The glacial rafts at Clava

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The Site of Special Scientific Interest at Clava [NH 766 442] is located at an altitude of about 150 m OD, 9 km east of Inverness (Gordon, 1993; Merritt et al., 2017). It includes several natural sections along the lower reaches of the Cassie and Finglack burns, both tributaries of the River Nairn (Fig. 83). The sections are cut into north-east to south-west-trending benches that slope gently both downstream and towards the centre of the Nairn valley, with the lower ones having been modified by glacial drainage in an ice marginal setting. The deposits, proved in sections and in boreholes, form a complex glacigenic succession that includes well-documented occurrences of 'Arctic' marine shelly clay. The latter, first described by Fraser (1882a, b) in a former clay pit [NH 7658 4411], excited considerable controversy in the Nineteenth Century. Some regarded it to be an *in situ* deposit thus supporting a ‘great submergence’ of the country during the Pleistocene, whereas others concluded that it had been transported by ice from offshore.

In view of the controversy over the origin of the shelly clay and its role in the whole concept of glacially-related submergence in northern Britain, a Committee of the British Association for the Advancement of Science was convened to carry out investigations on the deposits in 1892 (Horne et al.,1894). They made two excavations near the old clay pit (the 'Main Pit') and sank seven boreholes (Gordon, 1990). The clay was confirmed to be a marine deposit extending laterally for at least 170 m and reaching a maximum thickness of 4.9 m. It was essentially horizontal and had sharp contacts with adjacent beds above and below. There was little sign of disturbance, although cracks and fissures were noted. The shells are of shallow-water marine species representing an Arctic or sub-Arctic faunal assemblage. They were unscratched and generally well preserved with intact periostraca.

Clava was re-investigated and described by Peacock (1975). Although the original sections were no longer exposed, fresh ones had appeared along the Cassie Burn a few hundred metres to the south-west (Fig. 83, sites I-VI). In a stream-bed exposure (V), Peacock discovered diamict with fragments of marine shells (*Clava Shelly Till*) including an unabraded paired valve of the fully Arctic bivalve *Portlandia arctica*, which had not been recorded previously at Clava. The locality was investigated again in 1988 when additional exposures were logged (Merritt, 1990).

The glacigenic sequence.

For reasons of continuity the lithostratigraphy described below is that established by Merritt (1992): alternative names have been proposed by McMillan et al. (2011) (Table 4).

Table 4. Lithostratigraphy at the sites at Clava located in Fig. 83 (after Merritt, 1992).

	Unit	MP	FS	III	V	VI	< (m)
1	Morainic deposits					+	4
2	Finglack Till	+	+	+		+	10
3	Clava Gravel			+			8
4	Clava Sand	+	?	+			6
5	Clava Shelly Clay	+	?				4.9
6	Clava Shelly Till			+	+		3
7	Drummore Gravel	+	+				5
8	Cassie Till	+		?			7
9	ORS bedrock	+		near			-

The Clava site lies mainly on bedrock of pebbly sandstones and breccio-conglomerates of the Inverness Sandstone Group, part of the Middle Old Red Sandstone. The basal *Cassie (Suidheig) Till Formation (Fm)*, known only from the British Association boreholes, is probably a subglacial traction till. The overlying *Drummore Gravel Fm*, proved in boreholes and exposed at the 'Finglack Section' (Fig. 83 FS), comprises a weathered, stratified, matrix-supported, gravelly silty sand diamict with clasts mainly of sandstone and siltstone. It was probably formed as subaerial sediment gravity flows, in an ice-marginal or supraglacial environment.

The allochthonous units of the 'Clava Shelly Formation' include at least three members; the unfossiliferous *Clava Sand*, the *Clava Shelly Clay* and the *Clava Shelly Till*. The first two members form a conformable coarsening-upward sequence at the Main Pit (Fig. 83 MP & 84, dipping gently to the west. The Shelly Till is overlain unconformably by the Clava Sand at Section III (Fig. 85; Table 4). It is possibly a glacially disrupted, glaciomarine pebbly clay, but its origin, stratigraphical relationship and age are not absolutely clear. The Clava Shelly Till typically comprises stiff, matrix-supported, silty sandy clay diamict containing clasts of mainly micaceous gneiss, with some granite and sandstone. The sparse fauna of the diamict is interpreted as having inhabited a shallow-marine environment in fully Arctic conditions during a stadial episode (Graham, 1990; Graham et al., 1993). A broken, unabraded shell fragment, probably *Mya truncata,* yielded an infinite AMS radiocarbon date of >46,400 ^{14}C yrs BP (Merritt, 1992). The deposit has been folded at Section III, where a plane of glacitectonic décollement separates it from the overlying Clava Sand, which has also been deformed (folded) (Fig. 85).

The upper 3 m of the Clava Sand is a very compact silty fine to medium-grained sand with sparse 'lonestones'. It is generally pale to moderate yellowish-brown in colour, but in places it is dusky yellow with yellow and orange tints. At the Main Pit it shows poorly defined sub-horizontal lamination and includes poorly defined seams (up to 5 mm thick) of medium bluish-grey fine sandy silty clay and thin, ill-defined seams of fine sandy silt containing well-rounded granules and fine pebbles. The pebbles are mainly of metasandstone (psammite), but a single clast of decomposed micaceous siltstone has been recorded. The lower 3 m comprises a sequence of graded beds of moderate yellowish brown, micaceous silty fine-grained sand, each bed being about 4 cm thick with gradational contacts. The beds are stacked into crude packages, possibly varves, each being about 40 cm thick. The rhythmic cyclicity may indicate annual freezing within a marine fjord environment in which the Clava Sand was deposited (Merritt, 1992).

The basal contact of the Clava Sand is gradational at the Main Pit, picked out by a sharp change of colour that cross-cuts the bedding and descends gently towards the Cassie Burn. The colour boundary is clearly related to the position of the permanent water table and it is evident that the sand and the underlying shelly clay form a continuous coarsening-upwards sequence.

The lowermost 40 cm of the Clava Shelly Clay observed during the re-opening of the Main Pit for the QRA in 1990 comprised pale grey plastic clay, but the base of the deposit was not seen. The unit was overlain conformably by 60 cm of very stiff, silty clay containing a network of black carbonaceous streaks, closely spaced shears, and micro-faults that ramify through the deposit. It was overlain by poorly defined, graded beds of dark grey silty clay, clayey silt and silty very fine-grained sand. The lithologies of the clasts are as reported by Horne et al. (1894), importantly these are dominated by a micaceous gneiss of a type not common south of the Great Glen, with only 17 percent of locally derived Old Red Sandstone.

