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Valley fills and coastal cliffs buried beneath an alluvial plain: evidence from variation of permeabilities in gravel aquifers, Canterbury Plains, New Zealand.

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Abstract

The large-scale internal structure of the upper ca.10-200 m of the predominantly gravelly Canterbury Plains was investigated using the distribution of permeabilities. Groundwater transmissivity values, estimated from specific capacities, were divided by screen length to approximate hydraulic conductivities for ca. 3,500 wells and produce transmissivity and hydraulic conductivity maps. Semivariograms were used to examine spatial patterns over ca. 6,000 km^2 of the Canterbury Plains. Although the wells produce from different depths and aquifers, regions of distinctly low and high transmissivity and hydraulic conductivity are clearly apparent.

Five to six relatively high permeability corridors, 5-10km wide, trend subperpendicular to the present-day coastline and dissect areas of low permeabilities. These corridors are inferred to be infilled or buried valleys that may act as preferred groundwater flow paths. The valleys developed during the last interglacial stage (130,000-75,000 yrs B.P.) and filled during the last glacial stage (ending ca. 14,000 yrs B.P.). Whilst the valleys were incising, sea cliffs probably developed along a retrograding shoreline, truncating the valleys. Subsequent burial of the cliff results in an abrupt change in sediment structure and texture and aquifer properties across the cliff boundary. High permeability Christchurch artesian aquifers (e.g. Riccarton or Linwood Gravels) partially cover and abut the inferred former cliffs. These aquifers represent reworked outwash deposits analogous to the present-day Waimakariri River fan and floodplain.

Introduction

The Quaternary sediments that form the Canterbury Plains contain sandy gravel aquifers with a large groundwater resource that is important to the region's economy (Fig.1). An important aspect of the plains groundwater management is knowledge of the internal, or subsurface, sedimentary structure that influences the distribution of hydraulic properties.

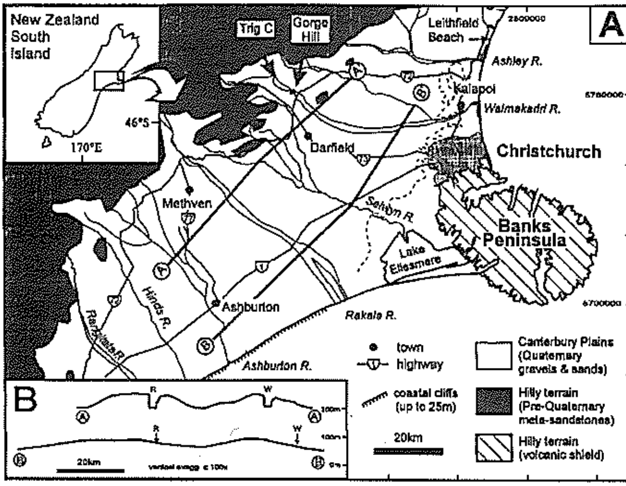

Figure 1 - A) Map of Canterbury Plains. The dashed line marks the western boundary of the upper confining layer of the Christchurch artesian aquifer system. B) Surface profiles through the Rakaia (R) and Waimakariri (W) fluvial fan illustrate the convexity of the fluvial fans and incison in their upper reaches. Minor rivers (e.g. Selwyn River) flow along interfan axes. Note the large vertical exaggeration.

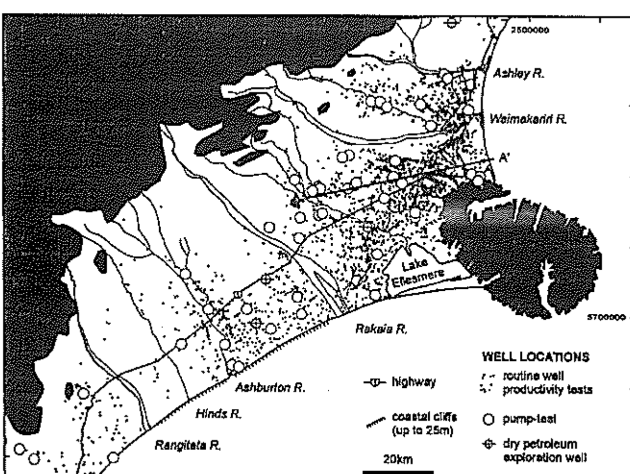

Figure 2 - Distribution of wells (3,800) having basic hydraulic data from which transmissivities were estimated in this study. Locations of 54 wells with pump-test data are also shown.

The subsurface structure of most of the plains is poorly understood, even though the plains have been penetrated by ca. 18,000 water bores and four petroleum wells, and have been the subject of a number of geophysical surveys. Water bores penetrating the Canterbury Plains (Fig.2) show the aquifers are characterised by rapid lateral and vertical changes in yields, transmissivities, and hydraulic conductivities. To make valid predictions using groundwater modelling, these rapid changes need to be averaged in a geologically meaningful way.

Wilsón (1973) demonstrated that well yields and geology could be correlated. His paper implied that influent seepage from rivers was a significant source for groundwater recharge, and followed hypothetical preferred flow paths or "ancient buried river channels" (p.114). Preferred flow paths form when units of high hydraulic conductivity are connected (Silliman and Wright 1988). The role of rivers as an important recharge for the Christchurch artesian aquifers has subsequently been confirmed by isotopic and geophysical work (Taylor *et al.*, 1989; Broadbent and Callander, 1991).

In groundwater modelling, the inherent heterogeneity caused by preferred flow paths is generally overcome by averaging hydraulic parameters. The parameters are averaged over an area which is large compared with the lateral dimension of individual flowpaths. It is therefore important to identify the scales of any preferred flowpaths so that basic assumptions related to continuity and differentiable functions are not violated. Identifying patterns in heterogeneity, or the large-scale structure of sedimentary deposits, and the development of a geological model to explain the observed patterns is therefore important for groundwater modelling, exploration, and management.

This paper addresses the large-scale internal structure of the upper ca. 10-200 m of the Canterbury plains. A geological model is proposed, based on the spatial distribution of hydraulic conductivities, estimated from basic hydraulic data. Possible preferred flow paths are identified and their relevance to modelling, exploration, and management outlined.

The geology of the Canterbury Plains: previous work

Subsurface depositional units are clearly identifable below the coastal area northeast of the Rakaia River mouth through to Christchurch and north to Leithfield Beach. Here gravels interfinger with finer grained estuarine to marine deposits (Suggate, 1958; Wilson, 1976; Brown and Wilson, 1988; Brown and Weeber, 1992, 1994), providing conditions for artesian water wells (Talbot *et al.*, 1986; Fig.1A). This coastal stratigraphy is difficult to correlate with the rest of the Canterbury Plains due to a lack of datable material and continuous marker beds (Brown and Wilson, 1988).

