prominent reflector. At several well locations, the lower sands are seen to correspond to reflectors onlapping axial basins at the surface of sequence I, while at least the upper part of the muds correspond to more extensive parts of the sequence (e.g. 49/10-1, 19-1, 24-4; Figs. 3.4, 3.10, 3.11).

Sequence III (complex surfaces) contains the fossiliferous marine zone within the upper mud-dominated lithofacies (Fig. 3.16). It is up to 100 m thick on seismic profiles within axial basins incised into underlying sequence II and up to 40 m of its lower part is observed in three wells (Fig. 3.16). The prominent reflector traced as the base of the sequence does not correspond to a recognisable lithofacies change. No reflectors were traced within the sequence. Upper sandy intervals are indicated in wells 49/13-4 and 19-1 (Fig. 3.16) and could be more common above the upper limit of well data.

3.4.3 Regional stratigraphy

The above results show that sequence III contains interglacial marine sediments and I here correlate it with the Egmond Ground Formation (Holsteinian stage). Sequences I and II correspond to the valley fill sediments proper of the Swarte Bank Formation (Elsterian stage) within which I recognise three members: lower sands, upper sands and muds (Fig. 3.17).

3.4.3 a) Egmond Ground Formation (Holsteinian)

Oele (1971b) proposed that shelly marine sands identified in Dutch blocks K17 and P5 were representative of the Holsteinian interglacial stage. Subsequent coring and seismic profiling suggested such sediments form an extensive tabular unit, mapped as the Egmond Ground Formation (Laban et al., 1984; BGS/RGD, 1984; 1986). The formation was not extended into the UK sector due to lack of sample control, but notes on the Indefatigable Sheet (BGS/RGD, 1986) indicated that it might be present within the upper parts of some Elsterian buried valleys the fill of which was mapped as the Swarte Bank Formation. The probable marine character of an upper seismic facies of the Swarte Bank Formation up to 30 m thick was also noted by Balson and Cameron (1985), Cameron et al. (1987), Long et al. (1988) and Balson and Jeffery (1991), who suggested that reflector configurations were indicative of deltaic influxes from the west and of relatively high-energy environments, such as mud flats and sandy channel-fills.

Cameron et al. (1989b, 1992) attributed the suspected marine sediments at the top of the Swarte Bank Formation to the Egmond Ground Formation. The latter authors described it as a tabular sheet with thicknesses of 8-20 m across much of the Indefatigable Sheet, locally thicker in the tops of Elsterian tunnel-valleys and resting with angular unconformity on underlying formations. They suggested it to record high energy, open marine conditions of the later Holsteinian, in contrast to the Sand Hole Formation identified to the west (Spurn Sheet - BGS, 1991) and interpreted to record a restricted, temperate environment of the early Holsteinian. The Sand Hole Formation was identified from marine muds within borehole 81/52A, beneath non-fossiliferous sands attributed to the Egmond Ground Formation, and stated to be 'entirely restricted' to a 25 km radius around borehole 81/52A on the basis of an interval of closely spaced, subparallel reflectors on seismic profiles. It was not considered that the Sand Hole Formation might be represented beneath the Egmond Ground Formation within the upper parts of the tunnel-valleys to the east.

The upper valley fill seismic facies corresponds to sequence III of this study (see Fig. 2.22), the marine character of which is confirmed by foraminiferal data from wells 49/10-1, 13-4 and 19-1 (Table 3.1). The sequence occupies basins up to 70 m in relief incised into underlying sediments and an upper part up to 25 m thick extends between some adjacent tunnel-valleys. The erosional base, and the complex reflector configurations, are suggestive of relatively high-energy marine environments. However, the lithological character of at least the lower 40 m of the up to 100 m thick sequence is mainly muddy (Fig. 3.16). The muddy character and the location within 'restricted' basins invites a correlation to the Swallow Hole Formation, despite the differing seismic character. Overlying sandy sediments are indicated in two wells (49/13-4, 19-1) and could be more prevalent at shallower depths as an upper layer that might correspond to the Egmond Ground Formation. However, this is not demonstrable from the well data, nor is it obvious that in such a setting of variable relief, low- and high-energy environments will necessarily succeed one another rather than co-exist.

There is thus little basis for a distinction between lower and upper formations within the tunnel-valleys. I follow Cameron et al. (1992) in referring marine sediments at the top of Elsterian valley fills (sequence III) to the Egmond Ground Formation. However, they may span the entire Holsteinian stage, not only a later part, and so be in part correlative to the Sand Hole Formation.

3.4.4 b) Swarte Bank Formation (Elsterian)

Oele (1971b) recognised 'fluvioglacial clays' in borings from the Dutch sector, beneath Holsteinian marine sands, and correlated them with the Peelo Formation (Elsterian stage) of the northern Netherlands. Shallow boreholes (<100 m) and seismic profiles subsequently showed such glaciolacustrine muds to be widespread across the Dutch sector in the upper parts of, and between, buried valleys (Laban et al., 1984; see Laban, 1995). These sediments and the underlying fill of the buried valleys were entirely mapped as the Swarte Bank Formation in the study area (BGS/RGD, 1986) and seismic facies within the valleys were interpreted to indicate that the upper muds overlie sediments of different character (Laban et al., 1984; Cameron et al., 1987; 1989b; Long et al., 1988;

Balson and Jeffery, 1991). The unsampled lower valley fill (chaotic seismic facies) has been suggested to comprise relatively coarse-grained sediments including subglacial tills, re-sedimented diamictons, slumped beds or gravelly coarse sands.

Balson and Jeffery (1991, Figure 163) and Wingfield (1990, Table 2) suggested that the Swarte Bank Formation was represented within BGS borehole 79/8 by 30 m of glacigenic mud. My interpretation of the borehole is that the valley fill includes the muds and 25-38 m of underlying sandy sediments (Fig. 3.15). Data from twelve other boreholes show that the Swarte Bank Formation is characterised by up to 350 m of sand-dominated lithofacies assemblages beneath up to 90 m of mud-dominated assemblages (Fig. 3.16).

The results appear broadly consistent with seismic predictions of muds over coarser sediments, although the dominance of the latter is contrary to expectations. This is in part because the mud- and sand-dominated assemblages do not correspond to the seismic facies or sequences (Fig. 3.16). The sand-dominated sediments include both sequence I (chaotic facies) and the lower part of sequence II (layered facies), while the mud-dominated sediments lie within upper sequence II. In other words, the seismic sequences (facies) do not correspond to distinctive lithologies - the lower (clinoform) sequence lies within sand-dominated sediments, while the upper (onlapping) sequence includes sands and overlying muds (Fig. 3.17).

It is thus possible to distinguish three distinctive depositional intervals, or members, within the Swarte Bank formation on the basis of combined seismic and lithological character. These are from top to bottom:

- muds (onlapping reflecting surfaces)
- upper sands (onlapping reflecting surfaces)
- lower sands (clinoform reflecting surfaces)

These depositional intervals are shown schematically in Fig. 3.17 and have also been included on individual well diagrams where possible. Their interpretation is considered in the next section, by way of comparison of the Swarte Bank Formation with the fill of other Elsterian tunnel-valleys.

3.5 Depositional Environments

The Egmond Ground Formation has been shown above to represent marine environments of the Holsteinian interglacial. Here I address the depositional environments of the Swarte Bank Formation, by comparison with other Elsterian tunnel-valleys in northern Europe. I am mainly concerned to show that the valley fill represents a succession of waterlain glacigenic sediments. Depositional processes within glacial basins will be considered in Chapter 4.

I first show that the vertical succession of sediments within the Swarte Bank Formation is closely comparable to the fills of other buried Elsterian tunnel-valleys in northern Germany, the Netherlands and East Anglia (Fig. 3.18). Borehole studies in all three areas have proven successions of waterlain sands capped by muds, both locally derived, interpreted to record a transition from glaciofluvial to lacustrine environments.

I then use the combination of borehole and seismic data available in the southern North Sea to refine this basic depositional model. The three intervals of the Swarte Bank Formation are proposed to record northwards progradation of glaciofluvial sands (backfill), followed by proglacial outwash of sands and proximal to distal accumulation of lacustrine muds (Fig. 3.17). The division of the lower sands is suggested to apply to other tunnel-valleys.

I begin by comparing previous interpretations of valley fills (3.5.1), then describe the three proposed depositional intervals (3.5.2).

3.5.1 Previous Interpretations

The Swarte Bank Formation is correlative to the fill of other buried tunnel-valleys interpreted to have formed beneath the Elsterian ice-margin in the northern European lowlands. Comparisons have been made with the valleys of East Anglia (Ehlers and Gibbard, 1991), the Netherlands (Laban et al., 1984; Cameron et al., 1987; 1989b) and northern Germany (Ehlers and Linke, 1989). The sediments filling these other valleys have in most cases been examined only by boreholes, with limited geophysical examination in East Anglia. Only in the Netherlands have they been given formation status (Peelo Formation, below).

The Elsterian stratigraphy of northwestern Europe from the Netherlands to Poland was reviewed by Ehlers et al. (1984). A series of buried tunnel-valleys 100-400 m in relief all contain similar vertical successions of waterlain sands overlain by muds, interpreted to record a transition from glaciofluvial to lacustrine sedimentation (Fig. 3.18). The valley fill sediments are everywhere overlain by Holsteinian marine, estuarine or lacustrine sediments. A similar sequence and interpretation has been proposed for buried valleys up to 100 m in relief beneath East Anglia (Woodland, 1970). The generalised vertical sequence is directly comparable to that of the southern North Sea (Fig. 3.18).

I review information on fill character in a) northern Germany, b) the Netherlands and c) East Anglia, followed by d) comparison with the Swarte Bank Formation.

3.5.1 a) northern Germany

Regional borehole studies of Elsterian tunnel-valleys were presented by Kuster and Meyer (1979) in Lower Saxony, Grube (1979) in Hamburg, Hinsch (1979) in Schleswig-Holstein, and Eissmann and Müller (1979) in eastern Germany. English-language reviews have been provided by Ehlers (1981), Grube (1983) and Ehlers et al. (1984). Most studies were based on a combination of drillers' records and geophysical logs (see section 3.2.4). Detailed studies of valleys beneath Hamburg, including a cored borehole, were presented by Ehlers and Linke (1989).

Valley fills are described as consisting of basal gravelly sands and overlying mainly fine sands (glaciofluvial facies) overlain by variably sandy muds of the Lauenburg Clay (glaciolacustrine facies). The Lauenburg Clay is a distinctive unit beneath northern Germany, over 150 m thick at the top of some valleys; sands and dropstones are common in lower layers. The underlying sands contain fine

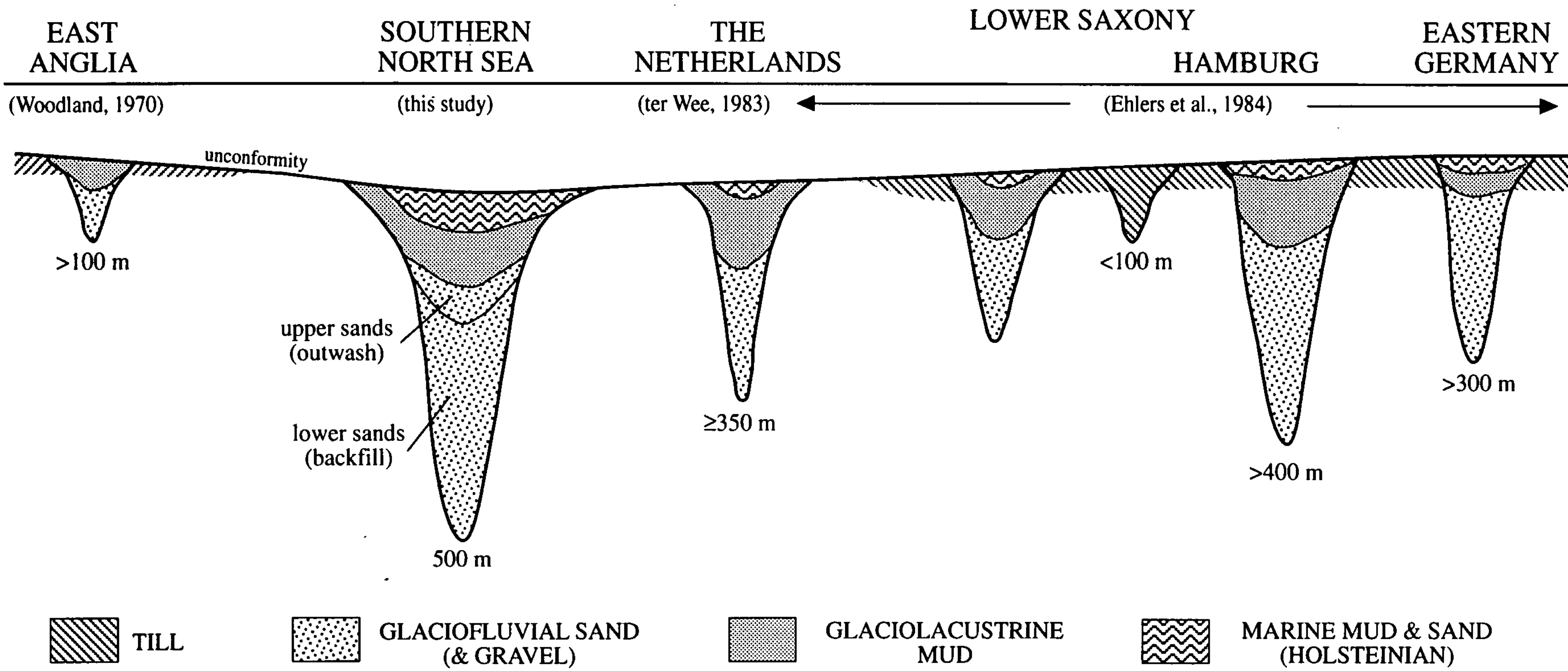


Figure 3.18 - Schematic stratigraphy of Elsterian tunnel-valleys across northwestern Europe, simplifed from Ehlers et al. (1984) with the addition of East Anglia and results from this study. The valleys appear to be incised into till, although Woodland (1970) argued that till and glaciofluvial sands were deposited concurrently. The seismic stratigraphical division of sands into lower backfill and upper proglacial outwash in the southern North Sea may apply to other tunnel-valleys. In Germany, meltwater sands locally include till at the base or on the flanks of the valleys, and rafts of Tertiary sediments (see Ehlers and Linke, 1989).

and coarse gravels (4-60 mm) of Paleozoic to Cretaceous age, a strong argument for the northern and so non-fluvial origin of the sediments (Kuster and Meyer, 1979). 'Local' gravels are also present, derived from the Tertiary sediments into which the valleys are incised: coarse gravel analyses include sandstones, mudstones and lignite. Rafts of Tertiary strata occur locally within the sands (Ehlers, 1981; Ehlers and Linke, 1989). Sediments interpreted as till are rare within the larger valleys (>100 m) but are recognised in places on the flanks and also as the complete fill of smaller (<100 m) valleys (see Fig. 3.18). A cored borehole (qho 4) in one valley (base 260 m bsl) showed a lower 15 m of gravelly sand (≤10% gravel, upper gravel layer <1 m thick) interpreted as interbedded meltwater sand and sandy till (Ehlers and Linke, 1989).

Ehlers and Linke (1989) presented results of 26 detailed well interpretations and correlations, along with a cored borehole, from deep (up to 400 m) valleys beneath Hamburg. A 35 km axial profile (17 wells) along one valley showed that uniform sands up to 280 m thick in the north grade southwards into a thinning wedge of interbedded sand, silt and clay layers. The interbedding is recognised over 30 km, and dips gently (≤1°) south; a tendency to fining up intervals is noted. A cross-profile shows that the beds form gently concave surfaces and include significant lateral facies variability (sand to silt transitions). The interbedded succession implies pulses of sedimentation from a source to the north, interpreted as ice-marginal meltwater discharges.

