ELSEVIER

Journal of Structural Geology 27 (2005) 587–596

JOURNAL OF STRUCTURAL GEOLOGY

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Numerical modelling of thrust structures in unconsolidated sediments: implications for glaciotectonic deformation

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Received 13 September 2004
Available online 17 March 2005

Abstract

The mechanics of thrust faulting induced by a differential vertical load has been studied by application of a dynamic finite element model. The modelled scenario is analogous to the glaciostatic load imposed onto an unconsolidated sedimentary substrate by a stationary ice sheet. The objective is to evaluate whether large-scale glaciotectonic deformation may be induced by *gravity spreading* in front of a stationary glacier or if a *push from the rear* is required. The model is capable of handling large strains and rotations combined with visco-elastic-plastic rheologies. Model parameters are derived from a Pleistocene large-scale glaciotectonic thrust complex in the eastern Danish North Sea. The plastic behaviour of the unconsolidated sediments is described by a time and pressure dependent Drücker–Prager yield criterion combined with a non-associated flow law. The modelling results show that gravity spreading in front of a stationary ice-sheet can form strain localisations and possibly cause thrust faults to nucleate in unconsolidated sediments several kilometres beyond the snout of the ice-sheet, provided a décollement surface is situated at a suitable depth relative to the effective load of the ice-sheet. Hence it is concluded that large-scale glaciotectonic thrusting may be formed by *gravity spreading* alone, and a *push from the rear* is not generally required to explain the observed deformation.
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Keywords: Numerical modelling; Finite element; Thrust faulting; Large-scale glaciotectonics

1. Introduction

Glaciotectonic thrust complexes may be regarded as small-scale analogues to thin-skinned thrust structures found in foreland fold-and-thrust belts that developed in response to crustal shortening and mountain building in areas of plate tectonic convergence (e.g. Croot, 1987; Pedersen, 1987; Aber et al., 1989; Klint and Pedersen, 1995; Huuse and Lykke-Andersen, 2000; Andersen, 2004). The mechanical behaviour of thrust structures has previously been studied by physical and numerical modelling and used as analogues to thrust structures in orogenic or accretionary wedge settings (e.g. Borja and Dreiss, 1989; Storti et al., 1997; Cobbold et al., 2001; Strayer et al., 2001; Costa and Vendeville, 2002; Panian and Wiltschko, 2004). In these models, lateral shortening was induced by a direct push of the sediment package from one end whilst the other end remained fixed.

Two conceptual models, which both act on a sediment package by inducing thrust faults, have been put forward to explain the mechanism of thrusting. In the first model, *gravity spreading* in front of a static differential load was used to explain the formation of thrust structures (Bucher, 1956; Price and Mountjoy, 1970; Elliot, 1976). The second model applied a direct *push from the rear* to explain the same kind of structures (Chapple, 1978; Davis et al., 1983; Dahlen et al., 1984).

In a glacial setting the *gravity spreading* model implies that the differential load of the ice-sheet (the glaciostatic force) is the cause of proglacial thrusting in the subsurface sediments (Rotnicki, 1976; van der Wateren, 1985; Pedersen, 1987; Aber et al., 1989). In the model invoking a *push from the rear*, the compressive stresses that initiate the thrusting are provided by the forward motion of the ice

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0191-8141/$ - see front matter © 2005 Elsevier Ltd. All rights reserved.
doi:10.1016/j.jsg.2005.01.005

(Croot, 1987; Aber et al., 1989). Different opinions thus exist on whether the dominant driving force responsible for ice-marginal thrusting is the glaciostatic force (*gravity spreading*) or glaciodynamic forces (*push from the rear*). Based on geological observations in a natural environment it is difficult to separate these forces, but this can be accomplished by numerical modelling.

This study is a first attempt to use the finite element model to model thrust faults as the result of a differential load with the shape of an ice-sheet. It is intended to use the model as an analogue to a simplified glacial setting where the mechanical response of the subsurface to the isolated load of an ice-sheet (the glaciostatic strain) can be studied.

The rheological properties of ice cause the 2D profile of an ice sheet to approximate a parabolic shape from its interior to its snout (Paterson, 1994). The ice sheet imposes a differential load onto the sedimentary substrate that leads to rotation of the vertical stress field such that large compressive horizontal stresses occur beyond the ice-sheet margin. This phenomenon has been described by Aber et al. (1989) who considered a 10-km-long ice sheet with a parabolic top descending from 1 km at the maximum ice-sheet thickness to zero at its snout. We use the same ice-profile in our modelling study. The steepness of the ice-sheet profile is dependent on the ability of the ice to flow over its bed, and the properties of the subsurface sediments thus play an important role (Paterson, 1994). Deformations inside the ice sheet can also change the ice profile (Kalin, 1971; Hudleston, 1992), but it would be rare for ice thickness to change by much more than 1000 m over a distance of 10 km at the ice-margin (Aber et al., 1989).

The geological parameters used in our model have been derived from the large-scale glaciotectonic Fanø Bugt thrust complex situated in the south-eastern Danish North Sea (Huuse and Lykke-Andersen, 2000; Andersen, 2004). The Fanø Bugt thrust complex, described in detail below, is comprised of large-scale glaciotectonic thrust structures formed during the Pleistocene glaciations when ice sheets repeatedly advanced into lowland areas characterised by thick successions of poorly consolidated sediments. In the present interglacial, glaciers resting on unconsolidated sediments are rare, and the mechanisms leading to the formation of large-scale glaciotectonic thrust complexes are thus difficult to study in situ.