The stratigraphy of the remaining plains is largely defined from exposures in incised river valleys and remnant river terraces. Fluvioglacial outwash surfaces are traced to glacial moraines to help define their chronostratigraphy (Gage and Suggate, 1958; Suggate, 1965, 1990). Soons (1968) describes a recurring pattern in which the Canterbury Plains are repeatedly incised by major rivers during interglacials and the incised valleys are subsequently filled with outwash sediments during each succeeding ice age. The location of these major river valleys may change with each interglacial. In the case of the ancestral Waimakariri River, for example, Wilson (1988) suggests that during the last interglacial period fluvial incision occurred south of Mesozoic basement (Torlesse) rock promontories of Trig C and Gorge Hill (Fig.1A). According to Wilson, this entrenched valley was subsequently infilled early in the last glacial by aggrading fluvioglacial deposits which eventually overtopped the valley in the later stages of aggradation.

With the exception of the coastal zone and river valley outcrops described above, the internal structure of the greater Canterbury Plains is largely unknown despite the many boreholes. Gravel lithologies in the boreholes are relatively uniform and cannot easily be subdivided into depositional units (Brown and Wilson 1988; Wilson 1988).

To date, only the larger scale geophysical surveys provide information about the internal structure of the greater Canterbury Plains (Atkins and Hicks, 1979; Risk, 1974, 1982; Broadbent, 1978). For example, a tentative interpretation (pers. comm. Broadbent) of old petroleum industry seismic reflection surveys indicate localised deposition of up to 600 m of gravel below Lake Ellesmere and Kaipoi (Talbot *et al.*, 1986). This interpretation is consistent with gravel thicknesses observed in wells drilled for petroleum exploration (Fig.2).

Resistivity surveys show significant lateral and vertical variation in the subsurface (Risk 1974, 1982; Broadbent and Callander, 1991). Risk identifies two distinct resistivity layers that are correlated from the Rakaia River to the Waimakariri River, the upper layer being ca. 20 m thick. Broadbent and Callander identified and mapped the flow paths for grounwater recharging the Christchurch artesian aquifers with fresh Waimakariri River water. However, Broadbent and Callander attribute most of the resistivity contrast to variation in water resistivities rather than geological structure (e.g. lithological or textural variations in the sandy gravels).

Up to two or three distinct layers are also interpreted from a refraction survey parallel to the Ashburton River (Atkins and Hicks, 1979). Broadbent (1978) suggested that the distribution of seismic refraction interval velocities between the Rakaia and Waimakariri rivers could be explained by either different gravel formations or by varying distribution of water within the gravels. He suggested that an incised valley may trend from Gorge Hill to Lake Ellesmere (Fig.1A), consistent with Wilson's (1988) interpretation based on sedimentary geometry.

Other smaller-scale geophysical work on the Quaternary gravels include reflection seismic and electromagnetic surveys (e.g. de Vel, 1984; Woodward, 1987, 1989; Bal, 1994). These surveys did not elucidate sedimentary structure within the gravels.

Development of the Canterbury Plains: processes and effects

The Canterbury Plains formed during glacial periods by aggradation and coalescence of fluvioglacial outwash fans comprising sandy gravels. Ice-age deposition was probably characterised by unconfined braided rivers with a high sediment load, with rapid shifting and pendulum sweeping of channels across the fan surface (Suggate, 1965; Soons, 1968; Wilson, 1985). Outwash sediment permeabilities generally improve with increasing distance from the foothills (Wilson, 1973; Scott, 1980). However, on average outwash sediments have lower permeabilities, due to their poor sorting and higher silt and clay contents, than interglacial and postglacial fluvial sandy-gravel deposits (Wilson, 1973; Scott, 1980). The glacial outwash fans grade into a low-gradient coastal plain about 10 km east of the present day shoreline (Herzer, 1981; Barnes, 1994). The sedimentology of the coastal plain appears to be comparable to that of the present-day Waimakariri coastal plain, which is characterised by sand, silts, swamp deposits, and estuarine muds.

Today, and probably in previous interglacial stages, the large fluvioglacial outwash fans are inactive and incised by their formative rivers: the Waimakariri, Rakaia, Ashburton and Rangitata (Suggate, 1963; Wilson 1985; Leckie, 1994). Unlike the glacial braided rivers, the present braided rivers are confined in a trench with narrow floodplains of ca. 2 kilometres width (Fig.1B). Maximum incision occurs at the foothills, where the major rivers are up to 100 m below the surrounding plains. As Wilson (1985) pointed out, the depth of incision decreases seawards towards a kink point, or zone of "minimal erosion" (Leckie, 1994), about 20 km from the modern coast. State Highway 1 follows this zone (Fig.1A). Soons and Gullentops (1973) and Wilson (1985) show incision, down the the knick point, is primarily due to a change in fluvial regime after the ice receded 14,000 years ago. Postglacial reduction in sediment supply resulted in rivers with higher erosive capacity and a predominantly degradational regime. Tectonic and isostatic uplift also contribute to incision.

The postglacial hydrological regime has resulted in the development of a flight of degradational terraces and of smaller fans on the lower plains, the apices of which are within the incised larger glacial outwash fans (Suggate, 1958, 1963; Wilson, 1985, 1988). Much of the sediment that contributes to these fans comes from reworking of the glacial outwash fan by degradation (incision) and bank erosion (Griffiths, 1979; Wilson, 1985; Basher *et al.*, 1988). Fluvial reworking of fluivoglacial outwash improves sorting and consequently permeability (Wilson, 1985; Scott, 1980).

Since the last glacial, the lower Waimakariri River has aggraded to form at least two highly permeable fans from reworked fluvioglacial outwash (Suggate, 1963; Wilson, 1988). The first fan probably developed during the time of rapid sea-level rise immediately after the ice age (Suggate, 1963). The Waimakariri River then flowed to the sea through the Avon-Heathcote estuary, reducing sediment supply to the north (Basher *et al.*, 1988). The nature of the coastline at this time is uncertain, but it appears to have been characterised by fine-grained sediments (sands, peats, and estuarine muds) south of the present day Waimakariri River mouth. The fine-grained sediments suggest a low-energy coastline (Brown and Wilson, 1988; Brown and Webber, 1994). In comparison, coarse-grained sediments north of the present-day river mouth suggest a higher-energy coastline and the presence of a "sea" cliff north of the Ashley River mouth certainly indicates an eroding coastline (Shulmeister and Kirk,1993). This northern sector of the Pegasus Bay coast is characterised by rapid local uplift (ca. 2.0 m/ka; Cowan, 1992) whereas the southern sector is subsiding (ca. 0.25 m/ka; Brown and Weeber, 1994).

The present-day geography shows the most recent Waimakariri River fan passes into a low-angle coastal plain that developed over the past 6,500 years. During this time, the plain prograded about 11 km into Pegasus Bay in a series of pulses, at an average long term rate of 1.5-2.0 m per year (Blake, 1964; Wilson, 1976; Basher *et al.*, 1988). Progradation is favoured by the change from rapid sea-level rise 6,500 years ago to relative stability (Brown and Weeber, 1992). The protection Banks Peninsula provides the northern coastline from erosive southerly swells (Kirk, 1969; Wilson, 1985; Shulmeister and Kirk, 1993; Leckie, 1994) is however a very important control.

Fine-grained coastal sediments form the uppermost confining layers for the Christchurch artesian aquifer system (Fig.1A). Similar deposits, representing earlier interglacials, form deeper confining layers (Brown and Wilson, 1988; Fig.3).