3.5.1 b) the Netherlands (Peelo Formation)

The Peelo Formation has been described by de Jong (1967), van Staalduinen et al. (1979), Zandstra (1983) and ter Wee (1983). The Peelo Formation includes a lower channel-fill (glaciofluvial) facies, within valleys up to 350 m deep, and an upper mainly muddy (glaciolacustrine) facies over 100 m thick at the tops of some valleys and extensive between valleys throughout the northern Netherlands (Fig. 3.18). The channel-fill facies is described as consisting of alternating fine and coarse sands with local mud layers. The upper facies is well known as potklei (pottery clay), a compact gray to black locally laminated mud with variable proportions of interlaminated silty fine sand. Sandy tills have been suggested to occur in two boreholes from the lower channel fill (Zandstra, 1983) but this interpretation has been questioned (ter Wee, 1983). Ter Wee (1983) noted that the lower channel-fill facies constitutes the entire valley fill in some wells in the south and that the upper potklei, where it extends beyond the valleys in the south, includes large areas of mainly sand (alternating coarse/fine) (also van Staalduinen et al., 1979).

The Peelo Formation is characterised by the incorporation of reworked Pleistocene, Tertiary and Cretaceous sediments, indicated by pollen and foraminiferal content (Zagwijn, 1974). Reworked Miocene pollen within the potklei is argued to derive from disaggregated lignite (van Gijzel, 1963). Heavy mineral assemblages within the potklei also resemble Miocene assemblages (Zandstra, 1983). Fine gravel analyses suggest the local occurrence of Scandinavian pebbles within the lower channel-fill and in the potklei at the type locality (Zandstra, 1983) but on the whole the sediments are comparable to the Pleistocene fluvial sediments of southern origin into which they are incised. Ter

Wee (1983) argued that this implied a southern source for the valley fills. However, Pleistocene and Tertiary sediments extend for hundreds of kilometres to the north of The Netherlands.

3.5.1 c) East Anglia

The small dimensions of the buried tunnel-valleys of East Anglia (<1km wide, ≤100 m deep) are presumed to reflect their incision into relatively resistant chalk (Boswell, 1914; Woodland, 1970). Regional borehole evidence regarding the composition of their fill, based almost entirely on drillers' logs, was used by Woodland (1970) to present a model of glaciofluvial to lacustrine sedimentation during deglaciation. A number of subsequent geophysical and borehole studies of small areas have not attempted to test his model, but have provided information consistent with his observations.

Woodland noted that drillers' records presented problems of interpretation, both as to consistency and as to the meaning of terms such as 'boulder clay'; he argued that mud-dominated sediments within valley fills were waterlain. He generalised two main fill facies: lower sands and gravels, and upper silts, clays and loams (i.e. mud). Both included fragments of chalk and (derived) flint, and the dark colour of the muds was related to a high comminuted chalk content. He did not recognise till within the valleys, although it was extensive between them. Instead he argued that the coincident distribution of tunnel-valleys and regional till sheets implied that both had been formed penecontemporaneously. Sands and gravels dominated the fill in a downstream (south, east) direction while overlying mud thickened upstream. The former he related to channel-filling by subglacial streams and the latter to lacustrine deposition as ice withdrew to the north.

Subsequent geophysical studies of some buried valleys, mainly using gravity and electrical surveys complemented by boreholes, have shown variations in valley fill composition related to the varying dominance of sand- versus mud-dominated sediments (Clarke and Cornwell, 1983; Barker and Harker, 1984; Cornwell and Carruthers, 1986). The latter two papers contained the use of 'boulder clay' in reference to mud-dominated sediments, although it was accepted that borehole data did not justify either the textural or genetic implications. Barker and Harker (1984) suggested that variations in fill composition within the Stour buried valley were consistent with Woodland's interpretation of a downstream increase in sand and gravel.

Cox (1985) presented additional borehole data, used to construct cross-sections across three buried valleys. He did not discuss the methods of drilling or interpretation but suggested that the valley fills included tills and in one borehole pre-glacial fluvial gravels. Ehlers and Gibbard (1991) noted that his cross-profiles show that the majority of fill sediments are gravels, sands and silts of meltwater origin, consistent with Woodland's original observations.

3.5.1 d) comparison with Swarte Bank Formation

To summarise, borehole studies of buried Elsterian tunnel-valleys from Germany to East Anglia provide evidence for a vertical succession of lower sand-dominated to upper mud-dominated sediments, the former representing the bulk of valley fills (Fig. 3.18). Diamicts interpreted as tills may be present locally. Both facies contain lithological indications of local derivation, from sediments similar to those into which they are incised, further demonstrated by pollen and foraminifera as young as early Pleistocene in age from The Netherlands. The vertical successions are interpreted to record glaciofluvial to lacustrine sedimentation.

This generalised depositional sequence is directly comparable to that of the Swarte Bank Formation, as characterised in section 3.2. The lacustrine character of upper muds in the Dutch sector was recognised by Oele (1971b), who compared them directly to the potklei of the northern Netherlands. A lacustrine character is also consistent with the non-fossiliferous character I identify in cuttings and cores. The waterlain character of underlying sands is evident only from BGS-79/8 cores (Figs. 3.15) but well data effectively demonstrate the sand-dominated and non-fossiliferous character of the majority of the sediments (Fig. 3.16). Also, both lower sands and upper muds of the Swarte Bank Formation contain indications of reworking from Cenozoic sediments, consistent with indications of local derivation in all other northern European valleys.

The results from other valleys, in particular geophysical logs and cores from northern Germany (Ehlers and Linke, 1989), support an interpretation of the lower sand-dominated lithofacies assemblage of the Swarte Bank Formation as mainly glaciofluvial in origin. Local occurrences of muddy facies interlayered with sands or overlying the valley base (e.g. 49/10-1, 10-2 - Fig. 3.6; 49/13-1 - Fig. 3.7) may be indicative of tills. Equally, they could record interbedding of glaciofluvial sand and lacustrine mud as in some German wells.

I conclude that regional studies concur regarding the waterlain origin of the large-scale fining-up sequences which characterise the fill of buried Elsterian tunnel-valleys in northern Europe, including those of the present study. The interpreted transition from glaciofluvial to lacustrine sedimentation is applied to the Swarte Bank Formation below.

3.5.2 Proposed Depositional Sequence

The regional distinction of glaciofluvial versus lacustrine sedimentation within tunnel-valleys has been based on a simple textural distinction of coarse- and fine- grained waterlain sediments, implicitly during a transition from proximal to distal glacial environments. I present a simple refinement of this basic depositional model based on the seismic distinction of lower and upper sands within the Swarte Bank Formation (Fig. 3.17). The three members of the formation (section 3.4.3b) are interpreted to record northwards glaciofluvial backfill (clinoform sands), proglacial outwash (onlapping sands), and proximal to distal lacustrine mud.

The three depositional units are briefly considered in turn, from top to bottom.

3.5.3 a) lacustrine mud (proximal to distal)

Non-fossiliferous muds of upper sequence II (Fig. 3.17) have been correlated to the potklei of the Netherlands (Oele, 1971b), which in turn are correlative to the Lauenburg Clay of northern Germany (Ehlers et al., 1984). These sediments record lacustrine sedimentation in glacial lakes prior to inundation by the interglacial Holstein Sea. In both areas, the 'clays' include vertical and lateral variations in sand content related to transitions from proximal (ice-influenced) to distal lacustrine environments (ter Wee, 1983; Ehlers and Linke, 1989). In northern Germany, the Lauenburg Clay in borehole qho 4 includes weak bedding which, if annual, indicates 2400 years of deposition (Wüstenhagen, 1984). The lacustrine muds of the Swarte Bank Formation may also record a long period of sedimentation during the transition from ice-proximal to distal environments.

3.5.3 b) proglacial outwash

Sand-dominated sediments conformably underlie muds within the lower part of sequence II and unconformably overlie clinoform sands of sequence I (Fig. 3.17). The sands correspond to reflectors which fill basins at the surface of sequence I. The subhorizontal, locally downlapping (Fig. 3.8) configuration of the sands records deposition beyond the ice-margin in lacustrine basins.

This interpretation is similar to that proposed by Ehlers and Linke (1989) to account for gently foreset interbeds of sand and mud within upper tunnel-valley fills beneath Hamburg. A similar interpretation may apply to southern sandy facies of the potklei in the Netherlands (ter Wee, 1983).

3.5.3 a) glaciofluvial backfill

The clinoform reflecting surfaces of sequence I correspond to sand-dominated sediments and represent the bulk of the Swarte Bank Formation (Fig. 3.17). Clinoform surfaces in part correspond to recognisable lithofacies changes within the sandy sediments (Fig. 3.6) and imply northwards progradation of glaciofluvial sediment. This is in the direction of glacial withdrawal and so is referred to as backfill. This is discussed in terms of subglacial sediment supply in Chapter 4.

Clinoform sands dominate all tunnel-valleys in the study area. By extension, I propose that equivalent sediments dominate the lower sandy fill of Elsterian tunnel-valleys across northern Europe. Beneath Hamburg, such sediments could correspond to the lower uniform sands up to 280 m thick (Ehlers and Linke, 1989). Woodland (1970) proposed that lower sands and gravels in East Anglian tunnel-valleys record deposition from subglacial streams during ice recession.

3.6 SUMMARY

Geophysical log and cuttings data from 13 exploration wells and cores from one BGS borehole allow the first complete vertical characterisation of lithology and microfossil content within buried tunnel-valleys in the North Sea. Valley fills consist mainly of sand-dominated sediments, overlain by mud-dominated sediments which contain an upper marine interglacial foraminiferal zone. Sands and muds both include interbedding and fining- or coarsening-up intervals and appear waterlain in core samples. Both contain foraminiferal, palynological and component (lignite, shell fragments, glauconite) indications of reworking from Cretaceous to Lower Pleistocene sediments. These results are closely comparable to those from other Elsterian tunnel-valleys in northern Europe, interpreted to record a vertical transition from glaciofluvial to lacustrine (to marine) deposition during deglaciation.

Correlation to the seismic stratigraphy shows that reflector packages (sequences) do not correspond to the lithofacies assemblages. Sequence I lies within the sand-dominated lithofacies, sequence II brackets the transition from sands to muds and sequence III corresponds to the marine foraminiferal zone. The latter is identified as the Egmond Ground Formation (Holsteinian interglacial). Underlying valley fill sediments proper of the Swarte Bank Formation include three members: lower (clinoform) sands, upper (onlapping) sands and onlapping muds. These are interpreted as glaciofluvial backfill, proglacial outwash and proximal to distal lacustrine muds. They are discussed in an ice marginal context in the next chapter.

CHAPTER 4
DEGLACIAL FORMATION

4.1 INTRODUCTION

The purposes of this chapter are:

- to propose a stratigraphical model for the formation of the tunnel-valleys by erosion and redeposition beneath the ice sheet margin
- to apply a depositional model to the progradation of the sedimentary fill
- to discuss the results in terms of en- and subglacial drainage processes during deglaciation.

In Chapter 2, seismic reflection evidence was presented for erosion to the south, deposition to the north and erosional overlap by younger cut and fill elements to the north, a morpho-stratigraphical architecture consistent with drainage to an ice margin receding to the north. The tunnel-valley basins also increase in size and spacing to the east in parallel with substrate thickness. In Chapter 3, borehole evidence was presented for a vertical succession of sands to muds interpreted as glaciofluvial backfill, proglacial outwash and lacustrine to marine deposits. The sediments contain evidence of derivation from sedimentary strata as young as early Pleistocene.

In this chapter, I propose that the tunnel-valleys formed by the headward excavation and backfilling of basins beneath the receding ice margin. This resulted in an axially diachronous stratigraphy, younging to the north, in which the vertical succession records the lateral migration of contemporaneous erosional and (proximal to distal) depositional environments. Basin dimensions indicate reworking of Cenozoic sediments beneath the outer 50 km of the ice sheet. Deposition may have lasted several millenia prior to the marine transgression with glaciofluvial sands deposited much more rapidly than lacustrine muds.

Fill progradation took place in the outer 5-20 km of the basins by the up-slope movement of water and sediment along backset surfaces. Backsets are natural expressions of glaciofluvial sedimentation along an ice margin, observed or inferred within esker ridges. I apply an eskerine depositional model to infer a distributed system of subglacial streams feeding subaqueous marginal fans within the larger tunnel-valley basins. The aggradation of sheet-like deposits by lateral and longitudinal fan migration would have been promoted by grounding-line migration during ice margin fluctuations. Erosional surfaces, tills and glaciotectonic rafts may also be intercalated within the backset glaciofluvial sediments.

Tunnel-valleys are argued to form in response to two factors: englacial recharge of surface water to the basal drainage system and deformation of the sediment bed. Surface ablation increases during deglaciation to dominate discharges by several orders of magnitude and tunnel-valley formation during this extended 'deglacial flood' provides a basis for rejecting alternative catastrophic models. Models for the formation of eskers during deglaciation indicate surface recharge across a sub-marginal zone tens of kilometres wide and eskers and tunnel-valleys are suggested to be equivalent and concurrent features. The tunnel-valley erosional record is influenced by substrate characteristics: the arborescent geometry, size and spacing are consistent with downward and headward tunnel growth in response to high englacial discharges to a stably deforming subglacial aquifer.

I begin by applying the model of migrating basins to the axial stratigraphy (section 4.2), discuss backfill processes along a receding ice margin (4.3) and ice sheet drainage during deglaciation (4.4).

4.2 MIGRATING BASINS

The model of elongate basins axially 'migrating' during ice margin recession is schematically illustrated in Fig. 4.1 - headward excavation (in response to drainage to the south) is accompanied by backfilling (by progradation to the north). The stratigraphical consequence is illustrated in Fig. 4.2 - an axially diachronous valley base and fill, in which the vertical succession records the northwards migration of contemporaneous erosional to depositional environments.

Headward excavation (and so basin migration) is a logical inference to account for the valley backfill. The model allows an interpretation of the prograding and onlapping elements of the axial stratigraphy in relation to time. Basal and fill characteristics are used in turn to estimate the dimensions of the active sub-marginal zone.

I discuss the inference of headward excavation (4.2.1), the axially diachronous fill record (4.2.2) and the estimated basin dimensions (4.2.3).

4.2.1 Headward Excavation

The inference of headward excavation is consistent with the need for a glaciofluvial sediment source during fill progradation and with the geometry of sub-marginal basins. The logic of this is illustrated in the 'forward' context of ice margin recession (Fig. 4.1).

Assume the downward excavation of some initial elongate basin: erosion is accompanied by sediment supply to the ice margin. Now allow recession of the ice margin, displacing the loci of erosion and deposition: headward excavation results in sediment supply to the distal portion of the basin. In effect, an 'active' basin migrates with the ice margin, displacing sediment from up-glacier to marginal locations.

'Headward' excavation is not exclusive of downward erosion. The two are naturally interrelated, in the same way that lateral progradation involves upward accumulation (Fig. 4.1). The arborescent plan form of the tunnel-valleys is compatible with the downward excavation of an initial horizontal drainage network, which also expanded headward with ice margin recession. The evolving form of such a drainage system is not readily derived from the basal unconformity, which represents the end product of such erosion. That is to say, the spatial distribution of vertical and horizontal components of erosion over time is conjectural.