Our modelling study attempts to explore the formation of large-scale glaciotectonic structures with length scales of the order of hundreds of metres. A priori, it seems plausible that the formation of such large-scale structures is different from the formation of push moraines that contain imbricate stacks of sediment slabs, tens of metres in length (e.g. Croot, 1987; Krüger, 1994; Boulton et al., 1999). Push moraines might be explained by the action of a push from behind (bulldozing) by an advancing ice-sheet, but the formation of large-scale glaciotectonic structures can be difficult to explain by the action of bulldozing alone, because of the low shear strength of ice compared with the shear strength of a sediment block of hundreds of metres thickness. Our hypothesis is thus that the differential load provided by the weight of the ice-sheet is the key to large-scale glaciotectonic deformation.

We employ a 2D dynamic finite element model to investigate the mechanics of glaciotectonic deformation, and we show that the model has the potential for modelling thrust faults formed by differential loading of a sedimentary succession. The modelling results enable us to describe the strain localisations imposed by a static differential load with the shape of an ice sheet resting on unconsolidated sediments. The results furthermore provide information regarding the importance of the dip and depth of the basal décollement surface. The modelling results suggest that large-scale glaciotectonic deformation can be formed by *gravity spreading* in front of a static glacier without the need to invoke a *push from the rear*.

2. Geological model

The geological model is based on the large-scale glaciotectonic Fanø Bugt thrust complex in the south-eastern Danish North Sea, first discovered by Huuse and Lykke-Andersen (2000) and later described in detail by Andersen (2004). The glaciotectonic complex is covered by seismic data off the west coast of southern Denmark in an area of 15 km E–W by 40 km N–S. The total extent of the thrust complex may be much larger, as only the western boundary of the thrust complex can be outlined on the seismic data; towards the east, north-east and south-east the deformations continue towards the shore, and similar structures can be observed onshore in an area 20 km east of the off-shore Fanø Bugt thrust complex (Andersen, 2004).

The Fanø Bugt glaciotectonic thrust complex consists of folded and thrusted sheets of sediment that have slid on a weakly dipping basal décollement surface (Fig. 1). Individual thrust sheets can be traced up to ca. 1.5 km in the dip direction and for several km in the strike direction. The basal décollement surface is dipping approximately 0.5° towards the WSW and is situated at depths between ca. 200 m in the east and about 360 m in the west. Internal décollement surfaces occur in parts of the complex, typically situated at a depth of 100–150 m (Andersen, 2004). The basal décollement surface is situated within clayey sediments of the lower Miocene Arnum Formation (Andersen, 2004). The deformed succession thus comprises Neogene sediments, which mainly consist of clays, silts and fine-grained sands (Rasmussen, 2004).

Based on the Fanø Bugt thrust complex, we constructed a simple two-layer geological model (Fig. 2). The principal outline of our model includes an ice-sheet of 1 km regional thickness, decreasing in thickness over a distance of 10 km towards its snout (Aber et al., 1989). The ice-sheet rests on a 1-km-thick sediment-pile with 250 m of brittle