In contrast to Pegasus Bay, the coastline south of Banks Peninsula is largely erosional and wave dominated, as it lacks protection from erosive southerly swells. As a consequence, the outwash fan deposits between the Rakaia and Rangitata rivers are dramatically truncated by coastal cliffs up to 25m high (Fig. 2). The cliffs are presently eroding at a rate of 1 m to 2 m a year, forming a retrogradational coastline (Kirk, 1969; Wilson, 1985). Leckie (1994) suggests that up to 40 vertical metres of outwash gravel is eroded by the southerly swells, 25 m of cliff plus 15 m down to the base of the shoreface. Coastal erosion and cliff development are also caused by undercutting by rivers flowing parallel to the coastline, due to entrapment behind a landward-advancing gravel barrier beach system (Shulmeister and Kirk 1993). The beaches are predominantly reworked outwash gravels supplied by the rivers and coastal cliffs. Most (95%) of the deposited gravel is rapidly abraded to mud in the wave swash-backwash zone (Gibb and Adams, 1982).

Development of cliffs in glacial outwash gravels by rising or high sea-level stands is common in New Zealand (q.v. Beu *et al.*, 1987; Suggate, 1990). Cliff development and subsequent burial results in an abrupt change in large-scale sedimentary structure and, most likely, in aquifer properties across the cliff boundary.

The northern coast differs from the southern coast only in the context of today's "snap-shot" geography defined by present-day sea level. Coastal plains also developed south of Banks Peninsula during falling sea level. The apparent difference now is simply because of the efficacy of erosive southerly swells during static or rising sea level.

Analysis of hydraulic data

It is well established that hydraulic properties largely reflect depositional processes. The following section analyses the distribution of hydraulic parameters in order to elucidate subsurface large-scale sedimentary structure.

Transmissivity, hydraulic conductivity, and specific capacity

Transmissivity (T) values allow us to roughly estimate probable groundwater production and is the rate at which groundwater flows through the entire thickness (D) of the aquifer under a hydraulic gradient of one. Hydraulic conductivity (K) is an indicator of permeability and is the volume of fluid passing through unit cross section in unit time under the action of unit pressure gradient, usually reported in terms of cm s^{-1} or metres per day. The relationship between these variables is (Driscoll, 1986):

$$T = K D \qquad [1]$$

Transmissivity is usually determined from data collected during controlled pump-tests, usually carried out over three, four, or more days.

Generally, pump-tests are costly and are therefore not performed routinely. Consequently too few pump-test data are available to adequately describe the spatial distribution of transmissivity in the plains (Fig. 2).

There are, however, many theoretical and empirical methods that may be used to estimate transmissivity from existing hydraulic data derived from routine well-productivity tests carried out by drillers (q.v. Logan, 1964; Walton, 1970; Kruseman and de Ridder, 1970; Driscoll, 1986). Most of these methods are subject to very large errors but the results give a general indication of hydraulic properties.

Basic hydraulic data from drillers' well-productivity tests, carried out for an unspecified length of time (usually hours), have been recorded for about 5,400 of the ca. 18,000 available water wells. These basic data commonly include the yield or discharge rates (the rates at which water can be pumped from the wells) and, less often, drawdown (the drop in water level in the well during pumping). Specific capacity (discharge per unit drawdown) is calculated from discharge and drawdown and may be used as an indicator of transmissivity — high capacities usually indicate high transmissivities and low capacities indicate low transmissivities (Walton, 1970).

Specific capacity is not a precise or accurate estimator for transmissivity; usually transmissivity is underestimated because wells rarely fully penrtrate the aquifer, or because of well losses and hydrologic boundaries (Walton, 1970). For the Canterbury Plains, well losses are probably the main reason for underestimation of true transmissivity (pers. comm. D. Scott, 1995). Specific capacity varies with duration of pumping and discharge. As pumping time or discharge increases, specific capacity decreases (Walton, 1970; Driscoll, 1986). Specific capacity has been calculated for the ca. 3,800 wells having both yield and drawdown data.

Vertical variation in hydraulic parameters, as defined by the producing interval, are influenced by the depth of the water table, discharge rates, and drilling costs. Wells generally penetrate only as deep as necessary to provide the required quality and discharge rate. The hydraulic properties of deep aquifers are therefore rarely measured, except in the greater Christchurch area where deep wells supply municipal drinking water (Fig.3). Eighty percent of the wells used in this study penetrate less than 60 m of plains, whereas the gravels are generally ca. 300-600 m thick (Talbot *et al.*, 1986).

The hydraulic properties of the unsaturated zone are also largely unmeasured. This zone is somewhere between the ground surface and the surface defined by the tops of the uppermost screens (assuming that screen tops were mostly set below the water table) (Fig.3). Zones with no screens, bounded above and below with screened intervals, are interpreted as

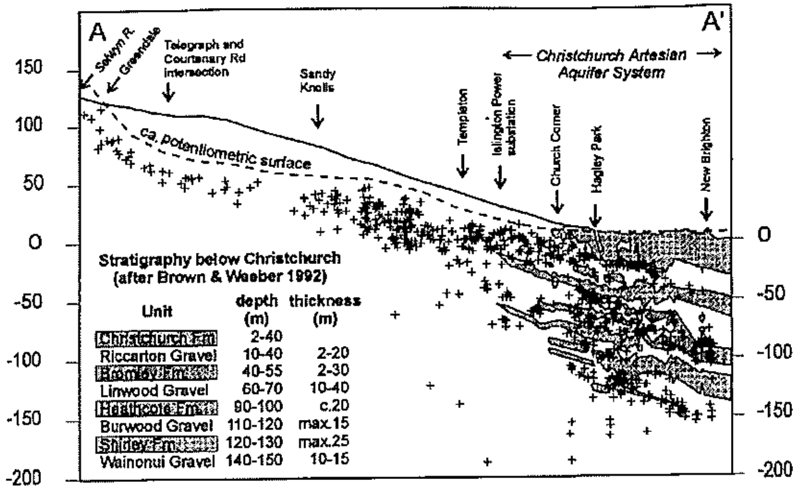

Figure 3 - Cross-section along line A-A' in Figure 2 showing distribution of the total depth of all wells 2.5 km either side of the section. Fine-grained confining layers of the Christchurch artesian aquifers are shaded (after Talbot, 1986). The section includes many wells with no reported hydraulic data. Commonly only the bottom two or three metres of a well is screened, the well depths therefore approximate the screened intervals.

aquitards or aquicludes (Fig.3). Such an interpretation presumes production zones are preferentially screened.

Estimating transmissivity and permeabilities

An empirical relationship between transmissivity, derived from controlled pump-test data, and specific capacity, calculated from drillers' discharge and drawdown data, has been established by linear regression (Table 1, Fig. 4).

$$\log 10\, [T(m^2/d)] = 0.9619 * \log 10\, [SPC(m^2/d)] + 0.611 \qquad [2]$$

Both Walton (1970) and Driscoll (1986) show that the theoretical relationship between specific capacity and transmissivity is linear. The transmissivity and specific capacity values follow log-normal distributions covering several orders of magnitude. The values were therefore log transformed for regression analysis. The residuals of high magnitude values may differ from the measured value by as much as an order of magnitude, therefore the estimates are not equivalent to the more quantitatively-derived pump-test data. The percentage error in the estimate is tabulated in Table 1.