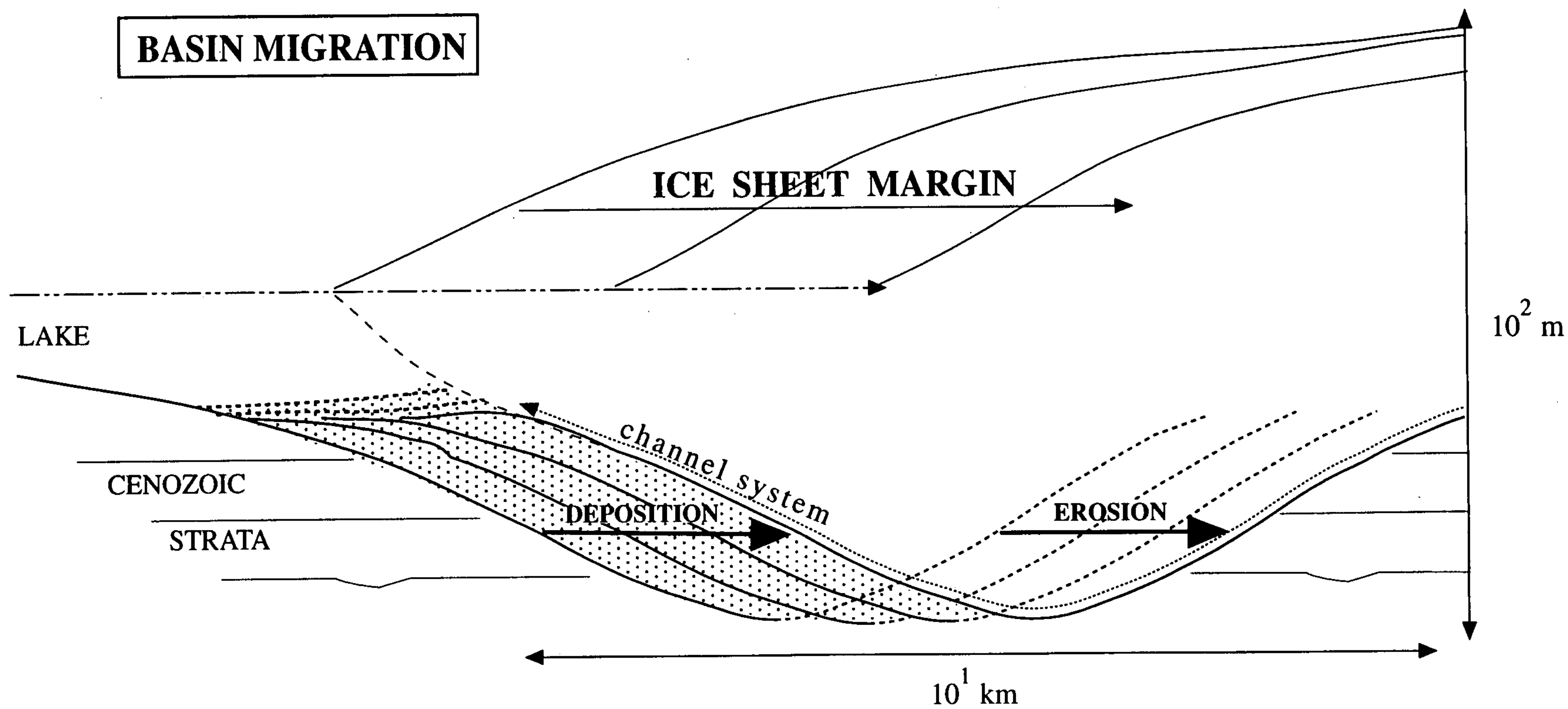


Figure 4.1 - Schematic illustration of proposed model of contemporaneous headward erosion and backfilling during ice margin recession.

The presence of axial sub-basins within the tunnel-valleys nonetheless implies temporal variability in some balance of vertical versus horizontal erosion, i.e. the size (or shape) of the 'active' basin varied during migration. This may have been influenced by the rate of ice margin recession: a decreased rate promoting downward erosion (and a corresponding accumulation of sediment at the ice margin), an increased rate favouring the opposite tendencies. However, axial relief was not observed to be correlative between adjacent tunnel-valleys, suggesting additional factors. Basin size and shape might also have varied in response to changes in discharge, or to the nature of the materials being eroded and transported.

The model requires the excavation of an initial basin, which represents a net loss of material to the system. Additional material would have been lost by long-distance transport of finer-grained material (water must leave the system, by moving away from or along the ice margin). These losses account for the incomplete fill of the tunnel-valleys. The expression of the net loss of material will depend on the size of the initial basin and the extent of its migration (no migration simply excavates a hole) and the composition of the material subsequently eroded (complete retention of sands, minimal retention of muds).

The net loss of material may be distributed across lesser basins at the surface of the backfill, or recorded in a single basin at the point where migration ceases. The former situation might correspond to the *Rinnenseen* of the Baltic lowlands, the latter to some of the large bathymetric deeps of the North Sea (which should therefore coincide with the headward terminations of tunnel-valleys). In any case, it is not necessary to invoke mechanisms of 'dead' or buried ice (e.g. Woldstedt, 1952; Kozarski, 1966/67; Wingfield, 1990) to explain the irregular axial relief of tunnel-valleys.

4.2.2 Axially Diachronous Fill

The model allows an interpretation of the chronostratigraphy of the axially prograding and onlapping fill (Fig. 4.2) that accommodates the observations of previous chapters. The duration of erosion and deposition are estimated as several millenia by comparison with other areas.

4.2.2 a) chronostratigraphy

Sequences I and II correspond to a vertical succession of sands to muds (beneath marine sediments of sequence III). These record the lateral migration of three contemporaneous depositional environments: glaciofluvial backfill, proglacial outwash and lacustrine deposition. The result is a series of diachronous intervals which include approximately isochronous surfaces (Fig. 4.2).

Prograding surfaces (sequence I) record successive stages of glaciofluvial deposition during basin backfilling (section 4.3) and are inferred to be subhorizontal in time. In contrast, their discordant bounding relationships are diachronous, the result of erosion below (valley base) and deposition above (I/II boundary). Proglacial outwash of sands was contemporaneous with glaciofluvial fill progradation (Fig. 4.1), so that backfill isochrons extend across the sequence boundary in continuity with onlapping surfaces (Fig. 4.2). However, the interval of outwash sands (lower sequence II) is

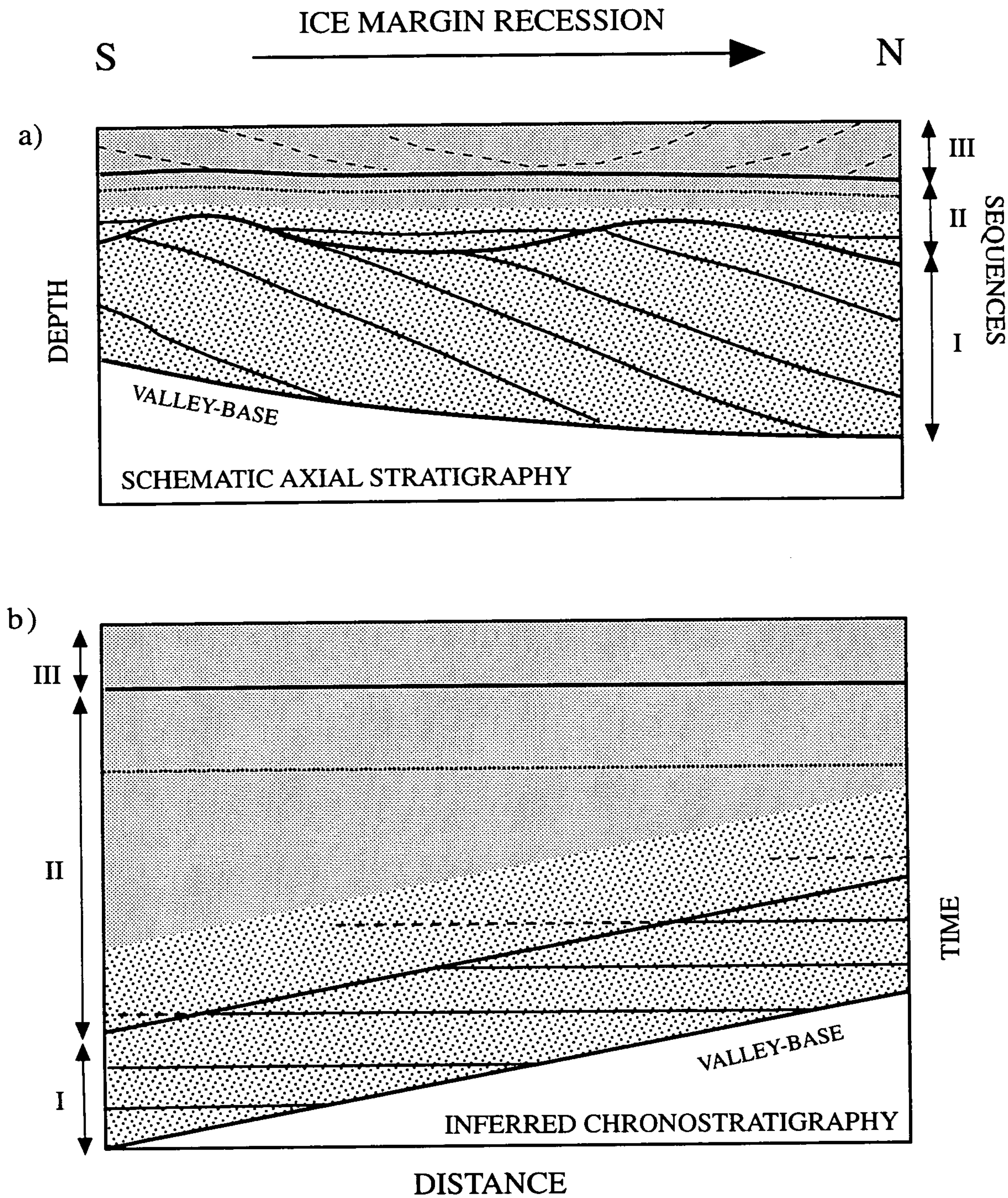


Figure 4.2 - Schematic a) axial stratigraphy and b) chronostratigraphy. Sequence I clinoform surfaces (backfill) approximate isochronous surfaces, bounded by diachronous sequence boundaries (see Fig. 4.1). Sequence II onlapping surfaces (proglacial outwash) are subhorizontal in space, but contemporaneous with subglacial (backset) sediment sources to the north and lacustrine muds to the south. The upper muds record deposition after the ice sheet has withdrawn from contact with the lake basins and the marine transgression (base sequence III) is effectively an isochronous event.

diachronous within successive axial basins at the surface of the backfill, progressively exposed within the lengthening proglacial lake basins.

Axial relief at the surface of the backfill records variable sediment delivery to the ice margin. The basins and sills were shown to be correlative between some adjacent tunnel-valleys (Fig. 2.27) and the lines of correlation invite an interpretation as ice margin lines. However, as with erosion, spatial variability in deposition could be due either to temporal variations in the rate of recession (thicker deposits along stable margins), or independent variations in the rate of sediment supply (due to discharge variations or the spatial distribution of erosion). Thus the lines of correlation may parallel ice margins, but it is not possible to infer whether the sills record more time than the basins (i.e. stable ice margins). The sequence boundary is therefore shown as linear in time on Fig. 4.2.

The concordant boundary between outwash sands and lacustrine muds (within sequence II) must be diachronous, as the sands were deposited proximal to the receding ice margin and would at all times have been accompanied by distal lacustrine deposition. However, the lacustrine muds are expected to be decreasingly diachronous with time as the uppermost interval will record deposition throughout the lake basins subsequent to the withdrawl of ice from the southern North Sea. The interglacial marine inundation of the basins is assumed to approximate an isochronous event (Fig. 4.2).

4.2.2 b) duration

The duration of erosion and deposition can be estimated from recession rates of the last deglaciation (Fig. 4.3) and from lacustrine sediments in other Elsterian tunnel-valleys.

Reconstructed rates of recession of the last northern hemisphere ice sheets are in the range of 10-250 m/a in the outer 500 km of deglaciation (Boulton et al., 1985; Dyke and Prest, 1987). However, rates of less than 50 m/a are only associated with the outermost ice limits, prior to more rapid recession of the margin across lowland areas (Fig. 4.3). Application of rates of 50-250 m/a to the 100 km latitudinal extent of the study area suggests between 400-2000 years of recession of the Elsterian ice margin.

In northern Germany, proximal to distal lacustrine accumulation of 115 m of Lauenburg clay in the upper part of Elsterian tunnel-valleys is estimated to have lasted 2400 years, on the assumption that stratification in a borehole (qho 4) represents seasonal accumulation (Wüstenhagen, 1984; Ehlers and Linke, 1989).

The duration of ice margin recession across the area and the subsequent period of lacustrine accumulation may thus have been comparable. However, at a given point in the tunnel-valley fill the amount of time represented by glaciofluvial sands (backfill and contemporaneous proglacial outwash) must be much less than that represented by overlying muds (shown schematically in Fig. 4.2). This can be shown using clinoform surface extents of 5-20 km and the range of recession rates above. Backfill across a distance of 20 km could have taken place in 80-400 years and across 5 km in 20-100 years. The sands are also much thicker than the muds so that accumulation rates of the former may have been several orders of magnitude higher than of the latter.

4.2.3 Basin Dimensions

Basal and fill characteristics can be used to estimate the dimensions of the migrating basins i.e. of the active sub-marginal zone of contemporaneous erosion and deposition. Backfill surface extents and the morphology of the valley base suggest dimensions of up to 50 km.

Backfill took place along prograding surfaces 3-20 km in length. Taken as a half-length of the migrating basins (Fig. 4.1) this implies erosion and deposition over distances of 6-40 km. This represents a minimum estimate as longer basins could have supplied sediment to a relatively narrow sub-marginal zone.

However, the valley base exhibits significant basinal closure over the latitudinal extent of the study area (100 km) and length scales of 6-40 km embrace a range of axial sub-basins. The largest single axial sub-basin in the study area is 45 km long (see Fig. 2.15). Such length scales also accommodate the offset of valley segments at angles, previously suggested to record erosional overlap of younger elements during ice margin recession (section 2.5).

As noted, the valley base records the end product of erosion so that the above estimates in effect apply to the final size attained by the basins. The smaller features along which the larger basins developed could have been much more extensive. However, the substantial erosion would have been accomplished proximal to the site of deposition. I therefore infer the tunnel-valleys to primarily record reworking beneath the outer 50 km (or less) of the former ice sheet margin.

The proposed dimensions are broadly consistent with downhole evidence for the composition of fill sands and muds which contain lithological (lignite, shell fragments) and micropalaeontological components derived from Cenozoic sedimentary strata as young as early Pleistocene in age. Such strata correspond to the Tertiary and Pleistocene succession into which the tunnel-valleys are incised.

4.3 BACKFILL PROCESSES

Backfilling of basins along a receding ice margin is a depositional style that has received limited attention in the literature. However, glaciofluvial depositional models include inferences of 'axial back-lapping' or 'shingling'. I apply an eskerine depositional model of 'longitudinally overlapping' subaqueous outwash to an interpretation of the prograding surfaces in the tunnel-valleys as backsets. This implies a distributed system of subglacial channels within the outer 5-20 km of the basins, debouching at the ice margin in a series of grounding-line fans.

I critically review previous work on backsets (section 4.3.1), and apply it to the tunnel-valley record (4.3.2)

4.3.1 Backsets

Backset surfaces have been observed within glaciofluvial sediments and inferred to exist as necessary elements of accumulation along ice margins. The progradation or shingling of such surfaces is implicit in models for the formation of eskers and related ridges during deglaciation. Such models derive from de Geer (1897, 1912) and more recently from the subaqueous outwash model of Rust and Romanelli (1975).

I review a) previous observations and inferences of backset surfaces, and b) the model of subaqueous outwash.