Shells, chiefly molluscs, were most commonly observed in the uppermost metre of the Shelly Clay during the re-excavation with the range of faunas being consistent with that described by Horne et al. (1894). There were, however, many laterally discontinuous beds (1 to 5 cm thick) with scoured bases containing angular to sub-rounded pebbles up to about 30 mm in diameter and broken shells only. These diamictic beds, not previously noted, were an integral part of the original (primary) sedimentary sequence, possibly representing mass flows resulting from storms. Intact shells were generally located within the layers between the diamictic beds, which also contain well-dispersed subrounded to well rounded pebbles, some with an epifauna including barnacles. The pebbles may have dropped from floating ice, but transportation by floating seaweed is also a possibility (cf. Gilbert, 1990).

Modern palaeontological analyses confirm that the Clava Shelly Clay formed in a shallow-marine environment in a high-boreal to low-Arctic interstadial climate with conditions apparently becoming increasingly estuarine during its deposition (Graham, 1990; Graham et al., 1993). The dinoflagellate cyst record suggests that there may have been two short, warmer interludes and, on the basis of calcareous nannofossils, the deposit can be dated tentatively as being no older than about 77 000 years. Amino-acid dating on specimens of *Littorina littorea* indicates that the Clava Shelly Clay is Middle Devensian in age and the results of radiocarbon accelerator dating confirm that the deposit is not younger than this age (Merritt 1992).

The upper units of the sequence are described below in the individual site descriptions.

Main Pit Section [NH 7658 4411]

The type section of the 'Clava Shelly Formation' (Fig. 84) is now mostly obscured and overgrown, but the contact between it and the overlying Finglack Till is periodically revealed in the back-scarp of a landslide that can be scraped clean. The top of the Clava Sand comprises yellowish brown, massive to indistinctly laminated silty fine to medium-grained sand with thin lenses of coarse-grained sand with scoured bases, and sparse dropstones. The unit is cut by a number of bedding-parallel to steep (60-80º) east-north-eastward dipping veins filled by finely laminated to brecciated sand, silt and clay (Fig. 86 a). Individual sand and silt laminae (< 1.5 cm thick) within the veins are graded, some containing intraformational rip-up clasts of silt and clay. The veins (< 15 cm thick) can be traced laterally for several metres and most taper upwards. They are interpreted to have formed as hydrofractures similar to those described by Phillips and Merritt (2008).

Close to the top of the sequence, the bedding within the Clava Sand is deformed into several moderately to gently inclined folds and thrusts. These deformation structures are locally truncated by the hydrofractures, indicating that folding and thrusting pre-dated the formation of these water-escape features. At the base of the overlying Finglack Till, the sand is very highly deformed with bedding having been variably transposed by a glacitectonic foliation. Parallel to the planar, sub-horizontal base of the till is a 40 cm thick zone of highly deformed, over-consolidated, silty fine-grained sand containing boudinaged lenses of diamict, stringers of coarse sand and plastically deformed fragments of clay. The hydrofractures present in the Clava Sand below are clearly truncated by this high strain zone (Fig 86 a,b). This relationship indicates that hydrofracturing occurred prior to ductile shearing that accompanied the deposition of the Finglack Till.

Cassie Burn Section III [NH 7638 4371].

The rafted sequence at this section is deformed by a number of thrusts that truncate bedding and earlier developed folds (Fig. 85). The Clava Shelly Till and Clava Sand in the lower part of the section are overlain by several metres of deformed (thrusted) sand and gravel identified as glaciofluvial ice-contact deposits by Merritt (1992), but named as the *Clava Gravel Member* by Phillips and Merritt (2008) (Table 4). The Clava Gravel comprises predominantly waterlain, poorly sorted, clast-supported gravel interbedded with pebbly, medium to coarse sand. Significantly, the gravel clasts are mostly composed of an allochthonous suite of lithologies composed of micaceous metasandstone with subordinate semipelitic gneiss, brown sandstone and granitic gneiss. The basal contact of the Clava Gravel is a gently north-eastward-dipping thrust (Fig. 85a). A small splay from this thrust deforms the upper part of the structurally underlying Clava Sand. The tip of this thrust is marked by a tight, isoclinal fold, which is itself truncated by a later reverse fault (Fig. 85c). A gently to steeply, north-eastward dipping 'high strain zone' (c. 50 cm thick) near to the base of the Clava Gravel (Fig. 85 d) is composed of highly deformed, laminated clay and silt containing lenses or boudins of sand and diamict (Shelly Till).

The Clava Shelly Till is exposed at the base of the section (Fig. 87) and in the river-bed, both here and downstream to the road bridge (sections IV & V). It is an olive grey, stiff, fissile, matrix-supported, silty clay-rich diamict containing clasts of micaceous metasandstone, quartzite and gneiss, with lesser amounts of red granite and red and white Devonian sandstone. Apart from the clasts of sandstone, which are generally angular, others are rounded to well-rounded, moderate to high sphericity, and polished, similar to the distinctive, allochthonous suite described in the Shelly Clay at the Main Pit by Horne et al. (1894). The diamict includes some discrete masses of fissile, dark bluish grey clay containing comminuted shells (including small gastropods) and sparse stones. A crudely developed layering in the diamict is deformed by a southward verging, north-eastward plunging and asymmetrical fold (Fig. 85 a). Although it is tempting to suggest that the discrete masses of grey clay within the Clava Shelly Till at this section may have been cannibalized from the Clava Shelly Clay during transport to the site, the faunal assemblages suggest otherwise (Peacock, 1975).

At this location both the Clava Shelly Till, and the structurally overlying Clava Sand, are separated by a gently dipping to sub-horizontal thrust. This thrust truncates a large-scale, monoclinal fold that deforms, and locally overturns bedding within the sand (Fig. 85 & 87). This relationship indicates that folding pre-dated thrusting. However, this does not necessarily imply that folding and thrusting represent completely separate deformation events. The Clava Sand forms a thrust-bound wedge that thins rapidly north-eastwards where it is cut out against the thrust at the base of the overlying Clava Gravel (Fig. 85 a). Soft-sediment deformation structures are common in the finer grained, lower part of the Clava Sand, indicative of locally high pore-water pressures/contents during deformation (Fig. 88). Bedding in the sand unit is also locally offset by small-scale, steeply dipping (50º-60º), conjugate normal faults, which are typical of rafts observed elsewhere in NE Scotland (Peacock and Merritt, 2000; Merritt and Connell, 2000).

Cassie Burn Section VI [NH 7645 4390].