Table 1 - Pump test transmissivities used to calibrate specific capacities calculated from data supplied by drillers and stored on the Canterbury Regional Council database. Percent difference is the residual using the equation in [2]. See Figure 2 for pump test locations.

	well	easting	northing	depth (m)	screen length (m)	yield (l/d)	draw-down (m)	specific capapcity (m^2/d)	trans-missivity (m^2/d)	ref	% diff
1	J38/0020	2354200	5662300	9.6	-	8.3	4.5	159	500	4	1.1
2	J38/0029	2358400	5659500	9.5	7.7	6.9	5.2	115	147	4	19.6
3	K36/0102	2401500	5712200	6.1	3.0	13.6	1.0	1,187	3,456	4	0.8
4	K37/0338	2400300	5692300	12.2	-	4.8	1.9	214	302	4	15.0
5	K37/0432	2408000	5702170	8.3	-	12.2	1.4	764	3,542	4	-4.7
6	K38/0025	2372450	5678070	12.0	3.0	27.0	4.7	495	10,973	4	-20.7
7	K38/0072	2380670	5660100	60.5	8.0	48.0	5.9	699	2,333	4	-0.6
8	L35/0185	2448200	5746000	110.9	11.2	13.6	33.5	35	240	2	-11.9
9	L35/0191	2446300	5745400	115.2	3.0	34.1	6.0	491	4,026	2	-11.2
10	L36/0023	2425400	5725500	109.5	13.5	18.2	2.7	582	2,640	2	-4.4
11	L36/0058	2432700	5739000	82.9	4.9	45.5	11.2	351	1,089	2	0.7
12	L36/0063	2439400	5735700	56.3	6.3	21.2	1.8	1,018	2,669	2	2.3
13	L36/0065	2439400	5737100	61.7	6.0	8.3	0.1	7,171	14,704	2	3.6
14	L36/0066	2437100	5735500	51.2	9.2	6.8	14.9	39	701	2	-24.6
15	L36/0067	2437800	5735900	45.0	6.0	9.1	6.1	129	1,037	4	-12.4
16	L36/0068	2438200	5736100	49.0	6.2	10.9	4.3	219	298	2	15.7
17	L36/0086	2440000	5736600	62.1	5.5	13.6	0.3	3,917	6,860	2	6.0
18	L36/0087	2447800	5736600	58.5	3.4	30.3	3.4	770	4,086	2	-6.2
19	L36/0093	2449800	5737600	61.9	4.9	35.2	0.5	6,083	20,132	2	-1.2
20	L36/0119	2434200	5728300	46.5	4.5	25.8	1.5	1,486	2,848	2	6.0
21	L36/0120	2433900	5722500	85.0	7.3	26.5	4.6	498	775	2	10.9
22	L36/0122	2439600	5729000	24.0	12.7	22.7	6.1	322	462	2	13.4
23	L36/0206	2440600	5729500	14.3	3.0	22.7	5.9	332	4,981	2	-17.9
24	L37/0008	2414640	5686480	29.5	6.2	27.0	9.5	245	69	4	58.1
25	L37/0012	2433400	5705800	43.3	4.6	39.0	34.7	97	1,190	1	-18.0
26	L37/0020	2425780	5696720	67.7	3.1	75.0	5.5	1,180	7,862	4	-8.5
27	L37/0021	2419000	5702300	72.0	4.6	31.5	7.5	362	930	1	3.5
28	L37/0026	2434530	5701000	35.0	3.0	7.6	4.6	144	733	1	-6.3
29	L37/0143	2412680	5692200	40.0	-	33.3	22.6	127	3,110	4	-24.5
30	L37/0374	2416650	5686000	45.4	3.0	66.7	19.9	290	691	4	4.9
31	M35/0470	2482200	5765600	30.2	5.9	13.3	3.0	383	360	4	21.1
32	M35/0847	2482400	5758800	98.0	6.2	105.0	11.2	810	130	4	61.3
33	M35/1104	2460900	5740000	55.4	5.9	24.2	17.8	117	507	2	-3.8
34	M35/1154	2461400	5744500	30.6	3.0	19.7	8.0	213	776	2	-1.4
35	M35/2689	2454580	5761100	12.0	0.3	30.3	1.6	1,636	6,508	4	-2.9
36	M35/2880	2475500	5768900	-	-	57.5	0.5	11,040	9,184	4	13.5
37	M35/3813	2483210	5740800	129.6	6.0	83.4	9.5	759	3,888	3	-5.8
38	M35/4897	2483000	5766000	22.3	3.0	26.2	2.5	905	776	4	19.6
39	M35/5135	2485620	5740180	155.4	5.1	56.9	13.0	378	778	3	6.9
40	M35/5212	2471500	5754400	32.6	3.7	19.0	17.9	92	372	4	-2.8
41	M35/6246	2457190	5761050	14.3	2.1	13.6	6.5	181	665	4	-1.5
42	M35/6312	2476400	5767700	199.6	-	1.0	12.0	7.2	1.3	4	1160
43	M35/6527	2459100	5759600	19.8	8.0	22.0	5.5	346	4,320	4	-16.0
44	M35/6733	2468660	5760330	-	-	24.0	9.2	224	6,048	4	-24.0
45	M36/0028	2458300	5733900	50.6	6.1	16.3	19.8	71	174	2	6.8
46	M36/0040	2450300	5738300	59.4	3.0	34.1	0.8	3,683	14,913	2	-3.2
47	M36/0175	2470700	5738500	21.0	3.2	45.0	3.7	1,065	15,552	3	-16.0
48	M36/0226	2463400	5738200	52.8	9.4	5.7	27.3	18	91	2	-7.1
49	M36/0516	2467000	5728600	19.8	4.3	41.7	2.7	1,334	2,505	2	6.4
50	M36/0699	2450400	5710500	14.0	3.3	53.0	1.5	3,053	31,400	2	-11.9
51	M36/0771	2456300	5718100	58.6	6.0	49.2	3.7	1,149	3,543	2	0.1
52	M36/0979	2485100	5739000	36.6	5.7	22.8	6.2	318	259	3	25.0
53	M36/1249	2467100	5729800	38.7	6.4	36.2	2.4	1,303	2,765	2	4.8
54	M37/0048	2455600	5706500	10.6	3.4	56.8	2.0	2,454	11,334	4	-4.5
			min	6.1	0.3	1.0	0.1	7.2	1.3		
			max	200	14	105	35	11,040	31,400		