4.3.1 a) observations and inferences

The term and concept of a backset was introduced by Davis (1890), who identified such surfaces within glaciofluvial deposits of New England, as natural counterparts to the proglacial deltaic accumulation recorded by top- and foresets (Fig. 4.4). He observed them as bedding planes within cross-bedded sands, the direction of cross-bedding indicative of deposition on surfaces ascending up to 20° (his Figure 3, reproduced as Figure 3B of Jopling, 1975). Davis reasoned:

> "If the feeding streams came from beneath the ice they must needs rise to flow over the delta surface to its front, and hence the backward growth at the head must have been at such points in the form of up-hill deposition." (p. 197)

He immediately applied his own concept to the head of a glacial valley flood-plain (sandur) referring to "...the back-set beds, which theory leads me to suppose must exist..." (p. 201).

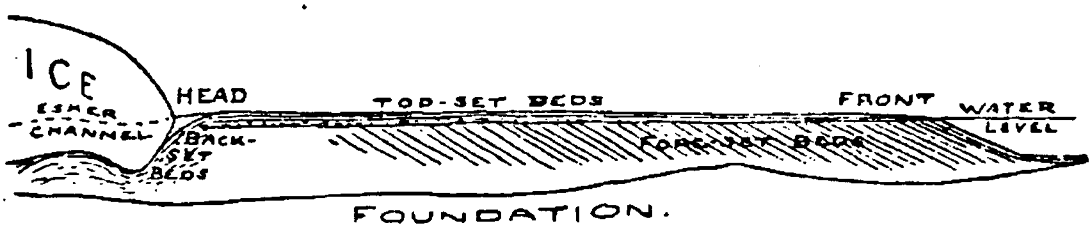
Jopling (1975) noted that Davis' backsets appear to be equivalent to *uppström* (upstream) bedding reported in eskers by de Geer. The term is generally recognised in Scandinavia although such beds are rarely reported (M. Punkari, personal communication). It is not clear whether de Geer observed them or inferred them to exist, as the term does not appear in his publications in Swedish (1882, 1897) or English (1912, 1940). However, they are a logical consequence of his model for the formation of eskers along a receding ice margin. De Geer showed that annual fans of proximal gravel to distal clay (varves) were:

> "placed as tiles, one over another, the uppermost always having their northern, or proximal border extending so much over that of the underlying as the ice-border had receded..." (1912, p. 245).

Eskers corresponded to the 'very handles' of these backset fans which de Geer related to the annual accumulation of deltas at the mouths of meltwater tunnels.

The stratigraphical concept of backset or upstream beds did not receive general recognition in the first half of the century. However, reverse depositional surfaces were identified in both modern and ancient glacial environments. Studies of the margins of Malaspina Glacier showed large scale surfaces inclined towards the glacier as a result of glaciofluvial deposition (e.g. Russell, 1893; Tarr, 1909). Equivalent surfaces were implicit in the interpretation of outwash cones at the mouths of tunnel-valleys in Denmark, with their apices tens of metres above the valley floors (Ussing, 1903), and their extension to the successive reverse slopes produced during recession (Werth, 1912).

The term 'backset' was re-introduced for fluvial stratigraphy by Power (1961). He observed backset surfaces inclined up to 25° within Pleistocene fluvial conglomerates and suggested them to record the upstream migration of anti-dunes during high velocity flow over an aggrading stream bed. This mode of formation was validated in flume experiments by Jopling and Richardson (1966) who produced backsets inclined up to 30-35° at high flow regime, beneath an induced hydraulic jump



Ideal Longitudinal Section of a Sand Plain.

Figure 4.4 - Backset formation proposed by Davis (1890).

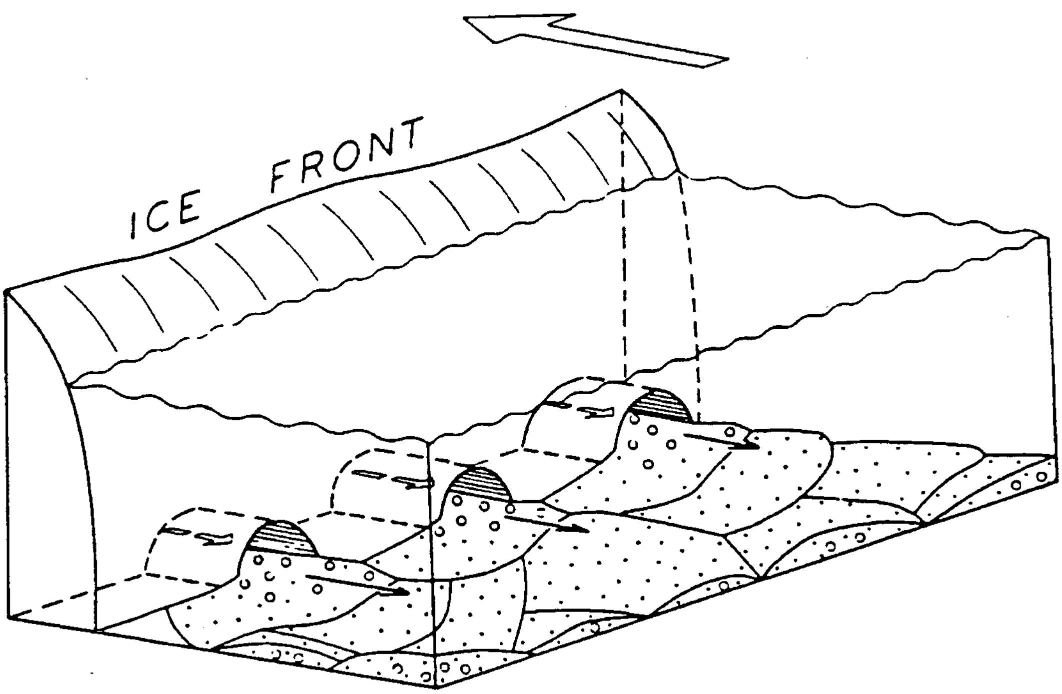


Figure 4.5 - Formation of sheet-like backset deposits by lateral and longitudinal migration of subaqueous fans during ice margin recession (from Rust and Romanelli, 1975).

across an obstacle to flow. They also produced foresets on the distal side of the obstacle and suggested that this compared to Davis' (1890) backset concept (Fig. 4.4). Backsets from various aqueous environments were subsequently attributed to anti-dune migration under high flow regime (Skipper and Bhattacharjee, 1978).

Banerjee and McDonald (1975) provided what seems to be the first example of glaciofluvial backsets since that of Davis (1890). Esker deposits in Canada were observed to include beds of gravelly coarse sand dipping up to 20° against the current direction (their Figure 3). They followed Jopling and Richardson (1966) in attributing the backsets to anti-dune migration under high regime flow in an open channel. However, Shephard (1973) had produced backsets inclined up to 20° in flume experiments at low flow regime and suggested that favourable conditions included a "pre-disposition surface" inclined upstream, decreasing stream power downstream and aggradation. This more general result may be applicable to the alternative of subglacial aggradation in closed channels.

Backsets imply the up-slope movement of water and sediment. A theoretical basis for up-slope movement was provided by Shreve (1972), who showed that glacial water flows in response to a potential gradient that is primarily controlled by ice surface slope and secondarily by basal relief. The effect of ice surface potential is to cause subglacial water to flow along a basal surface as though the latter were tilted down-glacier up to eleven times the ice surface slope (Shreve, 1985). Viewed in this sense, backsets may be regarded as 'effective' subhorizontal surfaces and their lateral progradation the result of vertical aggradation from streams at the base of the glacier (at low or high flow regime).

This seems to correspond to the situation envisaged by Davis (1890), and by de Geer (1912), whereby backward growth takes place as a natural consequence of subglacial streams which 'must needs rise' in response to ice marginal accumulation (Fig. 4.4). It is also compatible with the derived model of subaqueous outwash (Rust and Romanelli, 1975) described below.

The relatively steep gradients (up to 20°) of surfaces observed by Davis (1890) and Banerjee and McDonald (1975) imply lateral extents of less than 1 km if linearly extrapolated across reasonable sediment thicknesses. Bolduc (1993) inferred from lithological evidence that eskers in Labrador are composed of successive thin stratigraphical units 'shingled' up-glacier, with dips of <1° over distances of 5-15 km, that record sub-marginal accumulation during deglaciation. Backset surfaces of such gradients and dimensions would be difficult or impossible to recognise in field studies and this helps to explain why they have so seldom been reported.

4.3.1 b) subaqueous outwash

Rust and Romanelli (1975) presented a depositional model of outwash from laterally and longitudinally overlapping subaqueous fans in order to account for the sedimentological characteristics of large glaciofluvial ridges. They suggested that conditions for the time-transgressive formation of such deposits may have been widespread during Pleistocene deglaciation in the form of ponded fresh-water (or marine) basins along receding ice margins. The model has been successfully applied to other linear glaciofluvial deposits described as eskers or esker-ridges (e.g. Cheel, 1982; Thomas, 1984; Henderson, 1988; Deimer, 1988; Sharpe *et al.*, 1992).

Rust and Romanelli proposed their model in explanation of glaciofluvial sands and gravels, occurring in low (≤15 m) ridges up to 3 km wide and 15 km long, shown to have been deposited in up to 100 m of water. The sediments contained evidence of proximal to distal facies transitions characteristic of fan deposits, small-scale cross-bedding and large scale sandy channel fills. Interbedded tills occurred locally as well as ice-contact deformation structures. They envisaged the formation of irregular sheet-like deposits from *laterally* overlapping esker fans forming at the mouths of multiple tunnels. *Longitudinal* overlap of the fans during ice-margin retreat corresponds to backset formation (Fig. 4.5).

Rust and Romanelli used the term 'outwash' in preference to 'glaciofluvial' (deposits or activity), arguing that 'fluvial' cannot be applied to subglacial (roofed) streams. They acknowledged that this excludes some kinds of eskers from the category of glaciofluvial sediments. Their narrow distinction contradicts much common usage in which outwash and glaciofluvial deposits are broadly equivalent (e.g. Jopling, 1975; Sugden and John, 1976). I have distinguished backfill from proglacial outwash within the tunnel-valleys but this is merely a convenience of reference - both are composed of glaciofluvial sands.

4.3.2 Tunnel-Valley Record

The model of subaqueous outwash provides a basis for interpretation of the glaciofluvial backfill and outwash, in terms of a distributed system of subglacial streams feeding ice margin fans within the larger tunnel-valley basins. Sediment characteristics are accounted for in terms of lateral and longitudinal fan migration, with aggradation of sheet-like deposits promoted by channel shifting and grounding-line migration. The backfill may include intercalated tills, erosional surfaces or glaciotectonic rafts. Northward reduction of backset gradients in adjacent basins suggests changes in the glaciological factors driving water uphill (ice surface slope and basal water pressures). Proglacial outwash is inferred to have taken place by density currents.

I discuss a) backfill aggradation, b) backset gradients, and c) proglacial outwash.

4.3.2 a) backfill aggradation

The tunnel-valley fill backsets record sand-dominated deposition along surfaces 3-20 km in extent and 1-6 km wide. These dimensions are comparable to the ridges examined by Rust and Romanelli (1975). Deposition would have been subaqueous, in lake basins along the ice margin. The relief of the tunnel-valley basins implies water depths of up to 500 m.

I therefore propose glaciofluvial deposition from a distributed system of subglacial channels delivering water and sediment along gently upward-sloping surfaces (0.5-3.5°) to a grounding line position. Lateral and longitudinal overlap of channel-mouth fans would have led to the formation of sheet-like deposits (Fig. 4.5). Deposition may also have taken place within the channels after the manner discussed for esker-cores by Banerjee and McDonald (1975) and shifts in channel location over time would have further contributed to sheet-like deposits.

The depositional model accounts for the dominantly sandy character of the clinoform sequence recognised in downhole data. In addition, it can be applied to the textural cycles (both fining- and coarsening-up) and muddy interbeds inferred from well-logs (see Fig. 3.16). Vertical textural cycles can be attributed to migration of subaqueous outwash fans across a well location, either due to increasing distance from source (fining-up) or to progradation (coarsening-up). Muddy interbeds may correspond to sites of more distal or inter-fan deposition or to ice-contact deposition of till (Rust and Romanelli, 1975; Cheel, 1982). These interpretations may also be broadly applicable to the fill characteristics of northern German tunnel-valleys (Ehlers and Linke, 1989)

Interpretation in terms of fan migration accounts for the textural variations observed in individual wells, not all of which correspond to backset reflecting surfaces (section 3.4.2). However, it is less clear how it accounts for the backset surfaces themselves, which have axial extents of kilometres and at least some of which correspond to textural intervals correlative between wells (Fig. 3.6). An obvious solution lies in fluctuations of the ice margin, which in an aqueous setting corresponds to the migration of a grounding line position. Migration of the line of sediment delivery would distribute sediment across a backset surface in a manner analogous to marine transgression and regression. Fluctuations over a range of timescales are characteristic of ice marginal settings (e.g. Church and Gilbert, 1975).

Backset surfaces may also be generated by erosion, as I have shown for the discordant surfaces recording northwards erosional overlap by younger elements (section 2.5). These surfaces imply readvances of the ice margin during recession. By extension, lesser fluctuations of the margin could have lead to repeated erosion of previously deposited sediment and so a series of backset surfaces. However, the backsets (clinoform reflecting surfaces of sequence I) are generally concordant, and in any case an erosional mechanism does not account for their aggradation.

Imbricate surfaces inclined towards former ice margins are also characteristic of glaciotectonic structures. They are common within sedimentary strata due to the potential for elevated pore water pressures to reduce resistance to shear along bedding planes (Aber et al., 1989). Glaciotectonic rafts have been reported from boreholes through other buried tunnel-valleys (e.g. Potapenko, 1979; Ehlers and Linke, 1989; Ruszczynska-Szenajch, 1993). It is possible that such features are intercalated within the glaciofluvial backfill of the study area. However, deformation of the walls or base of the valleys was not observed on seismic reflection profiles. Shallow profiles indicate glaciotectonic structures of Elsterian age in the southwest of the study area (BGS/RGD, 1986) and to the south (BGS/RGD, 1984) but not in association with the tunnel-valleys. Glaciotectonism is promoted by elevated pore pressures and so may have been of little or no significance during the progradation of the fill in response to an efficient system of subglacial drainage.

4.3.2 b) backset gradients

The gradients of the backset surfaces may have been influenced by a number of factors, such as seasonal discharge fluctuations, sediment grain size, ice recession rates at different time scales, as well as the glaciological factors controlling water movement. However, the reduction in backset dip between southern and northern areas, common to adjacent tunnel-valleys, invites an explanation in terms of a change in the overall glaciological controls on drainage. I have suggested that backset surfaces might approximate water and sediment movement along effective subhorizontal surfaces, tilted up-glacier. If so, variations in backset gradient might be related to changes in the factors driving water flow: ice surface gradients and basal pressures.

Shreve (1972) showed that channelised water moves through and under glaciers in response to a fluid potential which is the sum of piezometric (ice-overburden) and atmospheric (gravity) gradients of pressure. Piezometric pressures dominate, so that lines of fluid equipotential are tilted up-glacier about 11 times the ice surface slope (Fig. 4.6). This assumes that water pressure in the conduits balances ice overburden pressure (effective pressure zero). Where water pressure is less (effective pressure greater), equipotential dip will be reduced to as little as twice the surface slope (Röthlisberger, 1972; Paterson, 1994, p. 114).

Overall changes in backset gradient (equipotential slope) could therefore relate to changes either in ice surface slopes, or in effective basal pressures. The observed reduction in backset gradient amounts to about 2-3°. This would correspond to a reduction of ice surface slope of as little as 0.2°, or up to 1° at higher effective pressures. Alternatively, higher overall effective pressures during deposition in the north would have resulted in a reduction in equipotential dip. The latter could follow variations in water flux, which in a strongly coupled system could dynamically influence ice surface slope (e.g. Arnold and Sharp, 1992).