This section, which occurs at the top of a steep, gorse-covered slope, exposes several metres of sandy diamict capped by very poorly sorted cobble gravel (Table 4; Fig. 89). The diamict is irregularly stratified, matrix-supported and includes thin, laterally discontinuous, graded, branching, cross-cutting and anastomosing seams and laminae of finely laminated silt, sand and fine gravel (Fig. 90). There is some evidence of re-sedimentation by fluvial processes in the form of 'rip-up' clasts together with penecontemporaneous slumping, scour-and-fill, cohesive mass-flows, 'pull-apart' structures and cracks. A lens of clast-supported gravel with a sharp, horizontal base and an arched top was formerly exposed (Fig. 89). Lenses such as these are thought to be indicative of subglacial fluvial deposition within channels (canals) in the soles of warm-based ice sheets (Benn and Evans, 2010). The diamict is more massive in the lower half of the section, where beds up to 0.5 m thick are interbedded with seams of sand (up to 10 cm thick) with scoured bases. Even the more massive beds of diamict typically include sub-horizontal wisps of white silt and fine-grained sand. Peacock (1975) interpreted the sequence as a supraglacial flow-till complex, whereas Merritt (1990) suggested that the deposits were more likely to have been formed by subglacial processes. The ramifying network of interconnecting to cross-cutting seams filled by finely laminated sand, silt and clay have now been identified as hydrofractures (Phillips and Merritt, 2008) (Fig. 86 c).

Finglack Section [NH 7688 7422].

Up to about 3 m of crudely stratified, very poorly sorted and silty gravelly diamict containing pebbles and large blocks of Devonian sandstone lies at the base of this abandoned, degraded river cliff section (Fig. 91). The unit (Drummore Gravel Fm) (Table 4) includes water-winnowed laminae and is interpreted as a debris flow deposit (Merritt, 1992). It is overlain by folded and thrust-bound slices of sediments matching those of the Clava Shelly Fm. The sharp, planar contact between the two units represents a glacitectonic décollement surface that probably developed at the base of the rafted sediments (Phillips and Merritt, 2008). The clay, silt and sands immediately above this thrust are similar in appearance to the upper part of the Clava Shelly Clay at the Main Pit, but are highly deformed (a glacitectonite *sensu* Benn and Evans, 1998) and show evidence of several phases of folding and faulting, as well as soft-sediment deformation (Fig. 91 b). This glacitectonite is overlain by dark grey pebbly diamict similar to the Clava Shelly Till exposed at Cassie Burn section III, which is in turn separated from the overlying Finglack Till by a prominent, gently south-eastward dipping thrust zone. Individual thrust planes within this zone are marked by thin layers of hard, sheared silty clay, with the upper and lower bounding thrusts being separated by a high strain zone composed of highly sheared clayey sand.

The rafted sediments are cut out towards the south-east of the section, where the Finglack Till rests directly upon the Drummore Gravel (Fig. 91). The till comprises a thick (at least 8 m), massive unit of yellowish brown, stiff, matrix-supported, stony diamict with a matrix of clayey fine-grained sand. Like many tills overlying Devonian rocks in the vicinity, it contains mainly faceted and striated clasts of sandstone and micaceous siltstone, many of which can be matched with strata cropping out immediately to the west or southwest of the site. However, the basal few metres of the till contains relatively abundant, well-rounded, high-sphericity and commonly polished pebbles of allochthonous micaceous metasandstone that have probably been cannibalised from sediments similar to those forming the rafts. The basal contact of the till constitutes a high strain zone similar to the one occurring at the Main Pit section described above.

Interpretation

The majority of the British Association Committee concluded that the Clava Shelly Clay in the Main Pit was *in situ*, indicating former submergence of the land up to about 150 m above OD (Horne et al., 1894). As evidence, they cited the assemblage of organic remains, their mode of occurrence, the extent of the deposits and their apparently undisturbed character. A minority of the Committee (Bell and Kendall), however, argued that there was insufficient evidence to reach a firm conclusion and doubted that there was any substantial evidence at all in Scotland for a great submergence. They questioned the widespread absence of shell beds and other traces of submergence and the lack of marine organisms in the overlying till. Although acknowledging certain difficulties, notably the extent of the deposit and the good preservation of the shells, they favoured an ice-rafted origin for the shelly clay, with a source area in Loch Ness, as indicated by ice-movement patterns inferred from striae and erratics.

Peacock (1975) concluded from his investigations at Clava that the shelly clay had been transported by ice and was part of an autochthonous melt-out till comprising reworked sea-floor material. Sutherland (1981), however, accepted an *in situ* origin for the shelly clay and presented a model relating these deposits to glacio-isostatically induced submergence in front of an expanding ice sheet. However, the deposits at Clava fit only partly into the overall distribution pattern of the high-level shell beds in Scotland and are apparently at too great an altitude to be fully explained by this model.

Following the re-opening of the Main Pit in 1990 and a critical re-examination of the sections described by Peacock, there is now unambiguous evidence to indicate that the shelly deposits at Clava have been deformed by glacial processes and a very strong convergence of evidence that they, like all other 'high-level' shelly clay occurrences around the Scottish coasts, considered by Sutherland (1981), have been transported glacially as rafts (Merritt, 1992; Phillips and Merritt, 2008; Merritt et al., 2014; Finlayson et al., 2010, 2014). The raft exposed at the localities investigated in the 19th Century has to a large extent maintained its integrity during transport, comprising an internally conformable, coarsening-upward sequence.

The smaller masses of shelly diamict (Clava Shelly Till) occurring upstream of the Main Pit are also interpreted to be rafts, but their Arctic fauna demonstrates that they are not simply lateral equivalents of the shelly clay in the main raft. The composition of the clast assemblage, both in the diamict and in the Clava Gravel at Cassie Burn section III, suggests that both units originated in the Loch Ness Basin. Merritt (1992) concluded that the rafts were all probably detached as a result of high pore-water pressure building up in laterally restricted aquifers beneath a confined glacier that flowed north-eastwards across the Loch Ness Basin (also see Phillips & Merritt, 2008). This glacier was deflected eastwards and upwards towards Clava by ice flowing from the northern Highlands through the Beauly Firth Basin during the build-up of the last Scottish ice sheet (Fig. 92). The rafts were stacked at the ice margin when the glacier entered the Nairn Valley before being overridden by the expanding ice sheet.