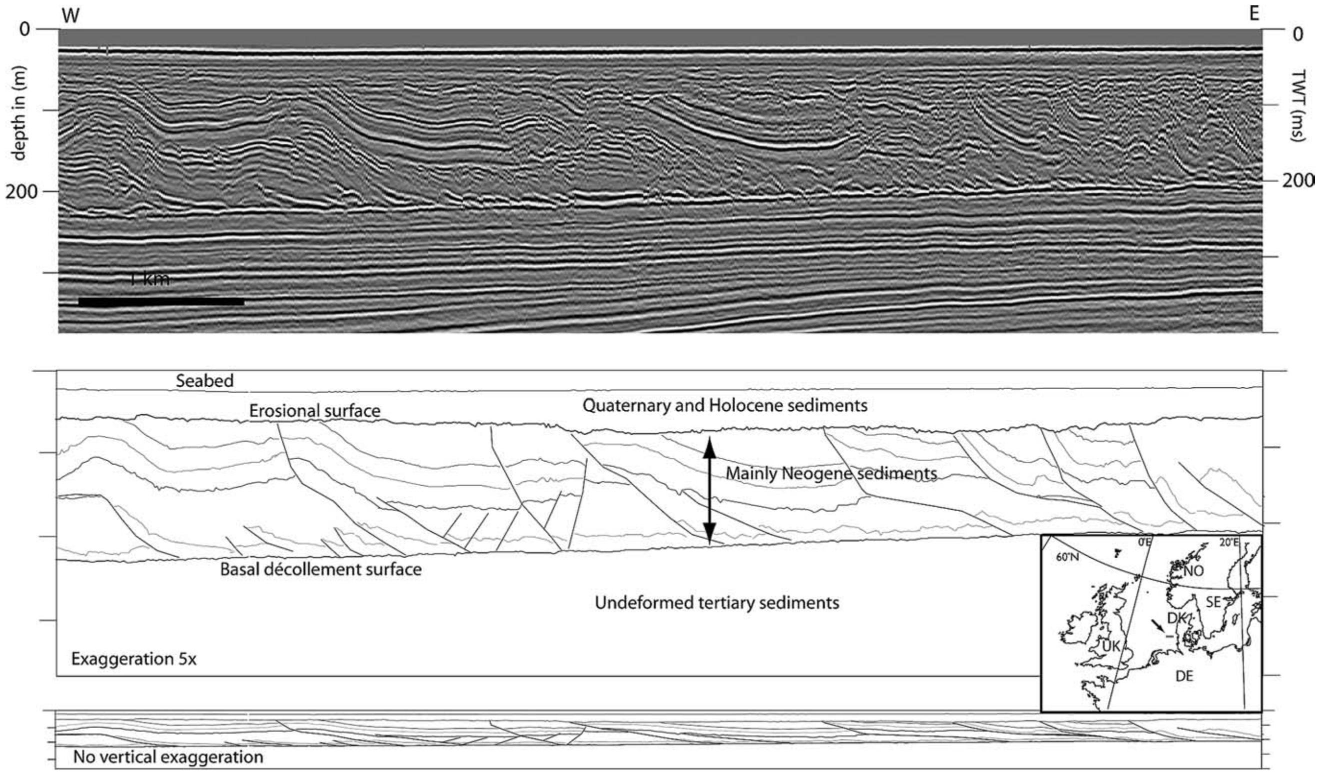

Fig. 1. High-resolution seismic line FL01-05 from the offshore Fanø Bugt glaciotectonic thrust complex. The seismic section reveals a series of thrust-sheets detaching at a weakly inclined basal décollement surface. The Neogene sediments involved in the deformation mainly consist of clays, silts and fine-grained sands (Rasmussen, 2004). Fault inclination is mainly towards the east, although some westward dipping faults do occur.

unconsolidated sediments on top of a 750-m-thick low viscosity layer. This ensures a total decoupling, which is required in order to simulate the conditions at a basal décollement plane like in the Fanø Bugt area (Fig. 1). The base of the model is allowed to move in a horizontal direction only, while the left and right vertical axes are fixed in the horizontal direction. The top of the model is free to move in any direction under the load of the ice-sheet.

3. Numerical model

A differential load with the parabolic shape of an ice-sheet

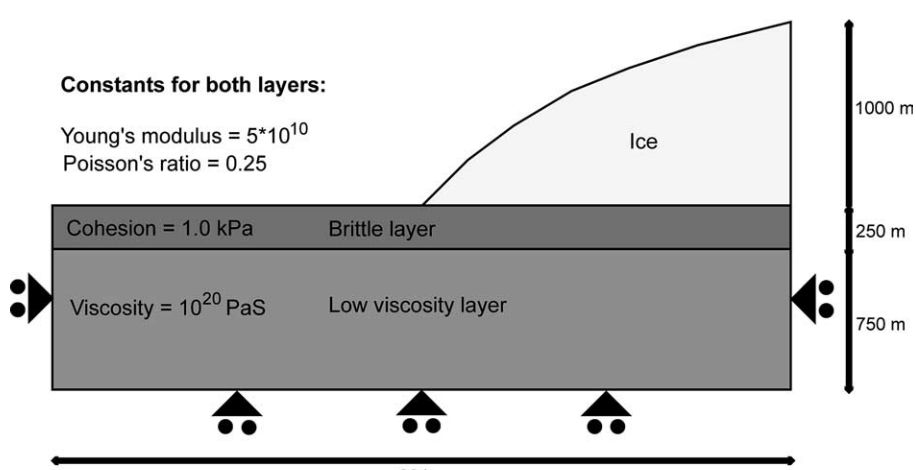

Fig. 2. The principle outline of the model includes a differential load with the parabolic shape of an ice-sheet, 1 km thick at its maximum tapering to zero thickness over a distance of 10 km. The ice-sheet is resting on a sediment pile, 20 km wide and 1 km thick. The sediment pile consists of a 250-m-thick brittle top-layer resting on a thick low viscosity layer that ensures a total de-coupling of the upper layer from its substratum. Small triangles and circles symbolise the boundary conditions: the model is allowed to move in a horizontal direction at the bottom, but not in the vertical direction; at each end the model is fixed in the horizontal direction, while the top of the model is allowed to move in any direction.

placed on a sedimentary substratum is modelled using a dynamic finite element model. The model can manage large strains and rotations combined with visco-elastic-plastic rheologies. The modelling approach described below is chosen because of its ability to represent both long-term elastic stress transfer and permanent plastic deformations in localized zones of failure.

The continuum model is based on convected co-ordinate transformations in a fully Lagrangian frame of reference. The Lagrangian Green strain tensor, η_{ij}, and its work-conjugate Kirchhoff stress tensor, τ^{ij}, are adopted when formulating the principle of virtual work in the reference configuration (Malvern, 1969):

$$\int_{V_0} \tau^{ij}\delta\eta_{ij}\mathrm{d}V_0 = \int_{S_0} T^i\delta u_i \mathrm{d}S_0 - \int_{V_0} \rho_0 g^i \delta u_i \mathrm{d}V_0$$

where $\delta\eta_{ij}$ and δu_i denote virtual strains and displacements, respectively. T^i represents surface tractions, ρ_0 is density and g^i the acceleration due to gravity. V_0 and S_0 are the volume and surface of the model in the reference configuration, respectively.