4.3.2 c) proglacial outwash

Rust and Romanelli (1975) suggested that overlying lacustrine (or marine) sediments serve as an indication of subaqueous outwash deposits and that the contact between the two need not be conformable. The tunnel-valley record shows that a significant unconformity indeed separates backfill from proglacial sediments, but it lies within sands rather than between sands and muds (Fig. 4.2).

The outwash sands are comparable to the top- and foresets of Davis' (1890) model in that they record deposition contemporaneous with that of the backsets (Fig. 4.4). The subhorizontal surfaces presumably record subaqueous deposition from density currents (under-, inter- or over-flows) as recognised in modern ice marginal lakes supplied with sandy sediment by meltwater streams (e.g. Gustavson, 1975; Church and Gilbert, 1975). A gently foreset sand lobe is recognised locally (Fig. 3.8) and might correspond to a proximal outwash fan. The sands are expected to undergo distal facies transitions to finer grain-sizes.

Ehlers and Linke (1989) described a wedge of gently foreset interbedded sands and muds within a tunnel-valley in Hamburg, which thinned south away from the ice margin and included lateral facies variability. The wedge appears comparable to the proglacial outwash facies and implies that sands may not everywhere be so neatly overlain by muds. Ehlers and Linke interpreted the sediments to record pulses of water and sediment from the ice margin, which they attributed to episodic release of water from subglacial reservoirs. In the next section, I argue for an alternative of episodic release of water from the ice sheet surface.

4.4 Ice Sheet Hydrodynamics

The preceding sections sought to accommodate the stratigraphical and sedimentological characteristics of the tunnel-valleys in terms of erosion and redeposition by drainage beneath the receding ice margin. This explains how and where the tunnel-valleys formed but not why. In this section, I review the hydrology of ice sheet margins during deglaciation and suggest two causal factors in tunnel-valley formation. The first is the role of surface recharge to the basal drainage system as the melting ice sheet passed through its own margin. The second is the deformation of an unlithified and permeable bed in response to basal drainage, which implies tunnel development in order to maintain stability. I show that the dimensions of the tunnel-valley 'migrating basins' correspond to the expected width of a zone of surface water influence and that their geometry, size and spacing are consistent with predictions of tunnel development on a deforming bed.

I begin with a review of the englacial drainage of modern glaciers and of former ice sheets as inferred from the geological record of eskers. This shows that surface water has access to the bed in a zone tens of kilometres wide and that discharges increase by an order of magnitude during deglaciation. This 'deglacial flood' provides a basis for rejecting catastrophic models of ice sheet hydrodynamics and for equating tunnel-valleys with eskers. I then review the nature of subglacial drainage over deformable sediment beds and show that high surface water discharges are consistent with models of arborescent tunnel development beneath ice margins.

I discuss englacial drainage in modern glaciers (4.4.1), deglacial drainage of ice sheets as inferred from eskers (4.4.2), and the basal drainage system over a deformable bed (4.4.3).

4.4.1 Englacial Drainage (Modern Glaciers)

An understanding of glacial drainage derives from a combination of study of glaciers over human time-scales and mathematical analyses. The following discussion draws on comprehensive reviews by Röthlisberger and Lang (1987) and Paterson (1994). Water in a glacier flows in response to a single potential but from two sources: the ice surface and its base. Surface water originates from melting and direct precipitation in the marginal ablation zone, basal water from melting due to geothermal and frictional heating over larger areas. Surface water nonetheless dominates and descends to join the basal drainage system somewhere beneath the ice margin. These results primarily derive from valley glaciers but can be extended to ice sheets.

I review the flow of water within a) valley glaciers, and b) ice sheets.

4.4.1 a) valley glaciers

Surface water volumes are at least one order of magnitude larger than basal melt in valley glaciers. Discharge is enhanced by spatial and temporal factors - it is restricted to the ablation zone and is characterised by discharge fluctuations over a range of time-scales, notably the seasonal cycle.

Seasonal variability in the surface velocity of many valley glaciers in correspondence with changes in basal water pressures implies that surface water moves to the base of the ablation zone to join the basal drainage system. Downward flow may take place via intergranular veins and conduits, but a negligible flux through vein systems implies that most water moves through conduits (see Paterson, 1981, p. 136-138). Surface water access to the bed does not appear to be limited to 'temperate' ice (at pressure melting point) as even colder 'sub-polar' glaciers exhibit seasonal velocity variations (see Paterson, 1994, p. 124-125; also Andreasen, 1985). For example, Iken (1972) inferred surface water access through a sub-polar glacier up to 300 m thick.

Price (1973) observed that even in temperate ice the "actual mechanisms whereby meltwater develops a channel system ... are poorly understood" (p. 101) and this remains true. Stenborg (1968, 1973) emphasised the importance of surface cracks and fractures in initiating downward flow via crevasses and moulins. Moulins draining surface water are known to reform annually at the same location (e.g. Holmlund, 1988) and one such example has been traced downward to at least 100 m depth (la Mer de Glace, Reynaud, 1987). However, little is known of drainage at greater depths. Glen (1954) and Weertman (1973) argued that a column of water of sufficient thickness, being of greater density than ice, could cause natural fractures to propagate downward to the glacier bed. Sugden and John (1976) provided an interesting analogy with karst circulation, by extension from Shreve (1972).

Shreve (1972) argued that a system of conduits grows downward from the three-dimensional intergranular vein system. Veins and conduits represent a balance between melting of the walls due to dissipation of heat and closure due to deformation of ice. Shreve showed that larger conduits tend to grow at the expense of smaller ones due to more efficient dissipation of heat, as well as to the influence of ice flow toward the conduits on water movement. Furthermore, Röthlisberger (1972) showed that, in the steady-state, conduits are at lower water pressure for higher discharges, so that larger conduits tend to grow at the expense of smaller ones. These factors lead to a hypothetical three-dimensional network of downward coalescing (arborescent) conduits which direct water sub-vertically to the bed beneath the ablation zone (Fig. 4.6).

4.4.1 b) ice sheets

The extension of this picture of glacial drainage to ice sheets is limited mainly by their greater thickness. Valley glaciers may be up to several hundred metres thick in their ablation zones, whereas ice sheets can be similarly thick within kilometres of their margins. In the absence of significant calving along aqueous margins, ice sheets may have ablation zones over 100 km wide underlain by up

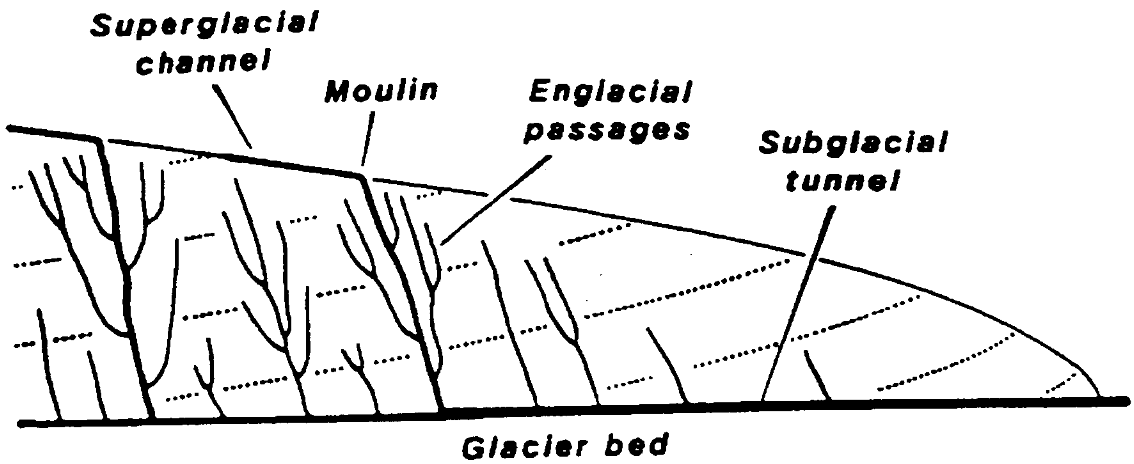


Figure 4.6 - Inferred englacial drainage system in an ice sheet margin. Surface water moves subvertically to the bed across fluid equipotential lines (dotted) through a downward coalescing network of conduits. Equipotential lines are tilted up-glacier 2-11 times the ice surface slope (vertical exaggeration 50-100 times). Water at the bed moves primarily in response to surface slope and secondarily to bed topography. From Shreve (1985).

to 1000 m of cold ice advected from the ice sheet summit (e.g. Arnold and Sharp, 1992). Cold, nominally impermeable ice might constitute a barrier to downward drainage, and so direct surface water subhorizontally towards the margin where it will have access to the bed in a relatively narrow zone of thinner ice (e.g. Boulton et al., 1995).

However, Drewry (1986) observed that "it seems clear from... theoretical considerations and radar observations that an extensive network of sub-horizontal channels cannot be supported within the body of a glacier" (p. 23). As seen above, although the means by which sub-vertical conduits might be established in temperate or cold valley-glaciers are not understood such conduits have nonetheless been inferred to exist. Their presence has also been inferred within the margin of the western Greenland ice sheet. Echelmeyer and Harrison (1990) reported seasonal upwelling of highly turbid water at the marine margin of Jakobshavns Isbrae, a fast-flowing ice stream up to 1000 m thick (which does not exhibit seasonal velocity variations). The estimated outflow significantly exceeded calculated basal melt discharges from which they concluded that at least part of the annual surface supply from the ablation zone (lower 80 km), which they estimated to account for over 90% of total meltwater discharge, came into contact with the bed at some point.

Surface ablation volumetrically dominates the drainage of ice sheet margins, by at least one order of magnitude (e.g. Boulton et al., 1995). I infer that this seasonal supply will descend via englacial conduits to join the basal drainage system at some point beneath the margin. This inference is supported by geological records of drainage beneath former ice sheets, which also indicate the width of such a zone of sub-marginal surface water influence.

4.4.2 Deglacial Drainage (Former Ice Sheets)

Eskers provide information on the hydrology of ice sheets during the last deglaciation of the northern hemisphere. Models of the formation of eskers indicate surface recharge to the bed via an englacial drainage system in a marginal zone tens of kilometres wide. I suggest that eskers and tunnel-valleys are equivalent and form concurrently. Surface water volumes would have increased as the ice sheet melted in a long-term discharge peak that corresponds to a 'deglacial flood' and which renders catastrophic release of water from subglacial reservoirs unnecessary.

I discuss a) evidence of sub-marginal surface recharge, b) equate eskers and tunnel-valleys, and c) consider tunnel-valleys and floods.

4.4.2 a) sub-marginal surface recharge (eskers)

Evidence for the sub-marginal influence of surface water derives from the interpretation of eskers, which record erosion and deposition by streams at the glacier bed in the form of arborescent networks of sand and gravel. Shreve (1972) suggested that the characteristics of eskers correspond to theoretical expectations of downward coalescing en- and subglacial drainage conduits within ice sheets (e.g. Fig. 4.5). He argued that the subglacial formation of eskers was favoured by high discharges and low ice surface slopes, consistent with surface recharge to the ice sheet base during deglaciation. He noted that most interpretations of eskers demonstrate a deglacial origin.

The original model of esker formation during deglaciation was that of de Geer (1897, 1912) based on deposits of the last deglaciation of Scandinavia. He correlated proximal esker cores with distal varve couplets which record seasonal supply of water from the ice sheet surface. He argued that:

> "...during the warm season of every year the melting-water from the surface of the great land-ice sank down through its crevasses and found its way along the bottom of the ice, where it was pushed forward under strong hydrostatic pressure..." (1912, p. 244).

In America, similar but independent views of ice sheet drainage during deglaciation were put forth by Hershey (1897) and Stone (1899) on the basis of esker characteristics and theoretical inference.

Banerjee and McDonald (1975) reviewed the nature of esker sedimentation and agreed with Shreve (1972) that eskers "are the products of a highly organised meltwater-flow system within the glacier and are common where an abundant supply of [surface] meltwater was available during deglaciation" (p. 153). They further concluded that eskers are formed time-transgressively by the headward migration of subglacial channels beneath a relatively narrow ice marginal zone. This zone was suggested to be up to 4 km wide, by reference to pebble transport studies in a Canadian esker. Donner (1965) summarised several lithological studies of transport distances in Finnish eskers, including of sand fractions, which showed that most material had travelled no more than 20 km from its source, and commonly less than 10 km.

Hebrand and Åmark (1989) studied eskers across 120 km of southern Sweden and presented a model of sub-marginal formation during deglaciation. They inferred that basal drainage processes were dominated by surface recharge from two main factors: the observed relation of eskers to seasonal varves, and the calculated insufficiency of basal melt alone in relation to esker (conduit) size and spacing. Esker spacing (average 5 km) was also used to make a rough estimate of the dimensions of the active sub-marginal zone as being "at least some tens of kilometres wide" (p. 77).

Bolduc (1992) studied arborescent esker systems up to 400 km long in Labrador and related them to sub-marginal deposition in conduits connected to an englacial drainage system which migrated headward during deglaciation. She recognised evidence for erosion and deposition by fluctuating flows due to surface recharge near the ice margin, in contrast to the steady subglacial flow inferred farther from the margin. The length of active channel was estimated from lithological evidence to be on the order of 25-30 km although lengths up to 100 km were possible from morphological evidence.

Mooers (1989a) presented evidence for the deglacial formation of eskers within tunnel-valleys in Minnesota, where segments 10-20 km long lead to outwash fans and begin at 'moulin kames'. The latter were interpreted to record seasonal surface melt reaching the bed through an englacial drainage system that migrated headward during ice margin retreat. Mooers emphasised the significance of a surface water source and noted the possible influence of such water in heating cold ice and so influencing the thermal evolution of the ice sheet margin.

Models of the deglacial formation of eskers thus invoke the influence of surface water beneath the ice margin in a zone up to 30 km wide which would have developed headward during recession. Mooers (1989a) showed that tunnel-valleys form in the same zone, and its dimensions are comparable to the 'migrating' sub-marginal basins I infer from tunnel-valley basal and fill characteristics.

4.4.2 b) eskers and tunnel-valleys

A sub-marginal and deglacial formation of tunnel-valleys during the melting of the ice sheet margin essentially corresponds to the original tunnel-valley concept of Ussing (1903-07) and Werth (1907-12), based on associations of erosional and depositional landforms with ice margin lines. The possibility of surface drainage to the bed during tunnel-valley formation had been suggested as early as 1884 by Jentzsch, and was assumed by later works by extension from de Geer's (1897) model of esker formation. Flint (1947) later observed that "if this interpretation of their origin is correct, then by no means are all subglacial stream courses marked by eskers" (p.160). Here I suggest that tunnel-valleys and eskers are equivalent cut and fill features and that where they coincide they formed concurrently.

Tunnel-valleys occur in sediment beds, whereas eskers are common to hard bed areas beneath former ice sheet centres, but their areas of occurrence overlap (e.g. Clark and Walder, 1994). Where eskers occur within tunnel-valleys, they are generally assumed to record deposition subsequent to erosion (e.g. Galon, 1983). Thus catastrophic models suggest a 'temporal disconnection' between tunnel-valley erosion and esker deposition during waning flood discharges (e.g. Wright, 1973; Shaw, 1983; Brennand and Sharp, 1993; Brennand and Shaw, 1994). Mooers (1989a) showed that esker segments occupied tunnel-valley segments, both formed during ice margin recession, but attributed the eskers to later depositional stages as a result of decreased discharges or increased sediment loads. The general assumption is that eskers are depositional features, while tunnel-valleys are erosive equivalents (e.g. Flint, 1947; Shreve, 1972; Boulton and Hindmarsh, 1987; Ehlers and Linke, 1989).