A revised model for the rafting at Clava

The dimensions and general sub-horizontal aspect of the assemblage of rafts exposed at Clava, coupled with the associated thrusting and large-scale folding of these bodies, are typical of many glacially-transported rafts described in the literature (see Stalker 1976; Moran et al. 1980; Bluemle & Clayton 1984; Aber 1985, 1989; Ruszczynska-Szenajch 1987; Broster & Seaman 1991; Benn & Evans 1998; Burke et al., 2009; Vaughan-Hirsch et al., 2013). In North America, large-scale rafts, 'megablocks' or 'pancake rocks' occur within areas of glacial-thrust terrain (Moran et al. 1980; Ruszczynska-Szenajch 1987), where the thrusted rafts of bedrock have been transported near to the base, or in front of the advancing glacier. It was generally thought that rafts were detached and transported whilst frozen to the base of cold-based ice sheets, melting out slowly following stagnation and decay of the enclosing ice. More recent studies, however, have suggested that failure, leading to detachment, is associated with elevated pore-water pressures that may occur along water-rich décollement surfaces within the subglacially deforming layer (Moran et al. 1980; Aber 1985; Broster & Seaman 1991; Benn & Evans 1998; Phillips & Merritt, 2008; Burke et al., 2009; Vaughan-Hirsch et al., 2013), at the base of the deforming layer (Kjær et al. 2006), or as a consequence of subglacial hydrofracturing by forceful upward dewatering (Boulton and Caban 1995; Rijsdijk et al. 1999; Phillips & Merritt, 2008). Aber (1985) concluded that unless soft sedimentary rafts remain deeply frozen they will be distorted and ultimately to lose their identity due to subglacial shearing and/or proglacial deformation. These frozen rafts, therefore, should only survive transport for a limited duration as they are liable to thaw, break apart and homogenise during transportation. Indeed, the irregular masses of clay observed in the Clava Shelly Till at Cassie Burn section III could have formed in this way. However, in general there is little evidence at Clava to support that the rafts of shelly material were frozen at any time during their history of entrainment, transport and final accretion/deposition. Macroscopic field observations and micromorphological evidence presented by Phillips and Merritt (2008) indicates that liquefaction, hydrofracturing and water-escape occurred repeatedly during the deformation history recorded by these sediments in response to fluctuating pore-water pressures. This evidence, coupled with the presence of large-scale detachments within the sequence, has resulted in a purely glacitectonic model for the rafting and emplacement of the marine sediments at Clava.

The assemblage of rafts exposed at Clava appear to occur within an envelope that is in the order of 900 m long by 300 m wide, and a few tens of metres thick, with its upper and lower boundaries formed by prominent décollement surfaces marked by high strain zones (Fig. 92). Internally, the envelope is dissected by a number of gently dipping, south-east-directed thrusts, with thrusting resulting in the imbrication and localised excision of parts of the sedimentary sequence (Fig. 93). The most intense deformation is observed within the Finglack section and Cassie Burn section III, which occur near to the margins of the envelope. The sequence at the Main Pit occurs towards the centre of the raft and is, therefore, less disturbed, recording the conformable stratigraphical relationship between the Clava Shelly Clay and Clava Sand.

The earliest recorded phase of deformation resulted in folding and thrusting. Thrusting was partitioned in the weaker clay-rich layers. Several generations of folds have been recognised recording the initial shortening of the sediment pile related to ice-push, prior to thrusting as the glacier continued to impinge on the sediments, followed by a later phase of folding in response to continued shortening within the hanging walls of the thrusts. Folding and thrusting were followed by several phases of liquefaction and water-escape, leading to the development of a locally complex network of hydrofractures. Liquefaction appears to have been accompanied by continued movement along the thrusts with these structures acting as a focus for water-escape. The repeated injection of fluidised sand and silt along the thrusts forming the margins of the envelope is thought to have aided transport, essentially lubricating these décollement surfaces.

Initial tectonism of the sediments may have occurred as an outlet glacier in the Great Glen overrode the deltaic sequence, including the underlying glaciomarine sediment that later became deformed into the Clava Shelly Till (Fig. 92 b). The early folding and thrusting that accompanied the detachment may have taken place in response to proglacial deformation, but as the glacier continued to advance north-eastwards along the Great Glen the detached slabs were completely over-ridden. Merritt (1992) suggested that drainage in front of the advancing Great Glen Glacier had been impeded by a the presence of a more powerful outlet glacier flowing out of the Beauly Glacier (Fig. 92 c). This restricted drainage would have assisted in the development of high pore-water pressures within the permeable Clava Sand. Fluid flow through the sedimentary pile would have been retarded by the presence of impermeable layers, such as the Clava Clay and Clava Shelly Till. Moran (1971), Banham (1975) and Moran et al. (1980), and subsequently Burke et al. (2009), have argued that the combination of high pore-water pressures and interlayered permeable and impermeable strata provide optimal conditions for the detachment of rafts.

A difficultly with this explanation is that the initial thrusting and folding of the Clava Shelly Fm appears to have predated hydrofracturing and water-escape, suggesting that pore-water pressures did not begin to increase until a later stage in the deformation history (Phillips and Merritt, 2008). Microtextural evidence clearly shows that the hydrofractures and related water-escape features accommodated several phases of fluid flow, indicative of fluctuating pore-water pressures, which suggests that water was, at least periodically, able to drain from the deforming sedimentary pile. An alternative suggestion is that the rafts were scavenged from the Loch Ness Basin , where ice had thickened sufficiently to have become warm-based, aided by an increase in confining pressure exerted by ice flowing through the Beauly Firth Basin. Similar conclusions have been reached at 'high-level' shelly clay sites around the Firth of Clyde that are now also considered to preserve glacially transported rafts (Merritt et al., 2014; Finlayson et al., 2014). Most of these rafts of sand, silt and clay are also closely associated with 'shelly tills' and were emplaced during an early phase of the last glaciation by ice flowing against adverse slopes, commonly after flowing through topographically-confined basins from which the rafts were detached. It is worthy of note that most rafts appear to have been detached whilst the Scottish Ice Sheet was expanding during the build-up towards the LGM, when deep permafrost is very likely to have formed around its margins.

Once a raft has become detached, many of the published models suggest that it is transported by incorporation into the base of a glacier (Clayton & Moran 1974; Bluemle & Clayton 1984). This model requires a compressive flow regime within the glacier, leading to an upward component of motion allowing the rafts to be sheared into the base of the ice as a result of the decreased effective stress caused by elevated pore-water pressure (Bluemle & Clayton 1984). At Clava, however, there is no direct evidence for the rafts having been frozen and included into the base of the advancing ice. Consequently, Phillips and Merritt (2008) offered an alternative glacitectonic transport model in which the thrusts and related high strain zones that form the margins to the envelope of rafts continue to accommodate displacement during transport beneath the glacier. Shear imposed by the overriding glacier would have been preferentially partitioned into these water lubricated décollement surfaces, effectively 'switching off' deformation in the raft and aiding in the preservation of primary stratigraphical relationships within the thicker parts of the ice rafted sequence. The fluctuation in fluid flow would have resulted in a 'stick-slip' pattern of movement, with displacement decreasing during periods of relatively low pore-water pressure. A fall in pressure probably accompanied renewed deformation of the sediments along the margins of the raft.

Emplacement or deposition of the rafts at Clava occurred on the up-glacier side of a rise in bedrock topography on the southern side of Strathnairn. This feature corresponds with a major change in bedrock geology from the Devonian Inverness Sandstone Group, in the north-west to much less permeable Neoproterozoic metamorphic rocks intruded by granites to the south-east (Fletcher et al. 1996; British Geological Survey 1997). The rafts were accreted against the bedrock 'high' as a result of Great Glen ice being forced south-eastward by ice issuing from the Beauly Firth (Fig. 92 c) (cf. Horne et al. 1894; Merritt 1992; Phillips & Merritt, 2008). This emplacement direction is consistent with the sense of displacement (to the south-east) recorded by a number of the thrusts that deform the Clava sequence. Accretion of the duplex-like, thrust-bound envelope of rafts (Fig. 93) would have aided the ice in overcoming the bedrock high. The accretion process is, therefore, predominantly glacitectonic and probably accompanied the deposition and penecontemporaneous shearing of the basal part of the overlying Finglack Till.