Visco-elastic behaviour is modelled using a non-linear Maxwell equation (Ranalli, 1995) relating visco-elastic strain rates to deviatoric stress states and effective viscosities while plasticity is modelled using a history and pressure dependent Drücker–Prager yield criterion combined with a non-associated flow rule (Rudnicki and Rice, 1973). The non-associated flow rule enables incompressible plastic flow even though the yield strength is pressure dependent and makes the onset of plastic shear bands possible even in a material strain-hardening environment (Hobbs et al., 1990). The Drücker–Prager yield criterion separating visco-elastic and visco-elastic-plastic stress states is written:

$$f = \sqrt{J_2} + \alpha(\varphi)(1-\lambda_v)I_1 - k(\varphi, C) = 0$$

where J_2 is the second invariant of the deviatoric stress tensor, I_1 the first invariant of the Kirchhoff stress tensor, $\lambda_v \cong 0.4$ the porefluid factor while $\alpha(\phi)$ and $k(\phi, C)$ are functions of the angle of internal friction, ϕ, and cohesion, C:

$$\alpha(\varphi) = \frac{2\sin\varphi}{\sqrt{3}(3-\sin\varphi)}$$

$$k(\varphi, C) = \frac{6C\cos\varphi}{\sqrt{3}(3-\sin\varphi)}$$

The Drücker–Prager yield criterion is a numerically efficient and widely used approximation to the well-known Mohr–Coulomb yield criterion (Gerbault et al., 1998). In particular, the linear dependence of yield strength to pressure is well captured by the Drücker–Prager criterion, as shown in Fig. 3.

During plastic deformation, the modelled material loose load bearing capacity through a decrease in the angle of internal friction:

$$\varphi(\varepsilon_p) = \begin{cases} 30^\circ - 10^\circ \dfrac{\varepsilon_p}{0.01} & \text{for } \varepsilon_p \le 0.01 \\ 20^\circ & \text{else} \end{cases}$$

where ε_p is accumulated plastic strain. This reduction in strength simulates many strain-softening processes such as grain size reduction and pore pressure build-up in fractures. The effect of strain softening is further deformation in plastic zones and thus an amplification of existing localisations.

In our model we operate with an elasto-plastic top layer consisting of unlithified sediments with a comparatively slight cohesion ($C = 1.0$ kPa). The bottom layer is visco-elastic with constant viscosity of 10^{20} Pa S. In both layers Young's modulus is 5×10^{10} Pa while Poisson's ratio is 0.25. The continuum mechanical theory is implemented in an in-house finite element code. For a more detailed description of the numerical model we refer to Hansen and Nielsen (2003).

The differential load is not part of the model, which only includes the brittle sediments and the low-viscosity substratum. However, the top of the model is subject to the differential load through the use of surface tractions.

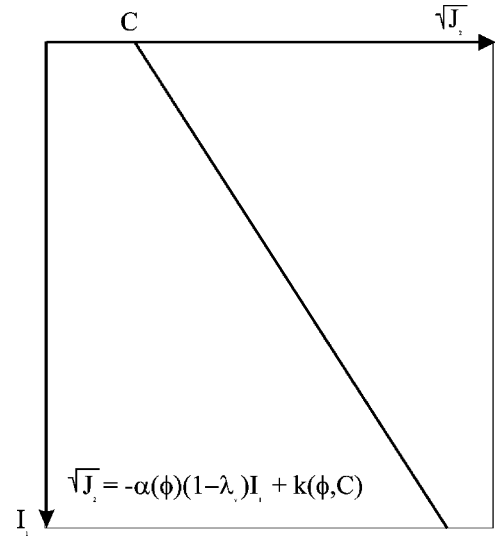

Fig. 3. The graph shows the linear dependence between yield strength and effective pressure of the Drücker–Prager criterion. I_1 is proportional to the effective pressure (the sum of lithostatic and tectonically induced pressure) while $\sqrt{J_2}$ is the effective stress at yield. The pore-fluid factor (λ_v) and cohesion (C) are assumed constant, while the angle of internal friction (φ) is a function of accumulated plastic strain (ε_p).

4. Model results and observations

Dimensions of the outer frame of the model, mechanical boundary conditions, material properties and differential load are identical for all simulations (Fig. 2). However, in order to study the mechanical response of the sedimentary substrate to a differential load like an ice-sheet, the model was run with three different scenarios of depth and dip of the décollement surface.

In the first model a differential load was placed on horizontally bedded, two-layered, 1-km-thick substratum. The brittle top layer had a thickness of 250 m, the underlying low-viscosity layer was 750 m thick (Fig. 2). This simulation shows how strain localisations (potential faults) appear in the foreland of the differential load, as a result of the horizontal stress-field developed by the pressure gradient beneath the load (Fig. 4a). Strain localisations develop in backthrust–forethrust pairs or as conjugate sets throughout the modelled extent of the glacier foreland (Fig. 4a).

The second model was run with a horizontally bedded substratum and a brittle top-layer of 400 m thickness. In this run no localisations occur in the foreland of the load. This result shows that a 400 m top-layer of brittle sediment is too thick a layer to deform by the size of the differential load used in this modelling.