However, tunnel-valleys record contemporaneous erosion and deposition, as do eskers. Eskers are also cut and fill features, composed of material derived from the adjacent till and redeposited within a few tens of kilometres (e.g. Donner, 1965; Bolduc, 1993). Their erosional areas are poorly expressed over hard beds, but they commonly occupy shallow depressions or 'trenches' within the surface of a till sheet (e.g. Shilts, 1984; Hebrand and Åmark, 1989). I suggest that where they occur over sediment beds their erosional areas are expressed as tunnel-valleys. The expression of glaciofluvial sedimentation in smaller tunnel-valleys may remain that of an eskerine ridge (e.g Andersen, 1931; Wright, 1973), which records the impression of the subglacial channel responsible for erosion and redeposition. Wider tunnel-valleys will be occupied by a distributed system of channels (Rust and Romanelli, 1975). Deposition in a deep basin will also act to transform the expression of backset glaciofluvial sedimentation from that of a ridge to that of the backfill succession recognised in the tunnel-valleys of the study area.

Thus I propose that tunnel-valleys and eskers record the same basic processes of sub-marginal erosion and deposition during deglaciation, but the geological expression of those processes changes due to the nature of the glacier bed. The reason for the development of tunnel-valley basins could simply be that sedimentary beds are easily eroded and this leads to downward and headward growth. However, theoretical considerations suggest that tunnels form in response to bed deformation and predict a relationship between substrate character and erosional morphology that is observed in the study area. This is considered following a rejection of catastrophist models of tunnel-valley formation.

4.4.2 c) tunnel-valleys and floods

Sugden and John (1976) observed that deglaciation corresponds to a natural variation in the supply of surface water on a time scale of centuries. A simple calculation showed that the water released by surface melting "in one ablation season could be the equivalent of 5-15 years' precipitation" (p. 297) or an order of magnitude increase. This implies that surface water volumes, already at least an order of magnitude greater than basal melt, will increase by another order of magnitude during the course of deglaciation. Such a discharge variation over time is in a sense a flood.

Arnold and Sharp (1992) provided a similar estimate based on a numerical model of the Fennoscandian ice sheet. They calculated an increase in total ablation on the southern margin of up to ten times, over six centuries of deglaciation (their Figure 9a). Again, this this is additional to the basic dominance of surface over basal supply. They emphasised the temporal asymmetry of ice sheets with respect to meltwater production due to surface ablation of a relatively narrow (100-200 km) marginal zone. They noted the potential influence of such a zone on the dynamics of the ice sheet as a whole and on the geological record of glaciation.

Sugden and John (1976) suggested that "perhaps when envisaging meltwater conditions during the decay of such [mid-latitude] Pleistocene ice sheets, it is fair to be thinking in terms of... jökulhlaup[s]" (p. 297). They were referring to fluctuations of surface water supply over seasonal to deglacial time scales. However, their comment anticipated a controversy between the model of deglacial hydrodynamics inferred from eskers and alternative models for the contemporaneous action of widespread subglacial floods, suggested to apply to tunnel-valleys.

Two models propose the episodic release of water from hypothetical subglacial reservoirs, through a frozen margin (Wright, 1973), or as areal sheet-floods (Shaw, 1983). The sub-marginal flood model of Wright (1973) was based on two assumptions: that the tunnel-valleys he examined represented a single system 150 km that recorded drainage to a single ice margin and that surface water was unable to reach the bed through a 'frozen' ice margin (cold ice). These assumptions were both shown to be untenable by Mooers (1989a). Subsequent applications of this model have assumed a frozen margin (e.g. Boyd et al., 1988; Attig et al., 1989; Wingfield, 1990) and have neither provided additional evidence in support of floods nor evidence against the alternative.

The model of Shaw (1983) is incompatible with the demonstrated association of tunnel-valleys with ice margins. The model is based in the first instance on the explanation of drumlin formation, by analogy of their form with smaller-scale water erosion marks. It then proposes that erosion and deposition were both subglacial, during the waning of floods from subglacial reservoirs beneath the ice sheet centre. Shoemaker (1992) suggested that water from the melting margins of the ice sheet would find its way to such reservoirs, a pattern of drainage that seems difficult to reconcile with theory. Alternative explanations of drumlin formation are available, including of their formation beneath receding ice margins (Piotrowski, 1987; Mooers, 1989b).

I discount both of these models in favour of the progressive formation of tunnel-valleys beneath the ice margin during the course of the 'deglacial flood'. The release of large volumes of surface water over time obviates the need for the storage and episodic release of subglacial meltwater.

4.4.3 Subglacial Drainage and Deformation

Subglacial drainage over sediment beds is of interest due to its coupled effects on the dynamics of ice sheet flow (Boulton, 1986). In permeable sedimentary beds, elevated subglacial pore water pressures lead to reduced resistance to shear and allow bed deformation as a component of glacier motion (Mathews and Mackay, 1960; Boulton and Jones, 1979). Shoemaker (1986) assumed that a system of channels in the bed would develop so as to maintain positive effective pressures and prevent deformation. Boulton and Hindmarsh (1987) argued that such a system will develop so as to allow stable deformation and grow by positive feedback into a system of tunnels beneath the ice margin.

Analyses of channel dynamics suggest different possible geometries for subglacial channel systems over deforming beds. I infer that surface recharge would promote downward and headward growth of arborescent systems. The size and spacing of the study area tunnel-valleys are compatible with a system of tunnels determined by the characteristics of a deformable subglacial aquifer as proposed by Boulton and Hindmarsh (1987).

I review a) channel geometries, b) the influence of surface recharge and c) apply it to tunnel-valley formation.

4.4.3 a) channel geometries

Boulton and Hindmarsh (1987) presented an analysis of channel dynamics over a deforming bed in terms of the flow of ice and sediment. Further analyses were presented by Walder and Fowler (1994) and Clark and Walder (1994), reviewed by Paterson (1994). These analyses predict different possible geometries for subglacial drainage systems over permeable beds (Fig. 4.7). Paterson (1994) followed Walder and Fowler (1994) in recognising two systems: arborescent 'R-channels' in basal ice, and braided 'canals' in the bed. However, Boulton and Hindmarsh (1987) had allowed the formation of larger, ice-filled features of arborescent geometry, which they referred to as tunnels.

An arborescent system of R-channels represents the natural extension of the downward coalescing englacial network (Fig. 4.6) envisaged by Shreve (1972) and Röthlisberger (1972). An inverse relationship between discharge and water pressure (Röthlisberger) implies further subglacial convergence to relatively few large channels at the ice margin (Fig. 4.7a). Water may move into, or

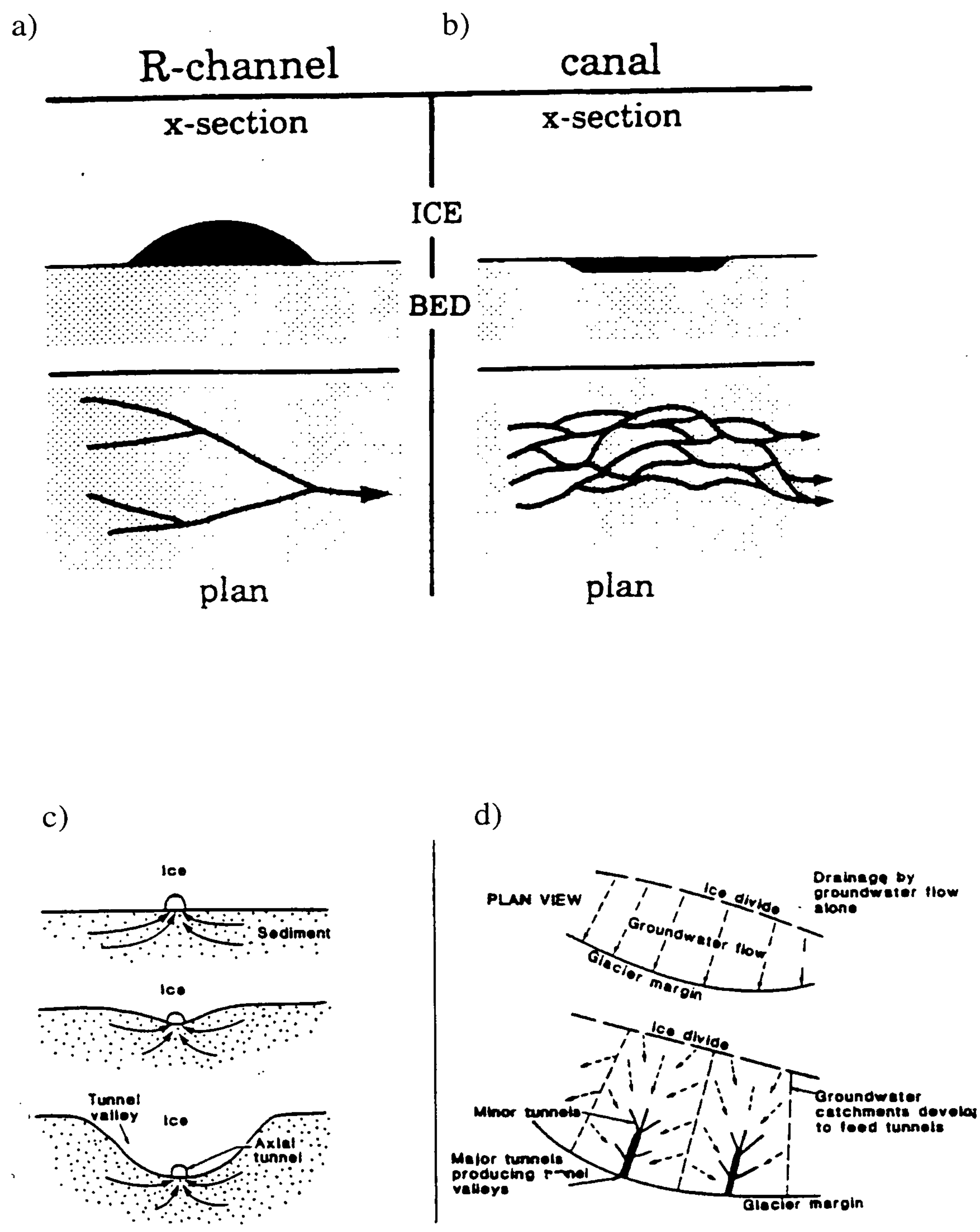


Figure 4.7 - Possible channel geometries over deformable sediment beds: **a,b**) arborescent R-channels in ice versus braided canals in sediment (from Clark and Walder, 1994); **c,d**) growth of large tunnels from smaller channel in response to sediment creep and fluvial erosion; an arborescent system allows headward and lateral growth in response to discharge fluctuations and ice margin recession (from Boulton and Hindmarsh, 1987).

from, the permeable substrate, but bed deformation plays no significant role in channel dynamics, which are a function of the flow and melting of ice.

Walder and Fowler (1994) used the term canal for their prediction of a flat ice-roofed channel incised into a sedimentary bed in response to bed deformation (Fig. 4.7b). Their analysis of channel dynamics (closure by ice and sediment flow, versus enlargement by melting of ice and erosion of sediment) yielded a direct relationship between discharge and water pressure. This implies no tendency for capture by larger conduits, and a braided system of broad, shallow canals (Fig. 4.7b).

Walder and Fowler (1994) argued that canals and R-channels could not stably co-exist, since lower pressure R-channels would tend to drain water away from canals. Which of the two is present would depend both on discharge, which they expressed as ice surface slope, and on the rheological properties of ice and sediment. Canals would be favoured by lower surface slopes (discharges) and pervasive bed deformation (low permeabilities), in contrast to R-channels. Clark and Walder (1994) therefore suggested that, beneath ice sheets, R-channels would occur over hard beds, and canals over soft (deforming) beds. However, they allowed that R-channels could occur over soft beds where sediment permeabilities were high, or ice-surface slopes (discharges) were high. The latter situation was noted to include ice sheet margins.

Boulton and Hindmarsh (1987) elaborated another influence on the geometry of subglacial channels. Channel enlargement due to sediment removal may be balanced by roof closure by ice flow leading to a lowering of the bed in the vicinity of the channel. Positive feedback leads to the formation of a deep, wide ice-filled 'tunnel' by one or more smaller drainage channels (Fig. 4.7c). The authors suggested the development of an arborescent tunnel system beneath ice sheet margins, with smaller channels converging to one or more trunk channels within each tunnel (Fig. 4.7d). An arborescent geometry allows lateral and headward expansion of the drainage system in response to ice margin or discharge fluctuations (Shoemaker, 1986; Boulton and Hindmarsh, 1987).

4.4.3 b) influence of surface recharge

These analyses do not fully address the effect of seasonal surface discharges beneath the ice margin, but inferences can be drawn to the effect that such discharges promote arborescent geometries.

Shoemaker (1986) considered that seasonal fluctuations in surface water supply would influence drainage geometries and that surface water arriving via an englacial system of arborescent conduits was likely to continue flowing in such a system (both features of the drainage of valley glaciers).

Walder and Fowler (1994) argued that if R-channels exist, they will drain water from canals and so be the stable drainage geometry. In their analysis, an arborescent system would be further promoted by higher surface slopes beneath ice margins due to the increase in discharges. Clark and Walder (1994) considered that surface water would recharge the basal system beneath a marginal zone but did not address the effect of the (dramatically) enhanced discharges.

Boulton and Hindmarsh (1987) did not address the source of the water that contributed to tunnel formation but it was noted that the system would be sensitive to perturbations. Higher discharges could be accommodated by larger or more closely-spaced tunnels and by lateral and headward expansion of the arborescent drainage system.

I conclude that an arborescent system of conduits and tunnels is theoretically probable beneath ice sheet margins due to several factors - higher ice surface slopes, seasonal high discharge of water through an englacial conduit system and the possibility of adjustment to discharge fluctuations. Positive feedback following bed erosion may lead to the growth of larger tunnels.

4.4.3 c) tunnel-valley formation

The tunnels envisaged by Boulton and Hindmarsh (1987) closely correspond to my interpretation of the southern North Sea examples - an arborescent system of sub-marginal basins that developed headward during recession. They suggested that tunnels will grow from smaller channels which lower the sediment surface by 'complementary processes of sediment creep and fluvial erosion'. My interpretation of the backfill implies that deposition took place via a distributed system of smaller channels within ice-filled basins. Seasonal discharge fluctuations would contribute to episodically enhanced sediment creep and erosion (summer discharges), which will be recorded in the sub-marginal accumulation of glaciofluvial backfill and outwash.

Boulton and Hindmarsh argued that the spacing of tunnels would depend on water supply and substrate transmissibility such that fewer tunnels would be necessary in more transmissive sediments. Spacing would also be related to size in that fewer larger tunnels would carry higher discharges. A dependence of size and spacing on the hydraulic properties of the substrate is recognised in the study area.

Tunnel-valley size, spacing and substrate thickness all increase eastwards across the study area (Fig. 2.21) towards the axis of the North Sea Basin. The physical character of the incised Pleistocene sediments approximates a large-scale coarsening-up sequence, from muddy prodelta to sandy delta-front and delta-top facies (Cameron et al., 1987; 1989a; 1992). The tunnel-valleys are incised into the upper sandy facies, which thickens proportionally to the east (Fig. 1.7a). The increase in sediment thickness thus corresponds to an increase in aquifer thickness. The change in tunnel-valley size and spacing can be seen as a response to the greater capacity of the aquifer to the east, i.e. a greater hydraulic transmissibility.

A similar correspondence between sediment thickness and tunnel-valley size may be seen in other areas. Elsterian tunnel-valleys beneath the Netherlands and northern Germany are up to 400 m deep (Ehlers et al., 1984), whereas to the east in Poland they only locally exceed 150 m depth (Mojski and Rühle, 1965; Marks, 1988). This reduction in relief coincides with the regional thinning of Cenozoic sediments from the North Sea Basin to the Polish Platform, from thicknesses over 1 km to generally less than 250 m (see Fig. 1.1; Ziegler, 1982). However, smaller tunnel-valleys also occur in areas of thick sediments within the North Sea Basin (e.g. Fig. 1.3), so factors other than thickness must also influence the erosional morphology.