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Figures

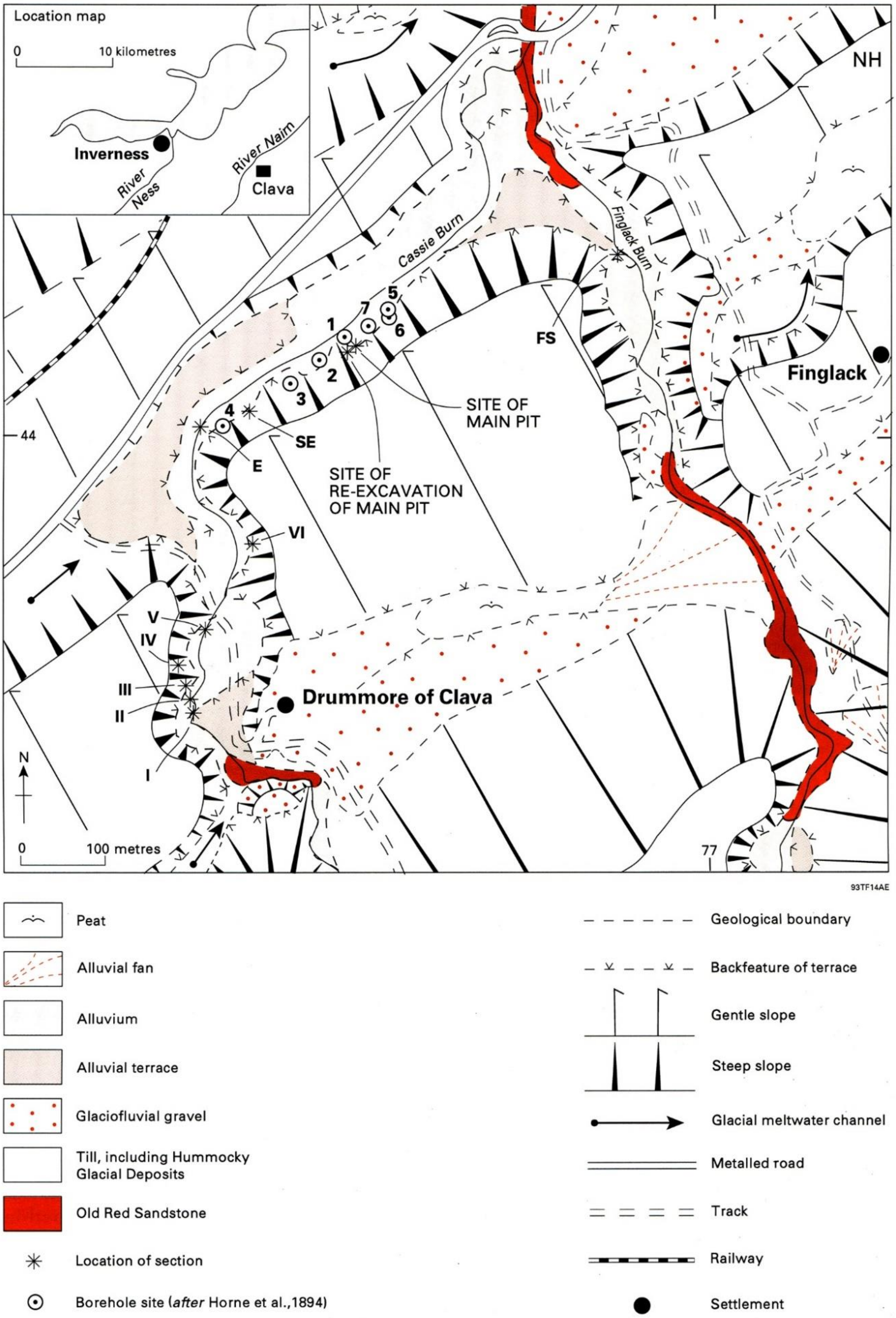

Figure 83. Simplified geological map of the Clava area showing the locations of the main sections studied (from Fletcher et al., 1996). FS Finglack section; I-VI Cassie Burn sections; other sites are described by (Merritt, 1992)