A third model was run with an inclined boundary between the brittle top-layer and the low-viscosity substratum. The boundary dips 0.5° away from the differential load. This is similar to the Fanø Bugt thrust complex (Fig. 1), ranging in depths from 200 m below the ice-sheet snout to 400 m in the far end of the model. In this case the localisations are concentrated close to the ice-sheet where the brittle layer is relatively thin. When the depth to the low-viscosity layer increases, the deformation intensity decreases (Fig. 4b). At some point the brittle layer becomes too thick to deform because the yield criterion and hence the load required for activating a brittle layer is dependent on the depth to the basal décollement surface. With the differential load used in this experiment, the nucleation of strain-bands stops when the décollement is situated at a depth of ca. 380 m (Fig. 4b), which is largely comparable with the limit seen in the Fanø Bugt dataset (Andersen, 2004).

In the first and third models (Fig. 4a and b) the strain pattern in the subsurface of the differential load shows high strain values below the maximum load and low strain values declining to zero strain beneath the point of minimum load.

A close-up of the zone in front of the differential load shows the distribution of localised strain-bands emerging from the décollement surface (Fig. 4). The mean distance between the nucleation of strain-bands in the first model is 432 m varying between a minimum of 150 m and a maximum of 690 m (Figs. 4a and 5). The mean distance is nearly the same in the third model (Fig. 4b); a mean value of 467 m varying between 210 and 630 m. The distance between nucleations seems to be in a random pattern (see Table 1).

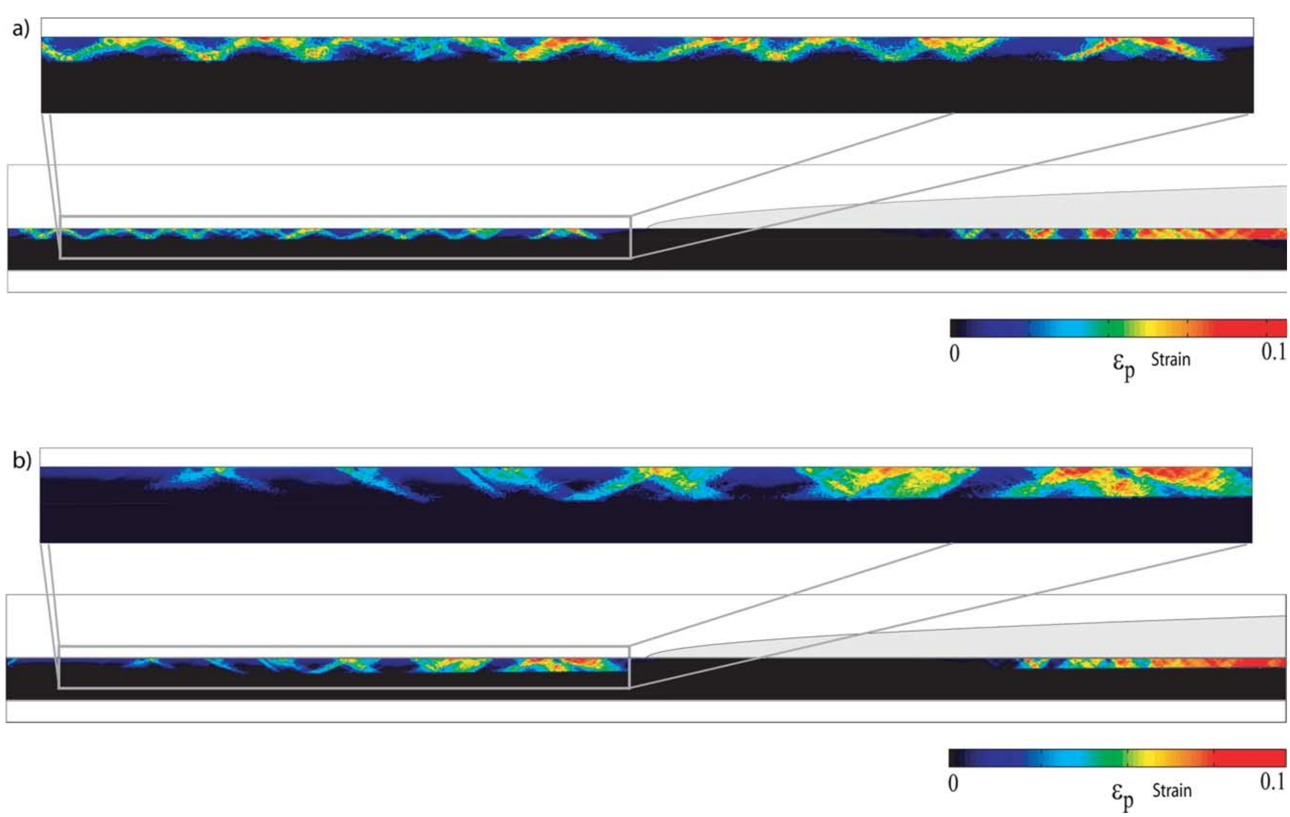

Fig. 4. (a) Strain distribution imposed by the static load of a stationary ice sheet onto the horizontally layered base case model (Fig. 2). The top part is a close-up of the deformation in the foreland of the ice sheet (see text for description). (b) Strain distribution imposed by the static load of a stationary ice sheet onto a model with a 0.5° inclination of the décollement surface. Strain localisation does not occur when the depth to the décollement exceeds 380 m.