source of transmissivity values

1	Scott & Thorpe 1986	3	Talbot *et.al.* 1986
2	Bowden *et. al.* 1983	4	Canterbury Regional Council files

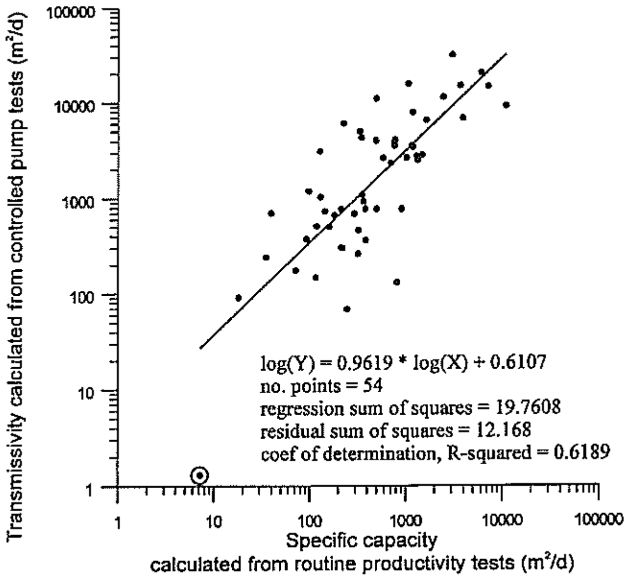

Figure 4 - Regression analysis of the relationship between drillers routine well-productivity tests (expressed as specific capacity) and transmissivity calculated from controlled pump-tests. The transmissivity explains 62% of the variance in specific capacity. The regression is significant at the 99% level as evidenced by high F-value of 86 and t-value of 2.74. The apparent outlier (circled) is believed to represent a valid test and is therefore not excluded from the analysis. The coefficient of determination: r^2 is 0.62 with the outlier and 0.55 without the outlier; there is not a significant difference in regression slope.

Using this relationship [2], T was estimated for 3,807 well locations where specific capacity data was available. Ninty-nine percent of the estimated transmissivities fall within the minimum and maximum range of the observed values (Table 1). Estimated transmissivities were also normalised by dividing them by the screen length wherever that was available (3,481 wells) to estimate hydraulic conductivity (i.e. K in equation [1]). The screen length may not necessarily equal the aquifer thickness, however the normalised value can be used as a relative measure of permeability. Both estimated transmissivities and hydraulic conductivities range over several orders of magnitude and are lognormal, as are the specific capacities on which the values are based.

Despite the fact that the wells produce water from different depths and aquifers, areas of distinctly low and high transmissivity are apparent (Fig.5). The most obvious example is an area of low transmissivity west of Banks Peninsula and adjacent to the narrow northeast-trending zone of high transmissivity between the Waimakariri River mouth and Lake Ellesmere.

Table 2 - Results of trend analysis of 3,806 specific capacity values from Canterbury Plains water bores. All fitted trends are significant at the 99% level.

Trend order	Goodness of fit %	Correlation coefficent	f-test
1	5.47	0.23	110
2	6.03	0.25	48.8
3	8.07	0.28	37.0

Figure 5 - Distribution of low and high transmissivities. To avoid clutter, only the lower and upper 20 percentiles of data are plotted.

Spatial variability in transmissivity and hydraulic conductivity

Data variability has two components: (i) the variability attributed to trends in the data and (ii) variability that is undefinable due to properties intrinsic to the phenomena, to too large a sampling distance or to measurement errors i.e. "noise". Knowledge about the variability structure of spatial data is necessary to interpret data or interpolate values (e.g. when contouring).

Trend analysis shows that there is a general southeasterly trend of increasing transmissivity and hydraulic conductivity, as noted by Wilson (1973). However on a Canterbury-wide scale, this trend only explains about 6% of data variability (Table 2). This suggests the distribution of the hydraulic parameters is much more complicated than previously recognised. The spatial varibility in the data has therefore been quantified by calculating a semivariogram.

Semivariograms compare the mean, or expectation, of the squared difference of many pairs of values as a function of the distance between the pairs. They are often used for geologic or hydraulic data (Journel and Huijbreght, 1978; Rogowski and Simmons, 1988; Eisenberg *et al.*, 1994; Ritzi *et al.*, 1994). Data for points that are close together are usually interrelated and have values of similar magnitude, whereas data for points located further apart will usually be less interrelated. As the distance between data points increases, the mean of the squared differences is expected to increase, resulting in larger semivariogram values. The semivariogram function can be written in the following form (Journel and Huijbreghts, 1978):

$$\text{semivariogram} = 1/2\ E\ [Z(x)\text{-}Z(x+h)]^2$$

where

E - is the expectation or average value over all paired samples at a given distance,
x - is a spatial coordinate location,
Z - is a value of a parameter (e.g. transmissivity), and
h - is the distance (or lag) from x.

Experimental and theoretical semivariograms are calculated from the estimated transmissivity and hydraulic conductivities (Fig.6). "Experimental" refers to the computed points and "theoretical" to the mathematical model fitted to the experimental semivariogram (Journel and Huijbreght, 1978). Three useful parameters describe the variability or correlation structure of the data (Cressie, 1988):

1. The nugget effect quantifies the precision that is obtained with the available data. Transmissivity and hydraulic conductivity, respectively have

nuggets of 48% and 47% of the overall sample variance, showing a high proportion of random small-scale noise that can be attributed to the same sources of error affecting hydraulic parameters from the Canterbury Plains that Scott (1980) identified; namely, inaccurate discharge and drawdown readings, variations in screen types and well construction, and well losses.

2. A range of ca. 16 km shows there is an appreciable amount of predictable continuity in the data despite the high proportion of noise. The range defines the sphere of influence, and relative weights, over which values in the data set are interdependent or correlated. Beyond the range, values are considered to be independent.

3. The sill defines the overall variance in the data set. The sill needs to be defined if the data is to be modelled using kriging techniques (Journel and Huijbregts, 1978).

The data were also tested for anisotropy. Directional semivariograms were computed using an angular tolerance of 22.5° on either side of lines coincident with depositional strike and dip, respectively oriented at azimuths of 45 and 135 degrees. The directional semivariograms have the same nugget effects and similar range as for the omni-directional semivariogram. However variance along depositional dip is only marginally larger than variance along strike. This difference is not considered large enough to require special modelling treatment along the depositional dip direction.

Spatial trends in transmissivity and hydraulic conductivity

The semivariogram analysis indicates there is sufficient integrity in the data to examine spatial patterns in both transmissivity and hydraulic conductivity. Both transmissivity and hydraulic conductivity data are used to estimate values at the nodes of a square grid with spacing of 2 km x 2 km by kriging, using the semivariogram described above. Kriging is a standard geostatistical technique and provides an unbiased, minimum-variance estimate of regional variables, while taking into account the high degree of noise in the data (Journel and Huijbreght, 1978). Both grids were contoured and are illustrated in Figures 7 and 8.