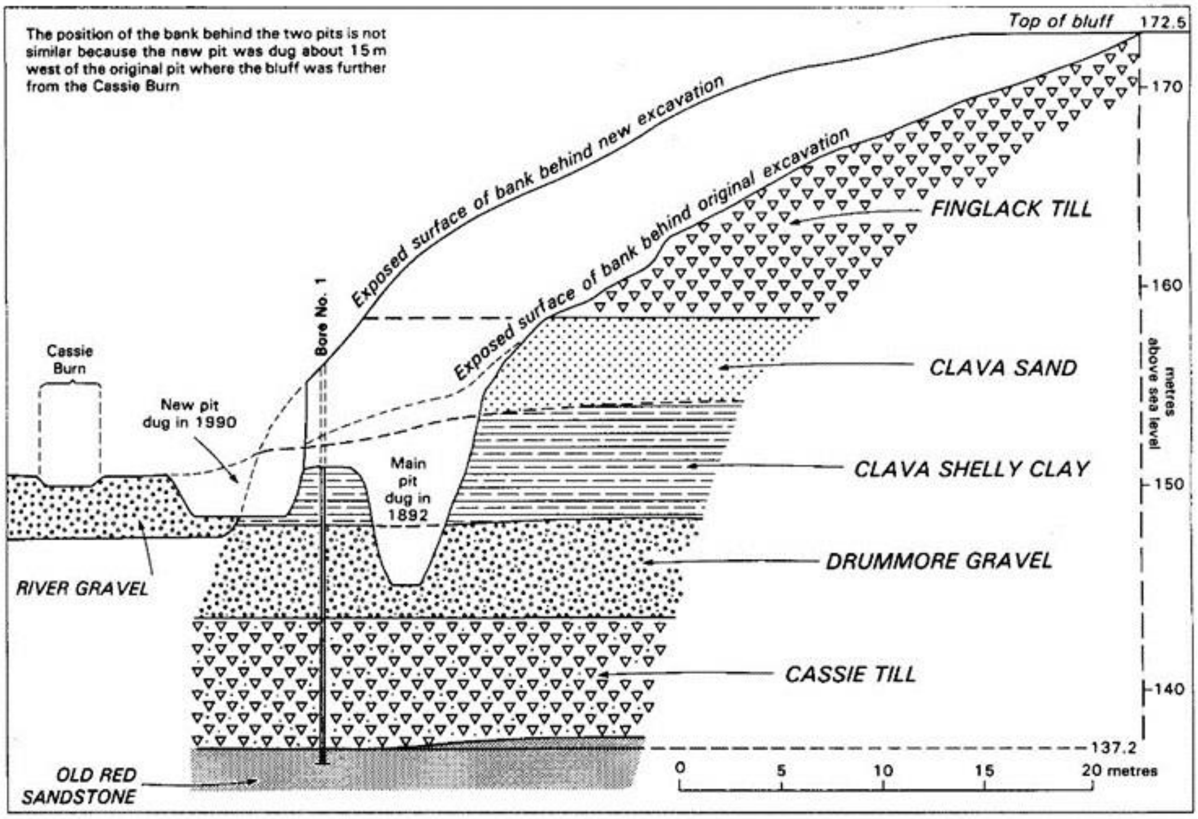

Figure 84. 'Main Pit' section at Clava showing the lithostratigraphical succession and the relative position of the re-excavation in 1990 (from Merritt, 1992).

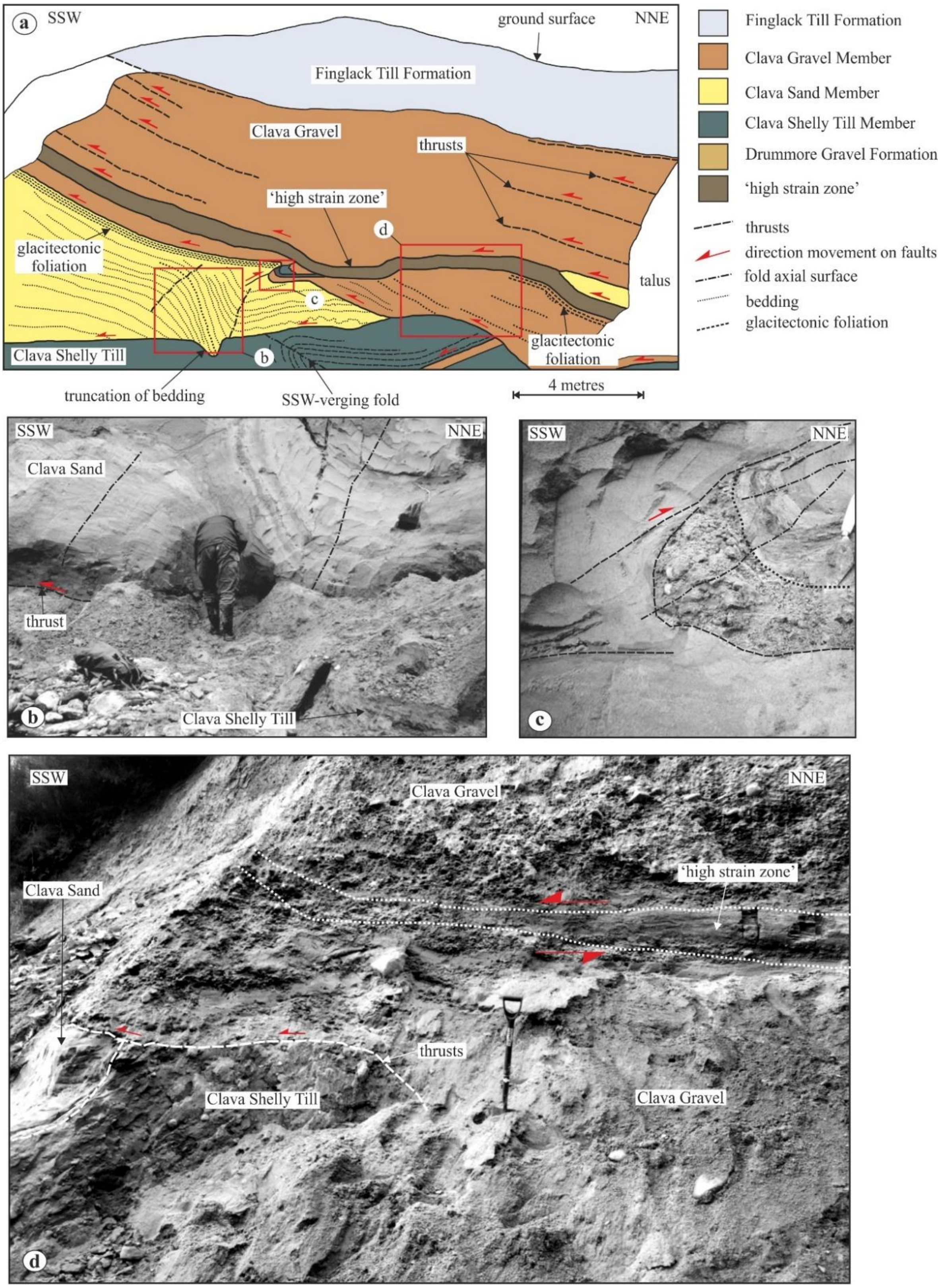

Figure 85. Cassie Burn section III (after Phillips and Merritt, 2008). (a) Section showing location of photos b-d. (b) Large-scale monoclinal folds which deform bedding within the thrust-bound wedge of Clava Sand. (c) Tight, gently inclined fold developed at the tip of a thrust. (d) Prominent SE-directed thrust deforming the basal part of the Clava Gravel. This is marked by a 20–30-cm-thick 'high strain zone' composed of finely layered sand, silt and clay.