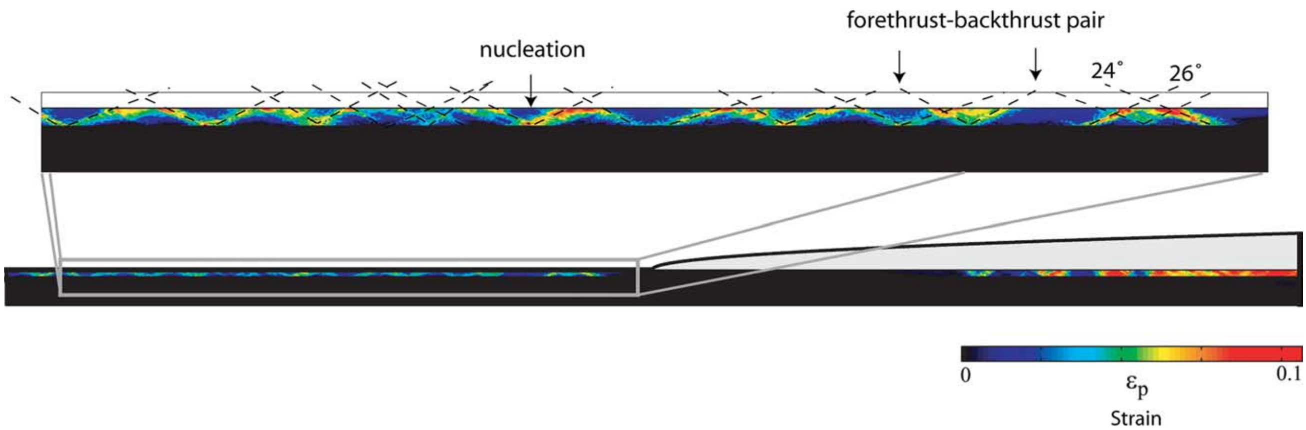

Fig. 5. Example of how the strain localisation/dip analysis is performed. Dashed lines indicate the strain localisations. The distance between points where strain localisations start at the basal décollement surface (nucleation points) is shown in Table 1. The dip between the dashed lines and the basal décollement is displayed in Tables 2 and 3. The definition of a backthrust–forethrust pair is illustrated on the figure.

An analysis of the dip of localisations (Tables 2 and 3) show that when modelling with a horizontal basal décollement surface (Table 2), strain localisations appear in backthrust–forethrust pairs (Fig. 5), with only 1° difference in the average dip of backthrusts versus forethrusts, but 3° average difference between the backthrust and forethrust as a pair. When analysing the model containing a 0.5° dipping basal décollement surface (Table 3) it appears that the development of forethrusts are preferred (57%) as opposed to backthrusts (42%), and not many backthrust–forethrust pairs develop (only two out of six possible). The average angle of forethrusts is 25° compared with an average of 17° for backthrusts, thus a difference of 8° (see Table 3 and Fig. 5 for explanation).

5. Discussion

Modelling of thrust faulting as a response to a push from the rear has been carried out in a variety of ways and serves as an analogue to deformation observed in fold-and-thrust belts (e.g. Borja and Dreiss, 1989; Storti et al., 1997; Cobbold et al., 2001; Strayer et al., 2001; Costa and Vendeville, 2002; Panian and Wiltschko, 2004). Modelling of thrust faults by exerting a differential load onto the substratum is another matter even though the mechanical response may be the same (cf. Strayer et al., 2001). We suggest that the latter is more relevant for the study of large-scale glaciotectonic deformation involving sediment sheets, hundreds of metres thick and kilometres long.

5.1. Glaciostatic stresses and strain localisations

The strain pattern observed in the sedimentary substratum due to a differential load can be explained by a rotation of the deviatoric stresses beneath the load; the vertical stress decreases and the horizontal stress increases from the point of maximum load to the point of zero load (Fig. 6) (Aber et al., 1989). Hence at one point the deviatoric stresses equal each other and a hydrostatic condition arises (see Fig. 6). At this point there is no deformation of the substratum (Figs. 4 and 6). Beyond the differential load the

Table 1
Distribution of strain localisations from the ice-margin towards the foreland for the two models shown in Fig. 4

Number of nucleations from the ice margin towards the foreland	Distance between nucleations (m) on the model with horizontal basal décollement	Distance between nucleations (m) on the model with a 0.5° dipping basal décollement
1	330	270
2	360	450
3	300	510
4	540	540
5	330	300
6	570	210
7	660	540
8	600	630
9	360	420
10	150	420
11	180	630
12	360	480
13	510	
14	690	

Table 2
Inclinations of strain localisations, sorted in backthrust–forethrust pairs for the model with a horizontal basal décollement surface (Fig. 4a)

Backthrust angle (°)	Forethrust angle (°)	Difference between backthrust and forethrust angles (°)	
	24		
31	25	6	
26			
32	26	6	
21	24	3	
28	29	1	
25	30	5	
22	30	8	
	28		
27	27	0	
25	24	1	
29	29	0	
25	23	2	
24	29	5	
26	27	3	Average angle (°)

maximum stress above the décollement is horizontal leading to compressional deformation (Fig. 4).