Existing Canterbury Plains transmissivity maps have been derived using the graphical analysis method outlined by Hunt and Wilson (1974), which is also based on specific capacities measured in existing wells. These maps cover only parts of the Plains (e.g. Hunt, 1975; Day, 1976; Bowden *et al.*, 1983; Scott and Thorpe, 1986; Lee, 1988; Anderson, 1994). However, corresponding areas of the Canterbury-wide map dervied by kriging can be compared with these earlier maps; the patterns generated by groundwater

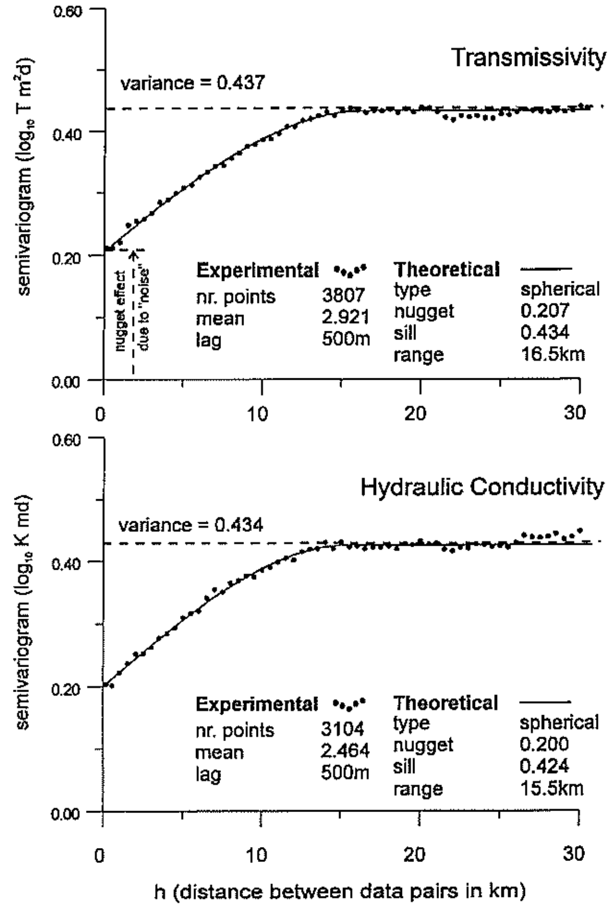

Figure 6 - Experimental and modelled semivariogram calculated for transmissivities and hydraulic conductivities. The semivariogram was calculated for 500 m intervals (lag). Each point represents the sum of 20,000-30,000 pairs.

modelling and geostatistical modelling are very similar. On average, however, transmissivities estimated by the Hunt and Wilson (1974) method are an order of magnitude larger than those estimated by geostatistics.

Figure 7 and 8 show the spatial complexity of aquifer parameters. The shading of areas with hydraulic conductivity values below 2.4 (log10 m/d) in Figure 8 emphasises the dissection of low conductivity areas by comparatively high conductivity corridors (5 to 10 km wide) subperpendicular to the coastline. These corridors do not necessarily coincide with present day river beds. The corridors are interpreted as

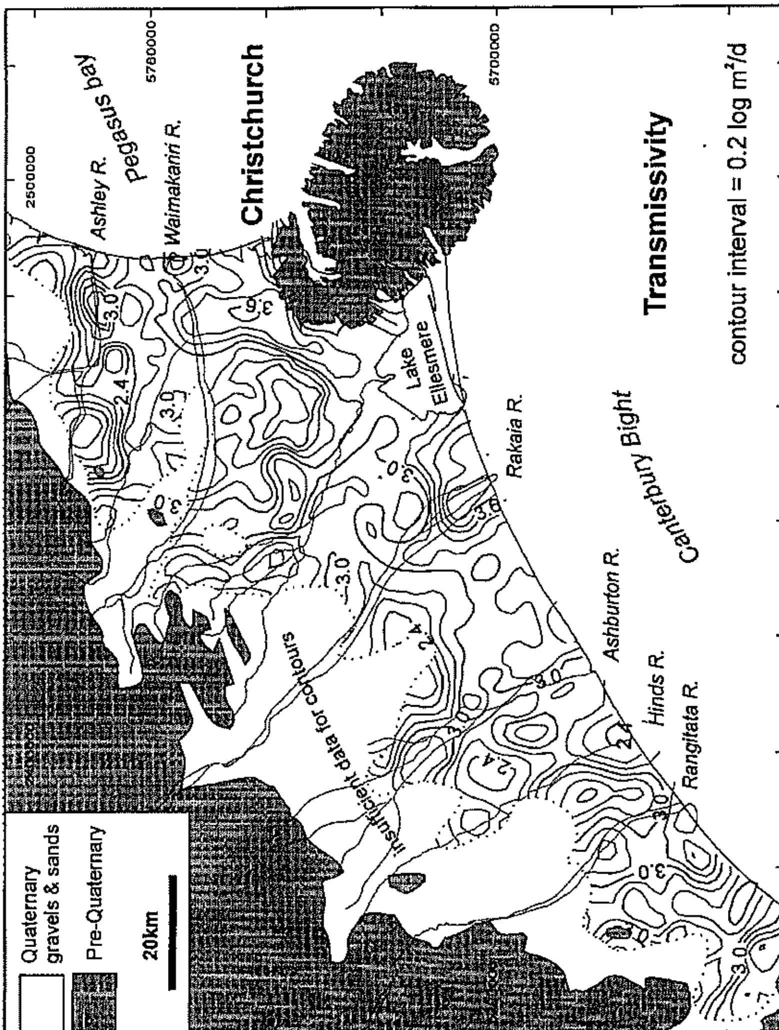

Figure 7 - Contour map of transmissivity. Contours are truncated along the margin of areas with limited or no data (data locations are shown in Fig.2).

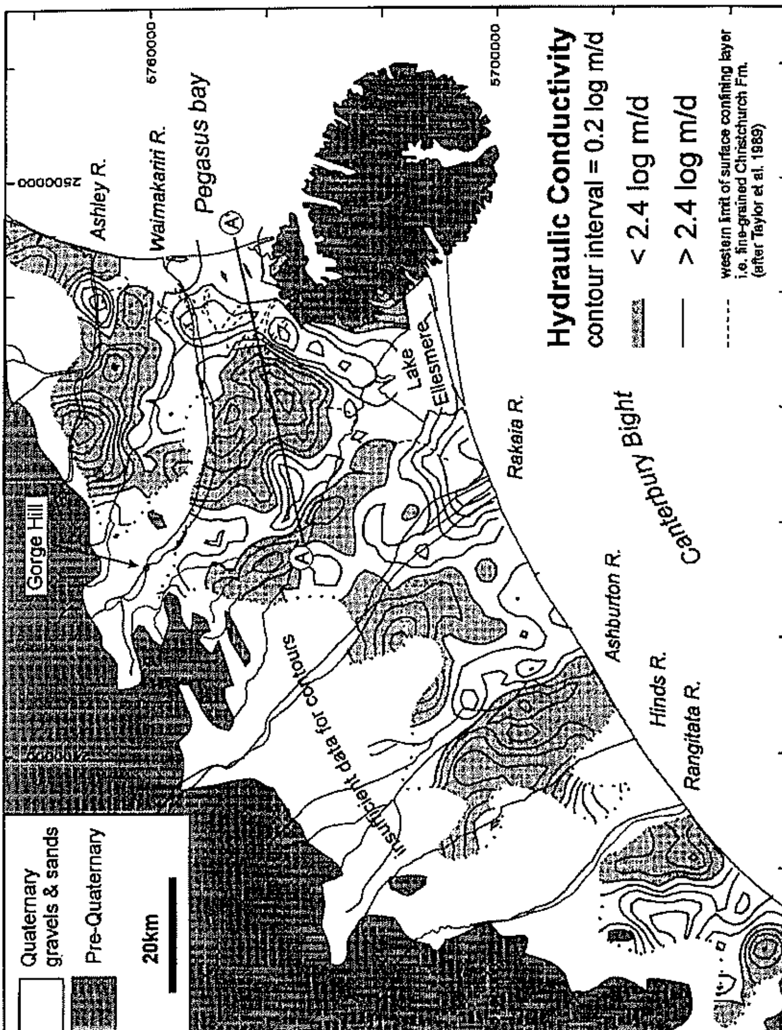

Figure 8 - Contour map of hydraulic conductivity. Contours are truncated along the margin of areas with limited or no data (data locations are shown in Fig.2). Shading emphasises the dissection of areas with low hydraulic conductivity by corridors and a coastal fringe of comparatively high conductivity. Subsurface stratigraphy and hydraulic conductivities along cross-section A-A' are shown in Figures 3 and 9.