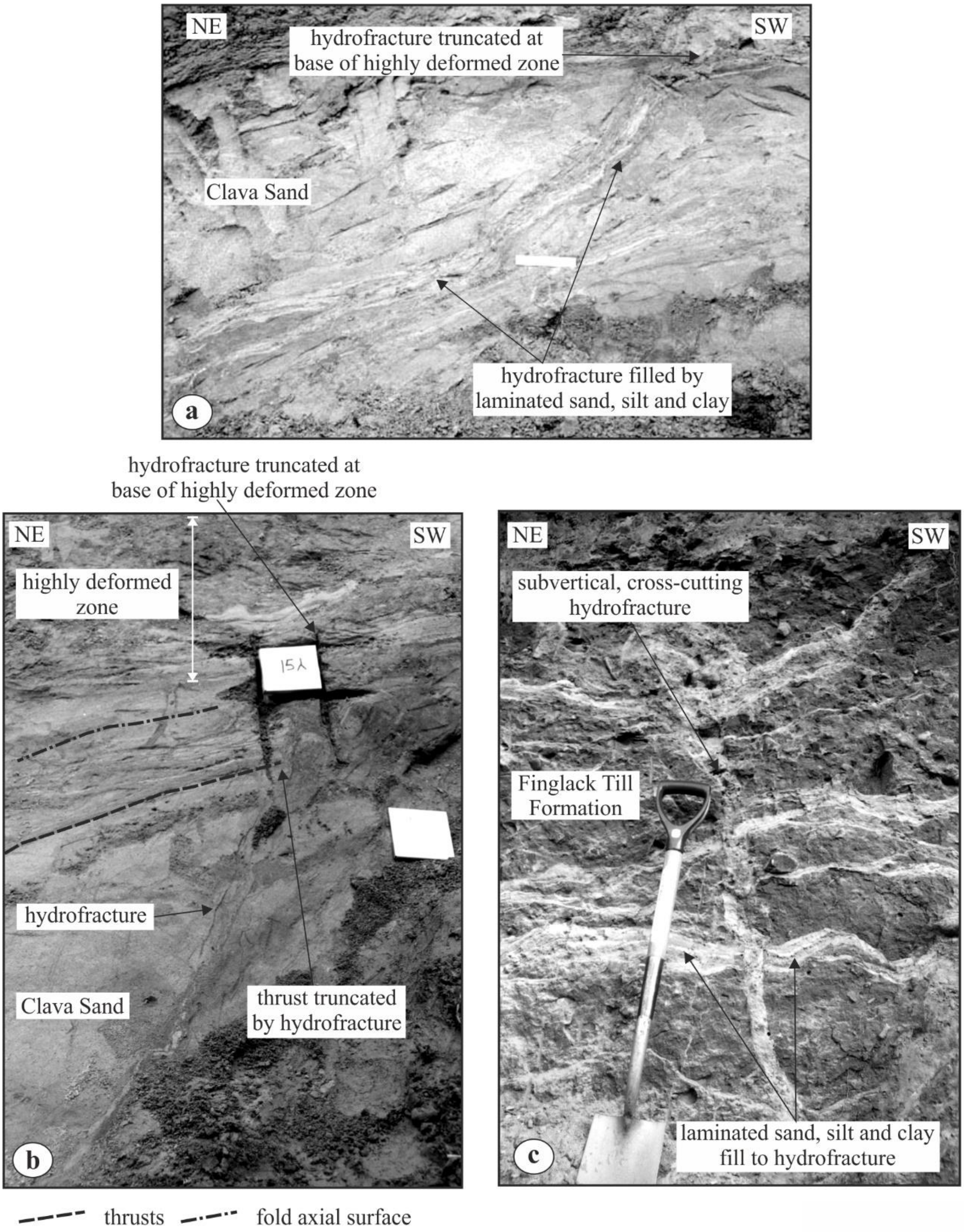

Figure 86. Laterally extensive hydrofracture systems at Clava (after Phillips and Merritt, 2008). (a & b) Hydrofractures within the Clava Sand exposed at the top of the Main Pit section. These features are truncated by a 'high strain zone' immediately below the base of the overlying Finglack Till. The location of blocks collected for sectioning are shown. (c) Network of cross-cutting, gently-dipping and sub-vertical hydrofractures locally developed within the Finglack Till at Cassie Burn section VI. The hydrofractures are filled by pale-coloured, finely laminated sand, silt and clay.

Figure 87. Folded Clava Shelly Till at the base of Cassie Burn section III.

Figure 88. Deformed and folded Clava Sand at Cassie Burn section III.

Figure 89. Cassie Burn section VI in 1988 showing possible subglacial canal-fill, above and to the left of the spade, and a gently dipping, sand-filled hydrofracture passing beneath the boulder.

Figure 90. Close-up of gently-dipping sand-filled hydrofractures at Cassie Burn section VI, as shown in Fig. 86 c.

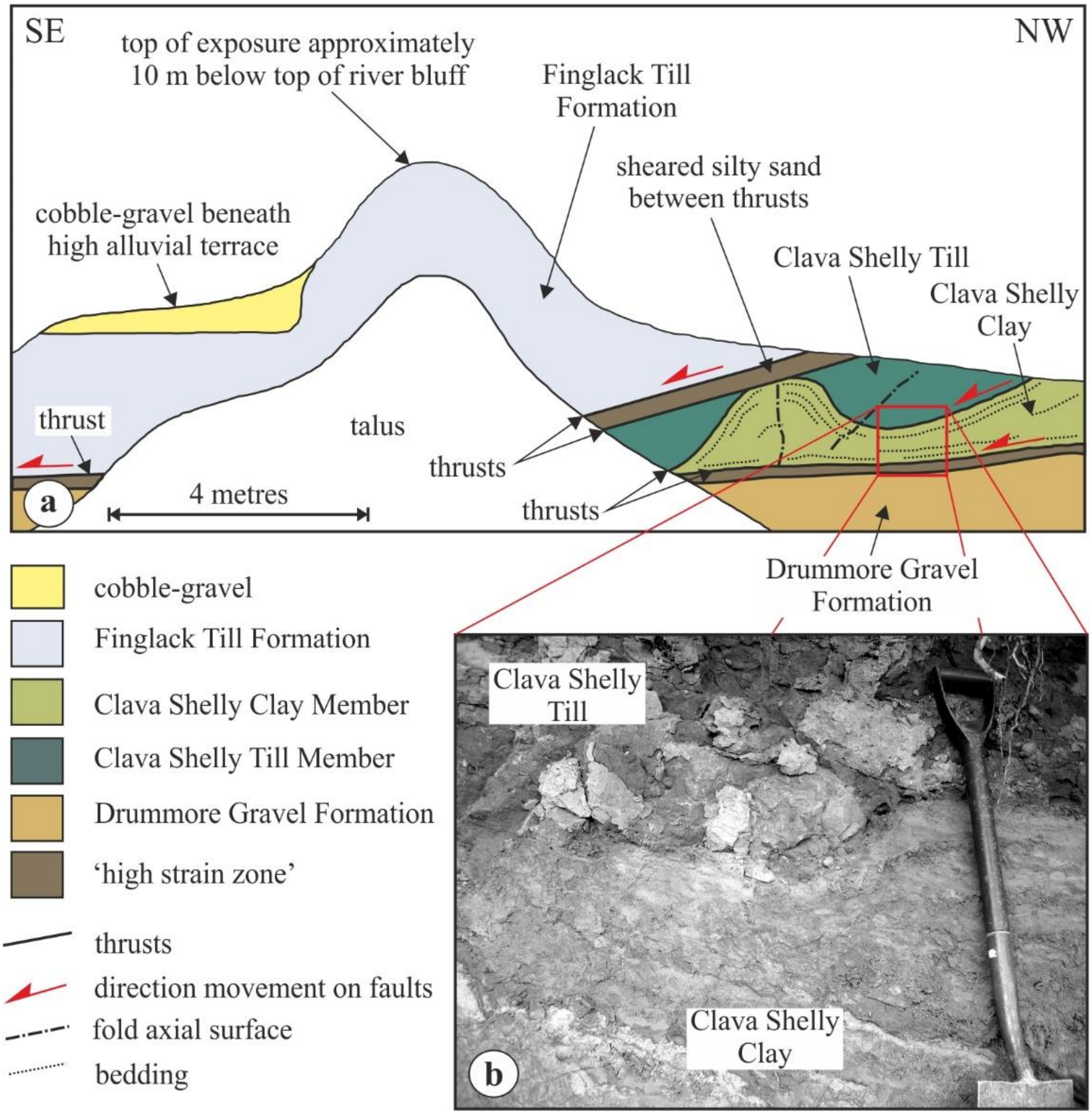

Figure 91. The Finglack Section (after Phillips & Merritt, 2008). (a) Schematic section showing the main deformation structures and location of photo; (b) Highly deformed Clava Shelly Clay exposed near the bottom of the section. Bedding within the clay has locally been transposed by a glacitectonic foliation. Both bedding and foliation are offset by a set of small-scale normal and reverse faults.