When comparing the model of a horizontal décollement (the horizontal case) with the model with a dipping décollement (the dipping case), there is no big difference in the distance between nucleation of strain localisation (Table 1). There is, however, a difference in the number of strain localisations; 14 in the horizontal case and only 12 in the dipping case. This difference is due to the fact that no strain localisations form when the depth to the décollement exceeds 380 m as it does in the distal end of the inclined-detachment model (Fig. 4b). When comparing the results of the strain-band-dip analysis (Tables 2 and 3) the effect of the 0.5° detachment-inclination is pronounced, such that the development of forethrusts is preferred instead of backthrusts and that backthrust–forethrust pairs are rare compared with the horizontal case. Also the dip-angles of backthrusts and forethrusts are markedly different in the dipping case, while there is nearly no difference in the dip-angles in the horizontal case (Tables 2 and 3). Hence it could seem like the formation of backthrusts 'up-hill' is difficult, even for small dips of the basal décollement surface away from the source of compression. This is consistent with the observations from Fanø Bugt, which show a prevalence of eastward dipping forethrusts, with subordinate backthrust development.

When comparing the results of this modelling study with the glaciotectonic setting, it is evident, as others have described before (e.g. Rotnicki, 1976; van der Wateren, 1985; Aber et al., 1989; Hart, 1990), that the glaciostatic load causes extension beneath the ice-sheet and compression in front of the ice. Our modelling study is, however, the first to show that strain localisations (potential thrust faults) may form far beyond the snout of the ice sheet, and to depths of hundreds of metres, only dependent on the location of a décollement at a suitable depth relative to the load of the ice sheet.

5.2. Importance of décollement horizons

The importance of a décollement horizon at a suitable level in the subsurface has been realised by several previous

Table 3
Inclinations of strain localisations, sorted in backthrust–forethrust pairs for the model with an inclined basal décollement plane (Fig. 4b)

Backthrust angle (°)	Forethrust angle (°)	Difference between backthrust and forethrust angles (°)	
25			
	25		
	29		
24			
26			
22	26	4	
	30		
24	17	7	
	35		
	18		
	20		
13			
17	25	6	Average angle (°)

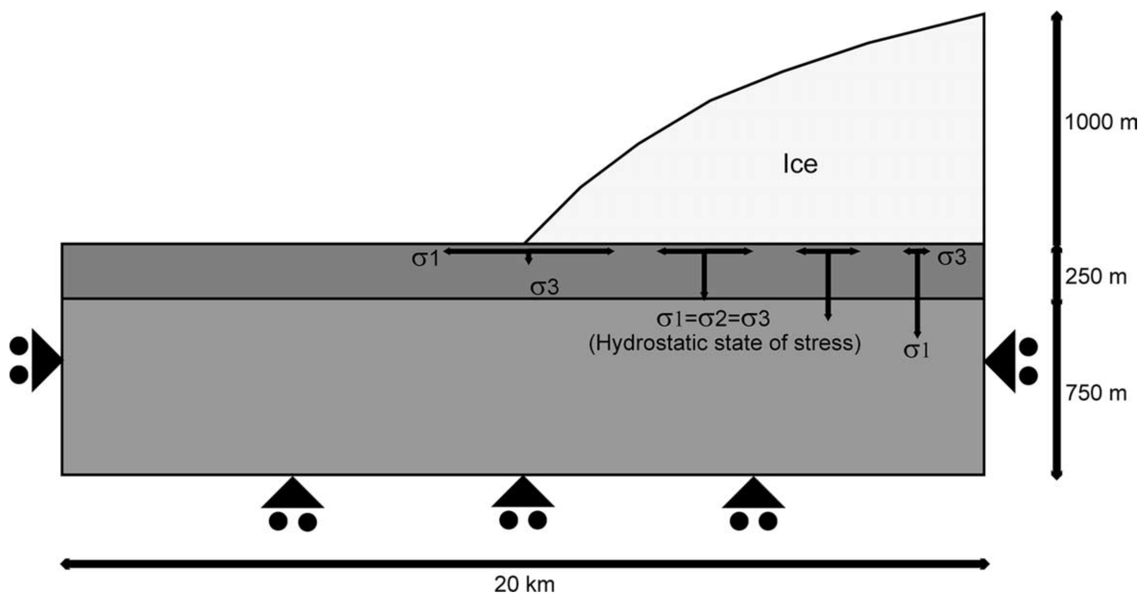

Fig. 6. Sketch showing the rotation of the deviatoric stresses beneath the differential load.

studies (e g. van der Wateren, 1985; Berg and Beets, 1987; Huuse and Lykke-Andersen, 2000; Bennett, 2001). The presence and efficiency of décollement surfaces are determined by the rheological properties of the substratum and circumstances that alter it such as permafrost or high porewater pressures (Boulton and Caban, 1995; Boulton et al., 1999). Porewater pressure and flow is important in the mechanics of thrusting, as it facilitates thrusting by reducing shear strength in fault zones (Hubbert and Rubey, 1959). The flow of porewater also plays an important role in the glaciotectonic setting by elevating groundwater pressures and thereby facilitates the formation of décollement surfaces (Mathews and Mackay, 1960; Boulton and Caban, 1995; Boulton et al., 1999).

5.3. Effects of pore pressure

We acknowledge that porewater pressure and flow may be important factors affecting thin-skinned thrust tectonics, but in this first attempt to model thrust faulting as an analogue to glaciotectonic deformation we have modelled the pure gravitational response of an unconsolidated sedimentary substrate to a differential load with the shape of an ice sheet.