conduits that follow infilled valleys that are larger in scale than the "buried river channels" of Wilson (1973). These may be preferred groundwater flow paths.

Another zone of high conductivity, perpendicular to the high conductivity corridors, fringes the present day coastline and the western margin of Banks Peninsula. This zone appears to truncate the low conductivity areas along an arcurate curve connecting the eastern to southeastern margin of the 2.4 (log m/d) contour. The truncation is particularly striking west of Banks Peninsula, but admittedly less well defined between the Rakaia and Ashburton Rivers. The marked truncation may delineate a former, now buried, cliffed coastline.

The maps do not reflect the conductivity distribution at depth. Some areas of high conductivity are surficial deposits of recent gravels that onlap older, less permeable, outwash deposits. An example is the area of high conductivities near the mouth of the Ashley River (Fig. 8). However, highly conductive units are not only restricted to shallow deposits but often occur at similar depths to adjacent units displaying low conductivity (Fig. 9). The vertical distributions of conductivities outline broad zones or units having similar values. Adjacent units may show order-of-magnitude differences in conductivities, suggesting juxtaposition of different depositional units. As suggested by the trend analysis, conductivities do not smoothly increase from west to east (or any other direction for that matter).

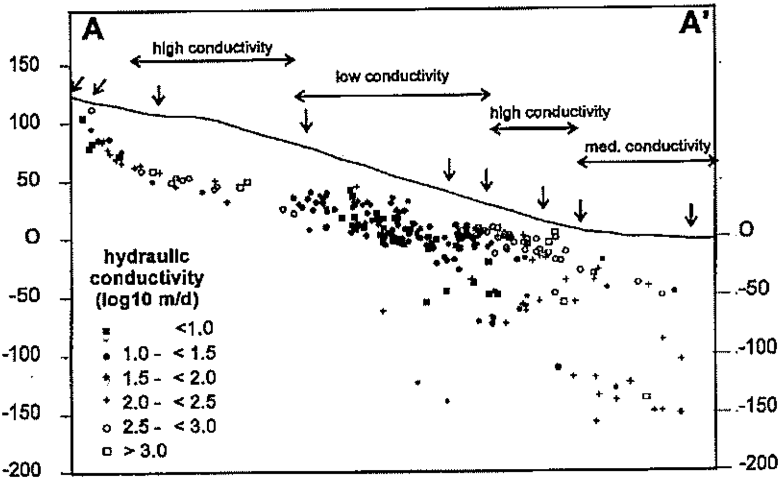

Figure 9 - Cross-section along line A-A' (Fig.8) showing distribution of hydraulic conductivities. Wells 2.5 km either side of the section are projected onto the section. Section can be compared with the subsurface straigraphy in Figure 3.

Towards a refined model

The observed hydraulic patterns provide a three-dimensional framework for the large-scale sedimentary structure of the Canterbury Plains. The structure reflects both past and present processes and can be explained in terms of the above described fluvial and marine processes during interglacial and glacial periods.

On the basis of the patterns, I suggest that the corridor of relatively high conductivity values that trends from Gorge Hill to Lake Ellesmere follows the now buried incised valley referred to by Broadbent (1978) and Wilson (1988). There is not enough information to accurately determine the depth and width of this early incision, but present-day incision shows that at least 100 m is possible. The buried valley's upper-plains western boundary is limited by the terrace outcrops of older (Woodlands and Hororata Formation) glacial outwash sediments (see Wilson, 1988) and underlies Darfield. By analogy, the other four, possibly five, northwest- and southeast-trending high conductivity corridors that cross the length of the plains are inferred to define valleys also incised during the last interglacial (Fig.8, Fig.10A). It is probable that maximum rates of incision occurred during the interglacial high sea-level stand shortly after deglaciation.

During the last interglacial, eustatic sea-level attained and sustained its highest level from ca. 125,000 to 122,560 years ago, i.e. the 2,500 years marked by the oxygen isotope substage 5e (Chappell and Shackleton, 1986; Martinson *et al.*, 1987). During this time a narrow seaway may have connected Pegasus Bay with the Canterbury Bight. Banks Peninsula would then be an island (Fig. 10A). A "Banks Island" is equivocal (c.f. Brown and Weeber, 1994). However recent work on diatoms also suggests Banks Peninsula was an island during the earliest part of the interglacial (Holt, 1995). Such an interpretation is supported by the presence of wave-cut platforms that indicate an interglacial sea-level about 6-8 m above present day sea-level (Lawrie 1993): an interpretation comparable with global data (e.g. Chappell and Shackleton, 1986). The platforms developed at intertidal levels as a consequence of substantial wave erosion along the northern edge of Lake Ellesmere.

During this high sea-level strong tidal currents, or swells, may have developed sea cliffs west of Banks Peninsula similar to those between the Rangitata and Rakaia Rivers today. Little if any deposition would have ocurred in such an erosive environment. The easterly truncation of the low permeability area west of Banks Peninsula (Fig.8) may delimit the now-buried conjectural sea cliff. The sea cliff roughly coincides with the western limit of the Christchurch Formation and the other underlying fine-grained formations (Figs. 3 and 8).