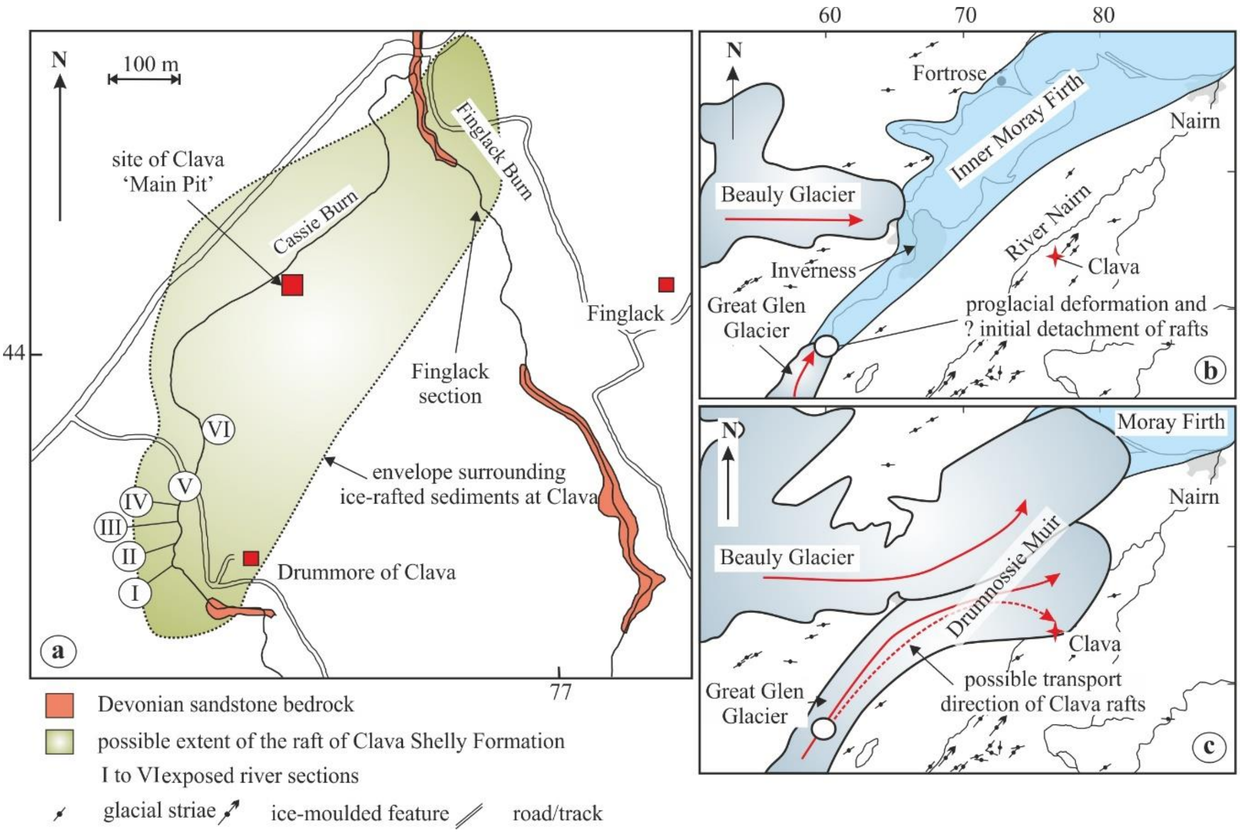

Figure 92. The possible extent and origin of the rafts at Clava (after Merritt, 1992; Phillips & Merritt, 2008). (a) Map showing the possible extent of the envelope of rafts exposed at Clava. (b & c) Reconstructions showing the transport of the rafts from their potential source in the Great Glen to their emplacement at Clava.

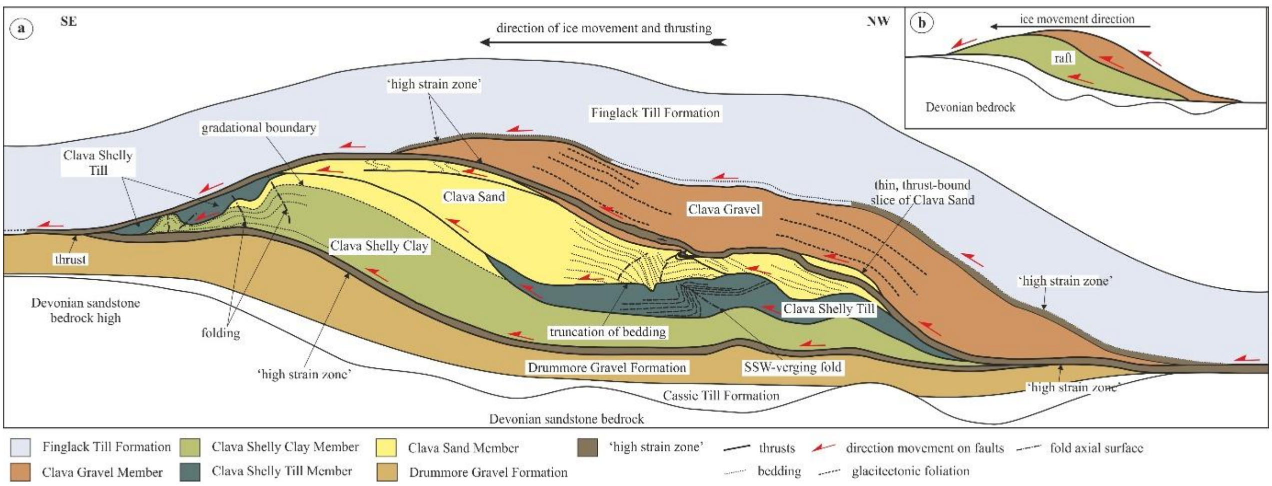

Figure 93. Schematic cross-section (not to scale) through the deformed, ice-rafted sequence exposed at Clava (after Phillips & Merritt, 2008). The relative positions of the main sections are shown; (inset) simplified model of the ice-rafted sequence at Clava showing the main bounding thrusts.