Although not modelled, it may be surmised that incorporation of porewater flow in the model would enhance the observed strain patterns, making the weak zones even weaker. This is suggested by a recent numerical modelling study (Strayer et al., 2001), which showed that fluid flow focused in zones of high shear strain-rates, reducing the pressure inside shear-zones and elevating pressure outside. Were this to occur in a glacial setting, the high strain zones in the foreland might eventually weaken to an extent where the mechanical coupling between the ice sheet and a point beyond its snout breaks down, thus halting the further development of thrusting beyond this point.

The porewater pressure beneath the ice-sheet snout may be so high that it reduces the effective load of the ice-sheet to zero. If this happens, the compression zone moves up-glacier to the point where the load of the ice-sheet again is being exerted on the subsurface. In this situation compressional deformation will happen below the ice-sheet margin, as has been observed in some modelling studies (Boulton and Caban, 1995). Thus the way in which the ice-sheet is coupled to its bed seems important for where the formation of glaciotectonic thrusting may occur, as also pointed out by e.g. van der Wateren (1995) and Bennett (2001).

5.4. Effects of ice-sheet movements

The fact that ice moves cannot be ignored, but the shear stress that are transferred from the moving ice-sheet to the subsurface are generally considered to be small compared with the stresses exerted by the load of the ice sheet (Aber et al., 1989). Hence the shear stress from the movement of the ice-sheet over its ground may be of minor importance for large-scale structures. Movement of the ice sheet is, however, important because the movement transfers the load of the ice laterally, thus imposing different stress conditions onto the sedimentary substrate. As an ice sheet moves relative to a point in the substratum it will first experience horizontal compression (in the foreland), then hydrostatic stress (below the ice-sheet snout), then horizontal extension (below the thick part of the ice sheet). As the ice sheet retreats, the same point will eventually experience compression again (cf. Hart and Boulton, 1991). The changing stress conditions and revival of deformation patterns might be the cause of progressive glaciotectonic deformation (Pedersen, 1987) where glaciotectonic thrusting is achieved in several phases of fracturing and faulting of the subsurface sediments. It might also be the reason why many large-scale glaciotectonic complexes appear to have formed during net retreat of the ice-sheet from the maximum extent, maybe followed by small re-advances (Hart, 1990; Boulton et al., 1999). Exposing the same sedimentary substratum to heavy compression twice or several times strongly enhances the chance of thrusting in the localised strain zones.

5.5. Summary of discussion

The modelling results indicate that the combination of ice-sheet load and depth to the décollement surface are primary controls on large-scale glaciotectonic thrusting (cf. van der Wateren, 1985). However, a subtle interplay between ice-sheet load, décollement depth, substrate rheological properties and pore pressure is needed to obtain the conditions leading to thrust formation. The modelling results indicate that bulldozing by an advancing glacier is not a pre-requisite for large-scale glaciotectonic deformation. Bulldozing by an advancing ice sheet is, however, still the most likely mechanism causing the formation of imbricate push moraines on a scale of tens of metres or less in thickness. The lateral motion of ice-sheets is important for large-scale deformation by varying porewater pressure in the substrate (Croot, 1988), and by superimposing deformation caused by varying stress regimes.

6. Conclusions

The modelling results show that thrust nucleation in an unconsolidated sedimentary substratum exposed to a static differential load can be modelled by means of a dynamic finite element model.

The subsurface stress pattern induced by the differential load displays a well-defined transition from a horizontal extension beneath the maximum load, over a zone where effective stresses decrease to zero in the transition zone of intermediate load, to a zone of horizontal compression from the point of minimum load and beyond. Compressive strain localisations (potential thrust faults) develop in response to the compressional regime far beyond the edge of the differential load. The obtained strain pattern is interpreted to be due to a rotation of the deviatoric stresses beneath the differential load by the mechanism of gravity spreading. A prerequisite for this strain pattern to develop is the existence of a décollement horizon at a suitable depth relative to the differential load.

The analysis of the distribution and dip of strain-bands showed that even a small dip of 0.5° of the décollement surface away from the differential load caused difficulties for backthrust formation. Furthermore, it was evident that strain localisations stopped developing when the basal décollement exceeded 380 m depth.

Interpreting the model as an analogue to the mechanics of thrusting in the sedimentary substrate of an ice sheet, it is evident that large-scale glaciotectonic thrust structures may form by gravity spreading alone without the need for a push from the rear by a moving ice sheet. The static load of the ice-sheet alone is thus sufficient to cause thrust nucleation, subject to the presence of an effective décollement horizon at a suitable depth relative to the effective ice-sheet load. Hence, it is concluded that the mechanism of gravity spreading may be responsible for the formation of large-scale (hundreds of metres) glaciotectonic thrusting as observed in Fanø Bugt and elsewhere. More advanced modelling studies should aim to explore the complexities of the subglacial environment by including parameters such as fluid pressure and flow, a variety of ice-sheet models, isostasy and motion of the ice.

Acknowledgements

The geological Survey of Denmark and Greenland and the Danish Research Agency financed the PhD-scholarship of LTA. Holger Lykke-Andersen and Peter Roll Jakobsen are thanked for valuable discussions. We would like to thank Ole Vejbæk, reviewers S. Hardy and E. Tavernelli and editor J. Hippertt for insights and helpful comments to the manuscript.

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