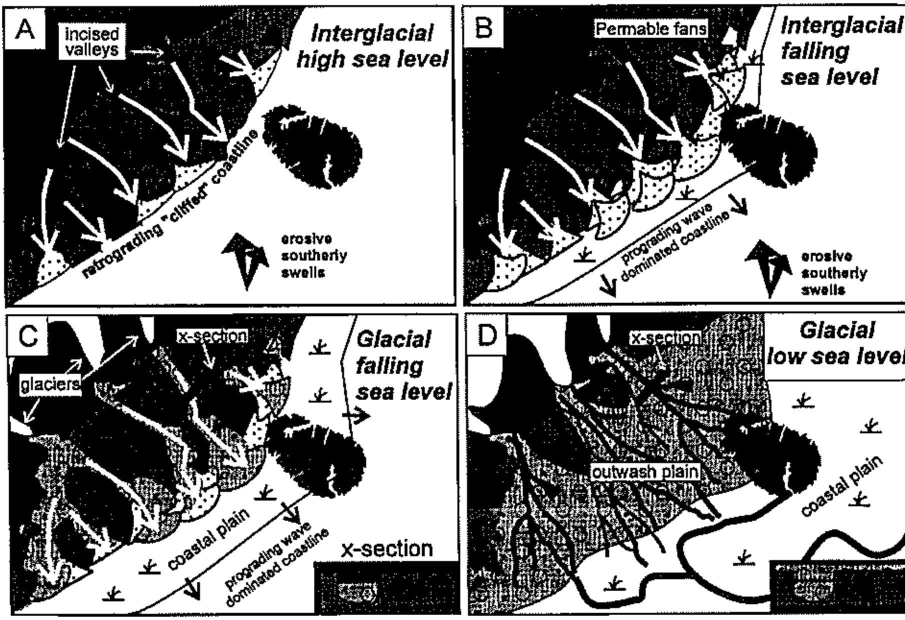

Figure 10 - Paleogeographic reconstructions illustrate the geological model described in the text. (A) Interglacial highest sea-level ca. 125,000 years B.P.. White arrows depict major rivers and incised valleys inferred from the contour map of hydraulic conductivities (Fig.8). (B) Falling sea-level results in deposition and preservation of reworked-outwash permeable fans that abut and cover a cliffed coastline, and a prograding coastal plain. (C) Advancing glaciers and aggrading rivers cause infilling of valleys. (D) Glacial maximum: unconstrained braided rivers spread glacial outwash over most of the Canterbury Plains. Coastal plain is developed out to shelf edge. The present day geography (not depicted) results from rapid rise in sea-level at the end of the ice-age.

After the oxygen isotope 5e interval, sea level fell and a tombola (land bridge) would have reconnected Banks Peninsula with the plains. The now relict sea cliffs were subjected to subaerial erosion and the tombola would have created a barrier protecting Pegasus Bay from the southerly swells. The falling sea level allowed deposition and preservation of faunas and sediments in estuarine and swamp environments characteristic of the Bromley Formation (and other confining layers). Bromley Formation sediments were deposited about 120,000 to 70,000 years ago between the last interglacial high sea-level stand and glacial low sea-level stand (Brown and Wilson, 1988; Brown and Weeber, 1994). This interpretation is consistent with the tentative correlation, by Brown and Webber (1992), of Bromley Formation with the marine regressive Lower Canterbury Bight Formation of Herzer (1981) and Barnes (1994).

The geography described above is compable to today's. Thus, the present-day prograding Waimakariri-coastal plain is not a suitable analog for the past higher (6-8 m) maximum sea level required to develop coastal cliffs, but is, however, a suitable analog for a lower static and falling sea level.

The high permeability aquifers (e.g. Riccarton Gravel), which partially cover and abut the inferred cliff, represent reworked outwash deposits analogous to those in the present-day prograding Waimakariri River fan. These gravels were primarily deposited during static sea levels comparable to today, or during falling sea level (Fig. 10B). The areas of very high permeability shown in Figure 8 suggest these reworked deposits did not prograde much further east than the limits depicted in Figure 10B. This suggestion is consistent with the limits for the Rakaia Fan imposed by Herzer (1981).

As the climate cycle moved into more glacial conditions, sediment supply increased and the fluvial regime shifted from predominantly degradational to aggradational. The valleys began to fill and were eventually completely overtopped in the later stages of aggradation, allowing unconfined braided rivers to sweep over the whole outwash plains in a pendulum fashion (Fig. 10C and 10D). The coastal plain is built eastward, keeping pace with falling sea level. The ice-age terminated about 14,000 years ago, resulting in a rapid sea-level rise and transgression, starting the cycle over again.

The glacial-interglacial climatic cycle is strongly asymmetric, and modulates the length of time sedimentary processes have to shape the Canterbury Plains and effect the large-scale sedimentary structure. Using oxygen isotope data, Broecker and Denton (1990) show that glacial-interglacial cycles over the past 700,000 years are dominated by a 100,000 year wavelength. A cycle begins with a buildup of ice over a series of steps, for most of the 100,000 years, with a consequent fall in sea level. No single point can be identified when climate shifts from an interglacial into a glacial phase. However at the culmination of an ice age, a critical ice mass is attained, setting up an instability that results in the climate system "flipping". The ice age then terminates with an extremely rapid melting of global ice and consequent rapid rise in sea level (ca. 2 cm a year) for a brief 5,000 to 10,000 years.

The described process of erosion and deposition, or cut-and-fill, most likely occurred during all past glacials and interglacials; possibly up to seven or more cycles are preserved. This repetition suggests the Canterbury Plains is a composite "stack" of numerous, possibly cross-cutting, cut-and-fill deposits. The architecture of the cut-and-fill is controlled by the varying rates of climatic, tectonic, and sedimentary processes operating on the system (Table 3). The higher rates largely define the cut-and-fill boundaries. Subsidence allows for preservation of the deposits.

Table 3 - Approximate rates of tectonic subsidence, eustatic sea level changes, and coastal changes operating on the Canterbury Plains.

Process	rate m/1000 yrs	comment/source
Subsidence (Christchuch sector)	0.03-0.1	Gibb 1986
	ca. 0.25	Brown and Weeber, 1994
Average sea-level fall from interglacial sea-level maxima to glacial minima	1.5	ca. 150m in 100,000 yrs; Chappell 1983 and Broecker and Denton 1990
Eustatic sea-level rise from sea-level minima at the end of glacial to interglacial sea-level maxima	20.0	from 150m to present-day level in 7.5ka; Chappell 1983 and Broecker and Denton 1990, and Gibb 1986
Coastal erosion (south of Banks Peninsula)	1,000-2,000	Kirk, 1969; Wilson, 1985
Coastal progradation (north of Banks Peninsula)	1,500-2,000	Blake, 1964; Wilson, 1976; Basher *et al.*, 1988

Implications for groundwater exploration, modelling, and management

1. The regional-scale geological model outlined here requires consideration and testing. If the model proves to be robust, it defines a large-scale framework for future groundwater modelling, exploration, and management of the Cantebruy Plains aquifers. The subdivisions of lower-order aquifers and aquicludes should be bounded within this larger framework.

2. The transmissivity and hydraulic conductivity patterns show that the general perception that aquifer properties improve in an eastward direction is only partially correct. The patterns suggest the presence of buried valleys with high yields, transmissivity, and permeability across large areas of the plains. The maps should therefore be useful for regional-scale groundwater exploration and management. However, more detailed analysis on a smaller grid and determination of local variations is still required for exploration of smaller areas or for optimal siting of a proposed water bore.

3. The geological model predicts that artesian aquifer conditions, similar to those of the Christchurch artesian aquifer system, occur only east of the arcuate curve between the Waimakariri and Rakaia rivers. This implies that it is unlikely that free flowing artesian conditions in the ca. 0-150 m depth range will be found south of the Rakaia River.

4. The possible preferred flow paths outlined above are an order of magnitude larger (kilometres or valley size) than the "buried river channels" (metres to 10's of metres) suggested by Wilson (1973). The valleys are analogues to "pipelines" or arteries containing the discontinuous buried river channels of Wilson. Future models need to incorporate the larger-scale 3-dimensional heterogeneity.

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