The composition of the substrate will clearly be of importance. For example, relatively small and narrow Elsterian tunnel-valleys of East Anglia are incised into relatively resistant chalk (Woodland, 1970) of relatively low permeability. Boulton and Hindmarsh (1987) suggested that for a given substrate composition, size will be related to spacing. They also noted that the large Elsterian tunnel-valleys were all formed beneath a relatively large ice sheet, and it can also be observed that they occur along its southern margin (maximum extent, and so larger discharges). The interplay of glaciological and substrata characteristics will presumably have been the determinants of tunnel-valley morphology along the former ice margins. However, their transversal characteristics are a composite record of processes beneath the receding and melting ice sheet margin.

Thus I conclude that tunnel-valleys record the influence of englacial drainage to a deforming sediment bed. Substrate characteristics determined the spacing of the tunnel-valleys and their size for given discharge. Surface melting determined the discharges, both their spatial distribution beneath the outer 50 km of the ice margin and their variability over time.

CHAPTER 5
CONCLUSIONS

The starting point of this research was that the location, manner and causes of tunnel-valley formation are pertinent to conflicting models of the hydrodynamical behaviour of ice sheets over sedimentary basins. The thesis results clearly indicate where and in what manner tunnel-valleys form: beneath the receding ice margin and by the action of streams within larger basins. This deglacial mode of formation is essentially comparable to the original tunnel-valley concept of Ussing (1903-1907) and Werth (1907-1912). The temporal and spatial setting suggests formation in response to two processes, melting of the ice sheet margin and englacial recharge of that water to a deforming sediment bed.

I conclude by summarising the evidence for a sub-marginal and deglacial formation (section 5.1) as a response to surface recharge and bed deformation (5.2). Finally, I consider possible future avenues of research (5.3).

5.1 A SUB-MARGINAL AND DEGLACIAL FORMATION

The study area tunnel-valleys lie on the southwestern flank of the Cenozoic North Sea Basin and within a few hundred kilometres of the maximum extent attained by the Elsterian ice sheet. Seismic reflection and borehole data reveal a morpho-stratigraphical architecture indicative of drainage south to the ice margin during its northward recession across the study area. Basal morphology and fill stratigraphy are interpreted to record reworking of sedimentary material beneath the outer 50 km of the ice sheet.

The valley base defines a series of broad (0.5-6 km), relatively shallow (up to 500 m relief) and steep-sided (5-40°) elongate basins with an arborescent plan form convergent to the south. Erosional overlap to the north by younger basal (and fill) elements resulted in apparent anastomosing patterns and the angular offset of segments up to 20 km long. The erosional morphology is comparable to that of tunnel-valleys of the last deglaciation for which convergent drainage to receding ice margins resulted in plan form complexity due to spatial overlap.

The incised Cenozoic sediments have influenced the erosional morphology, in the form of benches on the walls determined by stratal contrasts and local changes in valley plan form across faults. However, the complex arborescent plan form was primarily determined by the orientation and location of the ice margin. An eastwards increase in the size and spacing of the tunnel-valley basins was influenced by substrate thickness (section 5.2).

The valley fill is dominated by a seismic sequence (I) that records axial progradation to the north along clinoform surfaces up to 3.5° in gradient and 3-20 km in extent. Well data indicate sand-dominated sediments up to 400 m thick which include vertical textural variability in part correlative to the reflecting surfaces. The sands accumulated beneath the receding ice margin by the up-slope movement of water and sediment along backset surfaces. An eskerine depositional model (Rust and Romanelli, 1975) is used to infer a distributed system of subglacial streams feeding subaqueous marginal fans within the larger tunnel-valley basins. Sheet-like deposits were promoted by channel shifting and by fan migration during marginal (grounding-line) fluctuations. The backfill sands may contain intercalated erosional surfaces, tills and glaciotectonic rafts.

The depositional surface of the backfill is an angularly discordant boundary of irregular axial relief formed by variable sediment accumulation during recession. Axial basins and sills on the backfill surface are correlative between some adjacent tunnel-valleys along lines which may parallel ice margins. The depositional basins are an expression of the incomplete backfill of the tunnel-valleys. It is not necessary to invoke the former presence of buried ice, as has been suggested for *Rinnenseen* in the Baltic lowlands and bathymetric deeps in the North Sea.

The axial basins are filled with a sequence of onlapping reflecting surfaces (II) the upper parts of which extend throughout adjacent tunnel-valleys. Borehole data indicate that lower sand-dominated sediments, indistinguishable from the underlying backfill, are overlain by muds. This is interpreted as an upwards transition from proximal (outwash) to distal accumulation within lake basins. An upper sequence of complex reflecting surfaces (III) corresponds to an interglacial marine foraminiferal zone in the upper muds. Sediment contrasts do not correspond to distinct seismic intervals as has been assumed, rather sequence boundaries lie within the sands and muds.

The morpho-stratigraphical evidence is explained by the headward excavation and backfilling of basins beneath the receding ice margin. Migration of the basins resulted in an axially diachronous stratigraphy, younging to the north. The vertical succession records the lateral migration of contemporaneous erosional and (proximal to distal) depositional environments over several millenia. Basin dimensions imply reworking of sedimentary materials beneath the outer 50 km of the ice sheet. Microfossils and components within the valley fill are derived from sedimentary strata as young as early Pleistocene, comparable to the succession into which the tunnel-valleys are incised.

5.2 Response to Surface Recharge and Basal Deformation

The causes of tunnel-valley formation are sought in a combination of en- and subglacial drainage processes during deglaciation. The dimensions of the 'migrating basins' are comparable to the expected width of a zone of surface water recharge to the basal drainage system. The erosional morphology is consistent with theoretical expectations of tunnel development in response to high discharges to a deformable subglacial aquifer.

Ice sheets are asymmetrical with respect to meltwater production, in that deglacial melting corresponds to a long-term discharge peak or 'flood'. Surface discharges exceed basal contributions by several orders of magnitude and esker systems are interpreted to indicate a well-developed englacial drainage system within the outer tens of kilometres of receding ice sheet margins. The progressive release of the reservoir of water embodied in the ice itself obviates the need for episodic discharges from large subglacial reservoirs as invoked in catastrophist models for the formation of tunnel-valleys and other landforms.

Tunnel-valleys form by contemporaneous erosion and deposition during deglaciation, as do the esker ridges which they in some places contain. The two features are suggested to be completely equivalent and to form concurrently. The expanded erosional record of tunnel-valley basins may transform the expression of glaciofluvial sedimentation from that of a ridge to that of basin backfill.

Deformation of a sediment bed in response to elevated pore water pressures requires the discharge of excess water via streams expected to grow by positive feedback into tunnels with a geometry determined by discharges and substrate characteristics. The eastwards increase in tunnel-valley size and spacing parallels an increase in substrate (aquifer) thickness. Their arborescent geometry is consistent with high discharges from a downward-coalescing system of conduits to a basal system that expanded laterally and headward in response to discharge and ice margin fluctuations.

5.3 Future Research

The thesis is based on exploration seismic and well data collected incidentally from the shallow section in the course of examining deeper targets. The successful results demonstrate the potential of such data for investigations of thick Quaternary successions. Seismic reflection data provide information of high spatial resolution (below 100-200 ms) despite the relatively low frequencies used in exploration. Seismic coverage of the North Sea and other areas of active exploration is constantly growing and this vast database may be available for academic research on request.

3D-seismic data constitute an unrealised resource for the study of Quaternary sedimentary basins. Volumetric data allow the continuous exploration of spatial variability and extend the interpretive power of morphological analogy to plan form perspectives of the subsurface. For glacial geological research, this means that interpretive methods such as applied to landform assemblages of the last deglaciation of onshore areas may be extended to buried landscapes of successive glaciations offshore.

The occurrence of 3D-seismic data and tunnel-valleys throughout the North Sea provides a ready means of testing the general applicability of their proposed sub-marginal and deglacial formation. Horizontal seismic sections should reveal complex arborescent systems of valleys convergent towards former ice margins. Seismic profiles generated along tunnel-valley axes should reveal prograding sequences of backset surfaces. Deviations from such an architecture would invite explanation in terms of changing ice sheet and substrate hydrological conditions during the course of deglaciations.

The geological record of the last ice sheets is mainly composed of features formed during deglaciation. The thesis results reflect both the likelihood that this is also true of prior glaciations and the potential significance of processes operating within a relatively narrow and mobile marginal zone. Recognition of the diachronous nature of the morpho-stratigraphical record is fundamental to its meaningful interpretation in terms of glacial processes. In the case of tunnel-valleys and eskers, process models should accommodate the basic spatial and temporal context of receding, melting ice margins.

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ABBREVIATIONS

DGU	Danmarks Geologiske Undersøgelse (Danish Geological Survey)
RGD	Rijks Geologische Dienst (Dutch Geological Survey)
BGS	British Geological Survey
fm	formation
bsl	below sea level
Ma	million years
ka	thousand years
BP	before present
ms	milliseconds
TWT	two-way (travel) time
J	joules
Hz	Hertz
rf	radius Fresnel zone
CMP	common mid-point
CDP	common depth point (used interchangeably with CMP)
NMO	normal move-out
S/N	signal to noise ratio
K.B.	Kelly Bushing (drill rig reference level)

APPENDIX

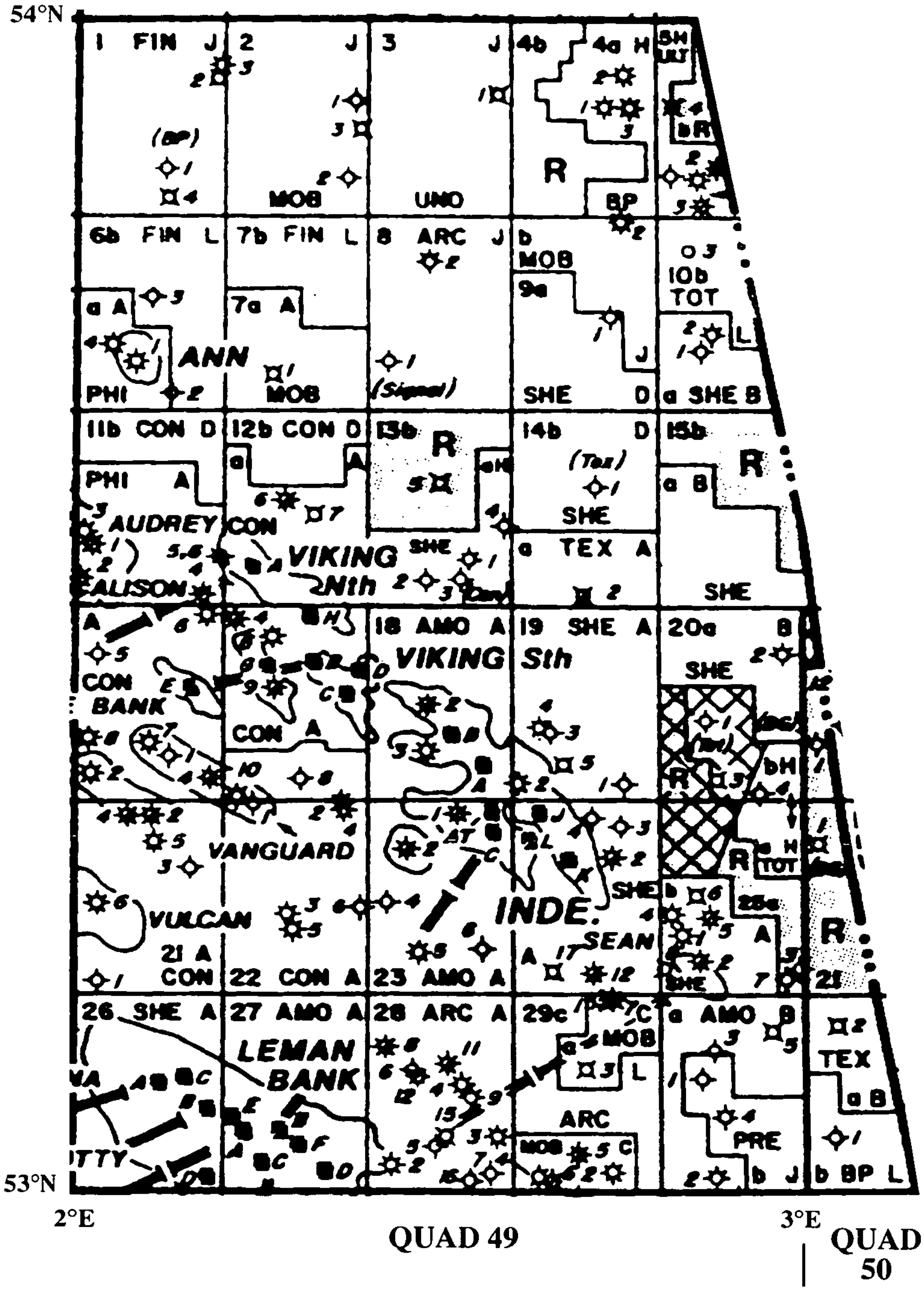
Shallow information available from exploration wells in the UK sector of the study area (Quads 49/50). Compiled from composite logs on file with DTI. Block locations (numbered 1-30, left to right), and many of the wells, are shown on the map below (from Amerada Hess, 1990 licencing round). UR = unreleased.

STRATIGRAPHY

Q = Quaternary
QT = Quaternary, Tertiary (undifferentiated)
Tert = Tertiary
Plio = Pliocene
Mio = Miocene
Eo = Eocene
Oli = Oligocene
Cr = Cretaceous
Jur = Jurassic
Tr = Triassic

LOGS

GR = gamma ray log
EL = electrical logs
SO = sonic log
cu = cuttings
mp = micropalaeontology



APPENDIX - QUAD 49/50 SHALLOW WELL LOG INFORMATION (1993)

WELL	OPER-ATOR	SPUD DATE	K.B. ft	1st RETURNS ft K.B.	1st RETURNS *m bsl*	BASE Q (QT) ft K.B.	BASE Q (QT) *m bsl*	Thickness Q (QT) *m*	Shallow Stratigraphy	Log Information (where known)
49/1-1	FIN	26.3.77	110	1214	*336*	-	-	-	*Eo*	
49/1-2	FIN	18.1.86	116	590	*144*	1550	*437*	*293*	*Q/Plio*	GR+cu
49/1-3	FIN	UR								
49/1-4	FIN	UR								
49/2-1	MOB	24.4.69	106	236	*40*	406	*91*	*52*	*Q/Cr*	GR
49/2-2	MOB	12.12.69	88	230	*43*	912	*251*	*208*	*QT/Cr*	GR
49/2-3	MOB	UR								
49/2-4	MOB	UR								
49/3-1	UNO	UR								
49/4-1	BP	15.6.84	42m	-	*158*	-	*212*	*54*	*Q/Plio*	GR+cu
49/4-2A	BP	2.4.86	32m	-	*200*	-	*360*	*160*	*Q/Plio*	GR+EL+SO+cu
49/4a-3	BP	UR								
49/5-1	ULT	25.3.67	95	700	*184*	-	-	-	*Tert*	GR+SO+cu
49/5-2	ULT	4.4.84	110	800	*210*	-	-	-	*QT*	GR+cu
49/5-3	ULT	3.8.85	119	1830	*522*	-	-	-	*QT*	GR
49/5a-4	ULT	UR								
49/5a-5	ULT	UR								
49/6-1	PHI	1.10.65	96	196	*30*	888	*241*	*211*	*QT/Eo*	GR+NL+cu
49/6-2	PHI	21.8.66	81	970	*271*	-	-	-	*Eo*	GR+SO+cu
49/6-3	FIN	8.3.69	84	1083	*304*	-	-	-	*Eo*	GR+SO+cu
49/6a-4		UR								
49/7a-1	MOB	UR								
49/8-1	ARC	28.10.66	101	338	*72*	748	*197*	*125*	*Q/Plio*	GR+cu (mp)
49/8-2	ARC	UR								
49/9-1	SHE	26.3.77	110	1214	*336*	1288	*359*	*23*	*QT/Eo*	GR+EL+SO+cu
49/9b-2	MOB	UR								
49/10-1	SHE	10.7.71	81	404	*98*	1395	*401*	*302*	*Q/T*	GR+El+SO+cu
49/10a-2	SHE	21.04.89	110	560	*137*					GR+cu
49/10b-3	TOT	UR								
49/11a-1	PHI	22.12.75	100	290	*58*	1034	*285*	*227*	*QT/Cr*	GR+cu
49/11a-2	PHI	13.8.83	112	1809	*517*	-	-	-	*Jur*	GR+EL+SO+cu
49/11a-3	PHI	30.3.84	120	400	*85*	1031	*278*	*192*	*QT/Cr*	GR+EL+SO+cu
49/11a-4	PHI	UR								
49/11a-5	PHI	UR								
49/11a-6	PHI	UR								
49/11b-8	CON	UR								
49/12-1	CON	1.1.69	106.5	189.5	*25*	907	*244*	*219*	*QT/K*	GR+SO+cu
49/12-2	CON	10.2.69	106.5	2400	*699*	-	-	-	*Cr*	uK
49/12-3	CON	18.12.69	105	800	*212*	845	*226*	*14*	*QT/Cr*	GR+cu
49/12a-4	CON	29.6.73	100	710	*186*	795	*212*	*26*	*QT/Cr*	GR+SO+cu
49/12a-5	CON	9.8.73	100	350	*76*	780	*207*	*131*		cu
49/12a-6	CON	3.8.74	103	1425	*403*	2480	*725*	*322*		GR+SO+cu
49/12a-7	CON	10.8.86	110	1190	*329*	-	-	-		
49/12a-8	CON	UR								
49/13-1	SHE	24.6.65	31	513	*147*	567	*163*	*16*	*Q/Plio*	GR+EL+cu

APPENDIX - QUAD 49/50 SHALLOW WELL LOG INFORMATION (1993)

WELL	OPER-ATOR	SPUD DATE	K.B. ft	1st RETURNS ft K.B.	*m bsl*	BASE Q (QT) ft K.B.	*m bsl*	Thickness Q (QT) *m*	Shallow Stratigraphy	Log Information (where known)
49/13-2	SHE	28.3.69	105	2800	*821*	-	-	-	*Cr*	
49/13-3A	SHE (Con)	3.7.77	112	1200	*332*	1513	*427*	*95*	*QT/Cr*	GR+EL+SO+cu
49/13-4	SHE	23.7.77	114	370	*78*	2208	*638*	*560*	*QT/Cr*	GR+cu (EL+SO@300m
49/13b-5	R	UR								
49/14-1	SHE (Tex)	22.12.67	98	295	*60*	990	*272*	*212*	*Q/Plio*	GR+cu (mp)
49/14a-2	TEX	5.6.89	124	373	*76*					GR
49/16-1	CON	16.10.69	100	300	*61*	1088	*301*	*240*	*QT/Cr*	GR+cu+SO@187m
49/16-2	CON	22.9.70	104	830	*221*	1071	*295*	*73*	*QT/Cr*	GR+NL+cu
49/16-3	CON	29.11.70	109	660	*168*	600?	*150?*	-	*Jur?*	GR+cu
49/16-4	CON	4.7.72	107	746	*195*	974	*264*	*69*	*QT/Cr*	GR+cu
49/16-5	CON	12.8.72	117	1140	*312*	1843	*526*	*214*	*QT/Cr*	GR+cu
49/16-6	CON	3.2.73	120	360	*73*	580	*140*	*67*	*QT/Cr*	GR+cu
49/16-7	CON	23.10.82	126	1060	*285*	-	-	-	*Cr*	
49/16-8	CON	27.6.83	114	1129	*309*	-	-	-	*Cr*	
49/16-9	CON	10.9.83	116	1200	*330*	-	-	-	*Jur*	
49/16-10	CON	19.11.83	101	1221	*341*	-	-	-	*Cr*	
49/17-1	CON	28.8.65	95	234	*42*	754	*201*	*158*	*QT/Cr*	GR+cu
49/17-2	CON	13.4.68	86	922	*255*	1066	*299*	*44*	*QT/Cr*	GR+SO+cu
49/17-3	CON	30.10.68	106	805	*213*	805	*213*	-	*Cr*	
49/17-4	CON	6.5.69	103	1015	*278*	1295	*363*	*85*	*QT/Cr?*	GR+SO+cu
49/17-5	CON	24.7.69	112.5	1108	*303*	-	-	-	*Cr*	
49/17-6	CON	8.9.69	105	750	*197*	-	-	-	*Cr*	
49/17-7	CON	15.11.69	107	984	*267*	-	-	-	*Cr*	
49/17-8	CON	31.10.72	112	360	*76*	1176	*324*	*249*	*QT/Cr*	GR+cu+SO@179m
49/17-9	CON	15.12.72	105	368	*80*	878	*236*	*155*	*QT/Cr*	GR+cu
49/17b-10	CON	UR								
49/18-1	AMO	8.4.66	92	300	*63*	732	*195*	*132*	*Q/Plio*	GR+EL+cu
49/18-2	AMO	3.2.67	95	288	*59*	713	*188*	*130*	*Q/Plio*	GR+SO+cu
49/18-3	AMO	10.6.67	84	304	*67*	676	*180*	*113*	*Q/Plio*	GR+SO+cu
49/18-4	AMO	23.10.67	90	315	*69*	1248	*353*	*284*	*QT/Cr*	GR+SO+cu
49/19-1	SHE	29.9.65	81	336	*78*	1099	*310*	*233*	*Q/Eo*	QGR+cu+EL±SO
49/19-2	SHE	27.11.67	84	350	*81*	770	*209*	*128*	*Q/Plio*	cu+GR+EL+SO
49/19-3	SHE	4.2.68	89.5	820	*223*	863	*236*	*13*	*Q/Tert*	
49/19-4	SHE	9.1.86	121	471	*107*	858	*225*	*118*	*Q/Plio*	GR+SO+EL+cu
49/19-5	SHE	UR								
49/20a-1	TOT	2.1.67	91	350	*79*	882	*241*	*162*	*Q/Plio*	GR+cu+SO@216m
49/20b-2	SHE	25.2.71	91	408	*97*	1100	*308*	*211*	*Q/Plio*	GR+SO+cu
49/20a-3	TOT	27.8.84	118	932	*248*	?	*?*	*?*	*Q/Mio*	
49/20a-4	TOT	UR								
49/21-1	CON	11.10.66	84	350?	*?*		*?*	*?*		
49/21-2	CON	14.5.70	103	600	*151*	778	*206*	*54*	*QT/Cr*	GR+cu
49/21-3	CON	27.3.72	98	760	*202*	1120	*312*	*110*	*QT/Cr*	GR+cu
49/21-4	CON	22.5.72	104	350	*75*	917	*248*	*173*	*QT/Cr*	GR+cu
49/21-5	CON	11.1.83	113	230	*36*	927	*248*	*212*	*QT/Cr*	GR
49/21-6	CON	5.3.83	113	192	*24*	440	*100*	*76*	*QT/Tr*	GR
49/22-1	CON	11.10.66	84	190	*32*	673	*180*	*147*	*Q/Plio*	GR+cu
49/22-2	CON	6.4.72	109	450	*104*	1107	*304*	*200*	*QT/Cr*	GR+cu
49/22-3	CON	29.9.72	115	370	*78*	1263	*350*	*272*	*QT/Cr*	GR+cu
49/22-4	CON	8.3.82	119	1070	*290*	-	-	-	*QT*	
49/22-5	CON	?	103	1171	*326*	-	-	-	*Cr*	

APPENDIX - QUAD 49/50 SHALLOW WELL LOG INFORMATION (1993)

WELL	OPER-ATOR	SPUD DATE	K.B. ft	1st RETURNS ft K.B.	1st RETURNS *m bsl*	BASE Q (QT) ft K.B.	BASE Q (QT) *m bsl*	Thickness Q (QT) *m*	Shallow Stratigraphy	Log Information (where known)
49/22-6	CON	UR								
49/22-7	CON	UR								
49/22-8	CON	UR								
49/23-1	AMO	22.9.66	93	300	*63*	713	*189*	*126*	*Q/Plio*	GR+EL+cu
49/23-2	AMO	9.7.67	83	300	*66*	621	*164*	*98*	*Q/Plio*	GR+SO+cu
49/23-3	AMO	2.9.67	85	302	*66*	742	*200*	*134*	*Q/Plio*	GR+SO+cu
49/23-4	AMO	26.1.70	95	325	*70*	620	*160*	*90*	*Q/Plio*	GR+SO+cu
49/23-5	AMO	UR								
49/23-6	AMO	UR								
49/24-1	SHE	4.11.67	91	798	*215*	847	*230*	*15*	*Q/Plio*	GR+SO+cu
49/24-2	SHE	29.8.68	100	300	*61*	883	*239*	*178*	*Q/Plio*	cu (+GR+SO@214m)
49/24-3	SHE	12.1.69	96	500	*123*	1120	*312*	*189*	*Q?/Oli*	GR+SO+cu
49/24-4	SHE	5.2.70	97	388	*89*	525	*130*	*42*	*Q/Plio*	GR+SO+cu
49/24-12	SHE	25.10.71	99	800	*214*	-	-	-	*Plio*	GR+SO+cu
49/24-16	SHE	22.2.74	96	400	*93*	?	*?*	*?*	*QT*	GR+cu
49/24-17	SHE	UR								
49/25-1	SHE	15.2.69	96	350	*77*	1083	*301*	*223*	*Q/Plio*	GR+cu (+SO@123m)
49/25-2	SHE	8.12.69	97	355	*79*	1045	*289*	*210*	*Q/Plio*	GR+SO+cu
49/25-3	TOT	15.1.71	88	333	*75*	1145	*322*	*247*	*Q/Plio*	GR+SO+cu
49/25a-4	SHE	1.6.82	106	2430	*708*	-			*Cr*	
49/25a-5	SHE	29.3.83	114	1510	*426*	?			*Tert*	GR+EL@26m
49/25a-6	SHE	22.6.86	113	448	*102*	?			*QT*	GR+EL+SO+cu
49/25c-7	TOT	UR								
49/25a-8	SHE	9.9.88 UR	116	500	*117*	?				
49/26-1	SHE	17.12.65	84	382?	*91?*	622	*164*	*73?*	*Q/Tr*	
49/26-2	SHE	21.4.66	88	368	*85*	527	*134*	*48*	*Q/Tr*	
49/26-3	SHE	18.5.66	82	329	*75*	515	*132*	*57*	*Q/Tr*	
49/26-4	SHE	10.8.66	80	626	*166*	652	*174*	*8*	*Q/Tr*	
49/26-5	SHE	20.2.66	80	503	*129*	563	*147*	*18*	*Q/Tr*	
49/26-14	SHE	1.11.68	100	558	*140*	558	*140*	-	*Tr*	
49/26-17	SHE	1969								
49/26-25	SHE	1970								
49/26-26	SHE	?								
49/26-E1	SHE	6.3.82	118	425	*94*				*?*	
49/26-F1	SHE	5.1.86	125	450	*99*	610	*148*	*49*	*QT/Tr*	GR+cu
49/26-G1	SHE	10.4.86	120	1600	*451*				*Tr*	
49/27-1	AMO	13.6.66	117	334	*66*	576	*140*	*74*	*Q/Cr*	GR+EL+cu
49/27-2	AMO	12.12.66	85	290	*62*	526	*134*	*72*	*Q/Plio*	GR+SO+cu
49/27-3	AMO	27.1.67	92	308	*66*	627	*163*	*97*	*Q/Cr*	GR+SO+cu
49/27-4	AMO	18.4.67	81	363	*86*	717	*194*	*108*	*QT/Cr*	GR+SO+cu
49/28-1	ARC	9.11.67	90	200	*34*	694	*184*	*151*	*Q/Tert*	GR+cu
49/28-2	ARC	25.1.68	96	206	*34*	723	*191*	*158*	*Q/Tert*	GR+cu
49/28-3	ARC	4.4.68	85	205	*37*	504	*128*	*91*	*Q/Tert*	GR+cu
49/28-4	ARC	2.11.73	117	748	*192*				*Tert*	
49/28-5	ARC	19.12.73	117	750	*193*				*Tert*	
49/28-6	ARC	21.1.74	116	726	*186*				*Tert*	
49/28-7	ARC	30.8.82	115	746	*192*				*Tert*	
49/28-8	ARC	25.2.83	116	780	*202*				*Tert*	
49/28-9	ARC	22.1.84	114	1462	*411*				*CR*	
49/28-10	ARC	7.4.84	123	2000	*572*	1495	*418*	-	*QT/Cr*	GR
49/28-11	ARC	6.7.84	113	1882	*539*				*Jur*	
49/28-12	ARC	4.8.84	115	1884	*539*	1067	*290*	-	*Jur*	

APPENDIX - QUAD 49/50 SHALLOW WELL LOG INFORMATION (1993)

WELL	OPER-ATOR	SPUD DATE	K.B. ft	1st RETURNS ft K.B.	*m bsl*	BASE Q (QT) ft K.B.	*m bsl*	Thickness Q (QT) *m*	Shallow Stratigraphy	Log Information (where known)
49/28-13	ARC	16.1.85	116	1889	*540*	1456	*408*	-	*Cr*	
49/28-14	ARC	7.3.87	121	540	*128*	1096	*297*	-	*QT/Cr*	GR+cu
49/28-15	ARC	UR								
49/28-16	ARC	UR								
49/29a-1	MOB	25.8.70	85	344	*79*	840	*230*	*151*	*Q/Eo*	GR+cu+SO
49/29b-2	MOB	19.4.76	102	530	*130*	≥1600	*≥457*	*≥326*	*QT*	
49/29a-3	MOB	27.7.85	111	400	*88*	850	*225*	*137*	*Q/Plio*	GR+cu
49/29b-4	MOB	19.5.86	94	400	*93*	645	*168*	*75*	*Q/Plio*	GR+cu
49/29b-5	MOB	UR								
49/29b-6	MOB	UR								
49/29a-7	MOB	UR								
49/30b-1	PRE(Amo)	5.3.68	100	349	*76*	1077	*298*	*222*	*Q/Plio*	GR+SO+cu
49/30a-2	AMO	27.12.69	89	287	*60*	1046	*292*	*231*	*Q/Plio*	GR+SO+cu
49/30a-3	AMO	22.11.73	106	420	*96*	?			*QT*	GR+cu
49/30b-4	PRE	UR								
49/30a-5	AMO	UR								
44/29c-1	R	29.2.84	85	686	*183*	980	*273*	*90*	*Q/Plio*	GR+cu
44/29b-2	FIN	22.4.85	85	740	*200*	922	*255*	*55*	*Q/Plio*	GR+EL+SO+cu
44/29b-3	FIN	UR								
50/16-1	BGC	19.4.84	98	1500	*427*	2749	*808*	*381*	*QT*	GR+EL+SO+cu
50/21-1	BGC	24.1.83	102	1220	*341*	1763	*506*	*166*	*Q/T*	GR+EL+SO+cu
50/26-1	BP	-	97	292	*59*	1480	*422*	*362*	*Q/Mio*	GR+SO+cu
50/26-2	TEX	UR								